Thermal model for origin of granitic batholiths

THE origin of granitic batholiths has been reviewed in the light of our understanding of the processes operative at destructive plate margins[1]. The thermal patterns at island arcs suggest that significant amounts of thermal energy must be released within downgoing oceanic plates, and that this energy may be transferred to the surface behind island arcs. In these regions heat flow values are abnormally high and andesitic volcanism is present. The high heat flow may be produced because the normal geothermal gradient is enhanced by penetrative convection of magmatic material which is derived from the downgoing plate. Many large granitic batholiths, such as those in Chile, are localised at plate margins, and some authors have proposed that the Sierra Nevada batholith originated in a similar setting along the continental margin in the downwarped portion of the lower crust. Dickenson[2] suggested that the magma for the batholiths was produced by partial melting within the downgoing plate, and was subsequently emplaced into the overlying continental crust. Bateman and Dodge[3], and Brown[1] suggested that emplacement into the lower crust, of andesitic material derived from the mantle, causes partial melting of crustal material, with the mixed product forming the magma for the granitic batholiths.

Experimental studies of the generation of granitic melts[1,5] from metamorphic rocks have shown that melts of granitic composition will occur under conditions of crustal pressure at temperatures of about 700–750° C provided that water is released by dehydration reactions. A granitic melt may be produced by partial melting of the crust at depths of 25–30 km, whereas granodiorites and diorites may be derived from partial melting at higher temperatures (800–950° C) and deeper crustal levels (30–45 km)[1]. At the continental margins where the crust is 40–50 km thick, the radioactive content and the thermal conductivity of crustal rocks, together with the heat flow values, suggest that the lower crust is at a temperature of around 700° C, although higher temperatures may be achieved locally, adjacent to mantle-derived magma. Using thermal considerations I have investigated the extent of partial melting in sialic crust adjacent to a deep seated intrusion.

The thermal effects which a magma induces in country rocks are considered first. The cooling of a magmatic body that is assumed to be instantaneously emplaced may be expressed by the time-dependent equation for heat transport in one dimension:

$$K(\delta^2 T/\delta Z^2) = \rho c_p (\delta T/\delta t) \qquad (1)$$

where K is the thermal conductivity, ρ is the density, and c_p is the specific heat. Solutions to the equation extended to handle partial melting, can be approximated by finite difference techniques: the governing difference equation solved for temperature is

$$T(Z, t+\Delta t) = [1 - 2K\Delta t/\rho c_p (\Delta Z)^2]\, T(Z,t) + K\Delta t/\rho c_p (\Delta Z)^2 \, [T(Z-\Delta Z, t) + T(Z+\Delta Z, t)] \qquad (2)$$

where $T(Z,t+\Delta t)$ is the temperature at time $t+\Delta t$ at a location Z; Δt is the time increment; and ΔZ is the distance increment. The finite difference approximation of the heat conduction equation for two different media, country rock and magma, requires that the respective values of ρ, K and c_p for each medium are used in equation (2). A boundary layer equation is then adapted for the contact between country rock and magma.

I have modified the numerical techniques to handle partial melting. Partial melting is typically an endothermic process, and the total amount of heat of fusion, H, necessary to melt a material is a function of the rock composition. The total melting of basalt may absorb 400 calorie g^{-1}, and 10% partial melting may absorb, proportionately, 40 calorie g^{-1}. The latent heat of fusion of rocks of granitic composition is about 50 calorie g^{-1} (ref. 6). Thus, if it is assumed that a metamorphic rock is composed of 50% granitic material with the other 50% of mafic composition, to melt the granitic fraction of this rock would require a minimum of 25 calorie g^{-1}. It can be assumed that H is absorbed linearly over a range of melting, or as assumed here, the melting process can be simulated by absorption of H at a constant temperature, similar to the melting of a single component chemical species. Although the melting of a metamorphic rock is probably more complicated, that assumption is probably a reasonable description of the fusion process[7].

If a point, Z, reaches the temperature of melting, TM, it absorbs heat according to the equation

$$\Delta H = c_p \Delta T$$

where $\Delta T = T(Z, t+\Delta t) - TM$. (3)

The temperature of the point, Z, is TM until the total heat of fusion, H, has been absorbed. Once H is absorbed, the point responds according to equation (2). The absorbed heat will be released—equation (3)—as the temperature falls to TM and solidification begins; the heat of solidification is assumed to be equal to H. Conservation of energy is maintained throughout the process.

A simple geological model may best illustrate the effect of partial melting in country rocks adjacent to a magmatic body in the lower crust. The following parameters are used: TM = 700° C; ρm (density of magma) = 2.7 g cm^{-3}; ρc (den-

Table 1 Partial melt zones around emplaced magma

Width of magma (km)	Initial temperature of country rock (° C)	TM of country rock (° C)	H (calorie g^{-1})	Total width (both sides) of zone of partial melting (km)
2	500	700	50	0.8
4	500	700	50	1.6
6	300	700	50	1.2
6 (no convection)	400	700	50	0.9
6	400	700	50	1.8
6	500	700	50	3.0
6	500	700	100	2.4
6	500	800	50	1.2
6	500	800	100	1.0
6	600	700	50	5.2
6	600	800	50	2.2
10	500	700	50	4.0

sity of country rock) = 3.3 g cm^{-3}; Km = 0.01 calorie cm^{-1} s^{-1}; Kc = 0.007 calorie cm^{-1} s^{-1}; T_m = 1,200° C (temperature of magma); H = 50 calorie g^{-1}; m = magma; c = country rock. Consider a slab shaped magma, 4 km thick, emplaced into a lower crust with an original uniform temperature of 500° C. It is likely that a mafic magma of this size, and a viscosity of 10^6 poise, will convect naturally[8]. One of the effects of convection within the magma will be to raise the temperatures in the country rocks above the maximum which would have been achieved if the magma was cooling without convection (Table 1). It is assumed that the temperature distribution within a freely convecting magma will be governed by the adiabatic law so that the temperature may vary with depth by anything from 0.3 to 1° C km^{-1}. Therefore the change in temperature from the top to the bottom of a convecting body 4 km thick will be only 2° to 4° C. The temperature distribution within and outside the magma at different times, for one half of a symmetrical model, is shown in Fig. 1. The temperatures reflect the heat

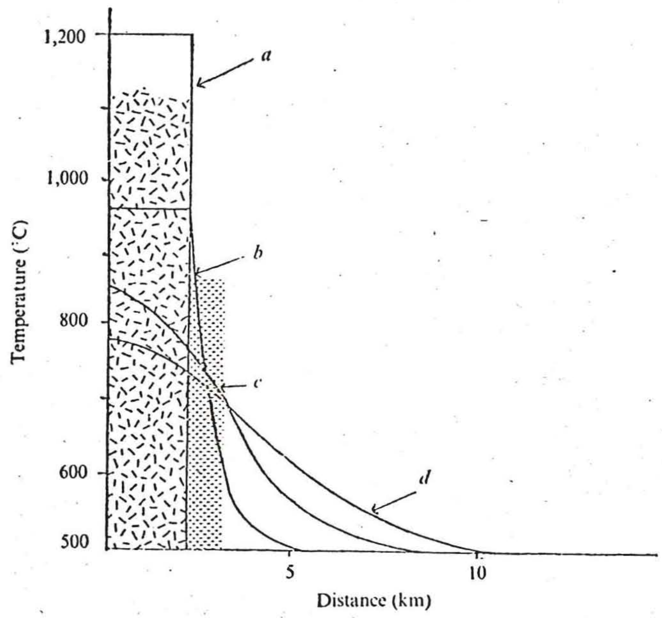

Fig. 1 The temperatures around a symmetrical magma 4 km thick: *a*, at emplacement; *b*, after 2 × 10^4 yr; *c*, after 10^5 yr; *d*, after 2 × 10^5 yr. Stippled area is the zone of partial melting. TM = 700° C; H = 50 calorie g^{-1}; initial temperature of the magma = 1,200° C. At T_m = 900° C convection ceases and the magma cools by conduction.

(H = 50 calorie g^{-1}) which is absorbed in the country rock as a result of melting once the melting temperature, TM (700° C) has been reached. After 40,000 yr, 0.8 km of country rock would be partially melted and the total partial melt zone on both sides of the magma chamber would be 1.6 km wide.

As most magmas derived from the mantle will have properties that do not vary greatly from those assumed here, the size of the partially melted zone around an intrusive will be principally a function of the thickness of the intrusive and of the initial temperature and composition of the country rock. Table 1 shows the width of zones of partial melting for several different examples. In each of these examples the parameters had the values already given here, and except in the cases specified in Table 1, convection within the magma was assumed.

Using H = 50 calorie g^{-1}, a magma 6 km thick intruded into country rock which is at 600° C, would partially melt a zone 2.6 km thick on each side. Within the zone of partial melting the amount of liquid granitic melt is a direct function of the original composition of the metamorphic rock. Thus, for example, with a metamorphic rock which was originally 50% granitic and 50% mafic, the zone of partial melt may be 50% liquid. A slab of magma 6 km thick, emplaced into country rock which is at 400° C, would have a zone of partial melting with a total combined width on both sides, of 1.8 km. If a magma were emplaced into a metamorphic rock of mafic composition with little granitic fraction, the width of the zone of partial melting would be reduced. This can be illustrated by taking H equal to 100 calorie g^{-1} instead of 50 calorie g^{-1}. The width of the zone of partial melting is reduced from 3.0 to 2.4 km thick (Table 1), but the amount of partial melting is still substantial. For a case in which the country rock has a higher temperature of melting (TM = 800° C) and a larger heat of fusion (H = 100 calorie g^{-1}) than the magma, the total width of the zone of partial melting on both sides of the magma would be only 1.0 km. As these properties may be more similar to a mafic metamorphic rock, the partial melting of metamorphic rock around a magma is probably limited. These results (Table 1) indicate that large portions of country rock with some fraction of granitic composition, can be melted under less extreme crustal conditions, and magma convection greatly increases the extent of partial melting in country rocks. Similar quantities of magma will produce, however, little melt in the upper crust (300° C).

If a substantial proportion of the zone of partial melting is liquid, then that liquid may mix with the magma. The amount of physical mixing between magma and country rock is partly dependent upon the relationship between the amount of time it takes the magma to solidify, and the time during which the country rock is melted. At the outer front of the zone of partial melting the country rock may still be melting after the mafic has cooled below its temperature of solidification. As long as the magma is above the solidus, however, the zone of partial melting in the country rock could mix with the magma and thus alter its composition. As an example, it can be assumed that solidification of the magma occurs at 900° C. Then an intrusive 6 km thick within country rock which had an original temperature of 500° C, will solidify after about 92,000 yr (Fig. 2). By that time the country rock would be partially melted out to 1.2 km from the contact, and most of the melt could mix with the magma before it solidified. If only a portion of the country rock melted, then limited amounts of granitic melt above and below the contact would be available for mixing. The refractory country rock material that does not melt may fall to the floor of the magma or be incorporated within it. In this manner the original magma could be diluted with both fused granitic material and refractory country rock.

The extent of partial melting around a slab shaped magma emplaced into a crust with a moderate geothermal gradient, is

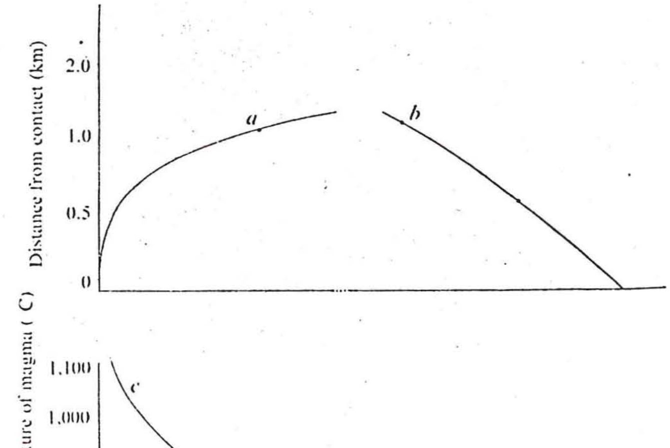

Fig. 2 *a*, The time which elapses between emplacement of magma and the onset of partial melting at increasing distances from the contact; *b*, the time which elapses between emplacement and the onset of solidification in the zone of partial melting at varying distances from the contact; *c*, the rate of cooling of a convecting magma.

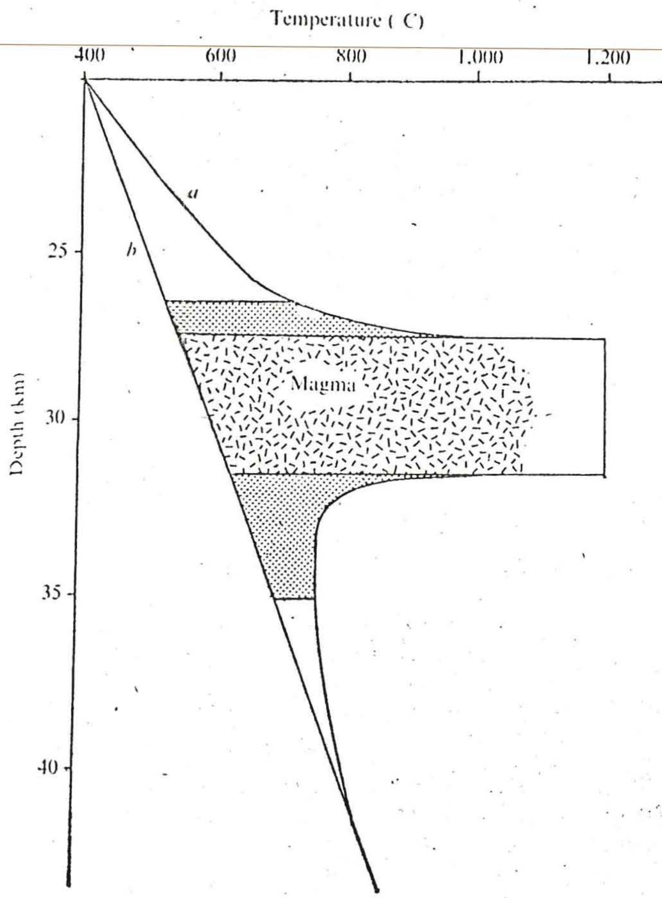

Fig. 3 Temperatures around an intrusive which is 4 km thick, and centred on a depth of 30 km: *a*, maximum temperatures; *b*, geothermal gradient; stippling, zone of partial melting. Initial temperature of magma = 1,200° C; *H*, = 50 calorie g^{-1}.

significantly different from that indicated by the model in Fig. 1. An asymmetrical orientation of the heat fluxing from the top and the bottom of the magma is produced. Taking an initial geothermal gradient of 20° C km^{-1}, and a magma slab 4 km thick emplaced at 30 km depth, the country rock temperatures below the magma would be at least 80° C more than above the magma, and those temperatures would increase with depth (Fig. 3). In this thermal model the values of the parameter given already define the initial boundary conditions, and crustal rocks initially hotter than 700° C (35 km depth) are assumed not to undergo partial melting. The maximum temperatures achieved in the country rock in a case such as that are shown in Fig. 3. About 1 km of country rock above the upper contact, and about 4 km below the magma, would undergo partial melting. The total width of the zone of partial melting would exceed the initial thickness of the magma. In that case, the total heat flux from the top of the magma would be about twice that from the bottom. As much as 5 km of the country rock would have been melted partially, whereas in the model shown in Fig. 1, the partially melted zone is only 1.6 km thick adjacent to a slab of magma of the same size, with initial country rock temperatures of 500° C.

The process of partial melting and solidification of the country rock adjacent to an intrusive, takes approximately 2×10^5 yr in the cases discussed. Solidification of the mafic parent rocks at 900° C takes about 10^4–10^5 yr. Thus, the zones of partial melting adjacent to a magma may have a higher mobility for a considerably longer period of time than either the solidified magma or the lower crustal material that has not undergone partial melting. These zones of partial melting may serve as sites for the emplacement of other material derived from the mantle, and further partial melting in the lower crust may follow.

These thermal models suggest that there are a number of possible ways in which an intermediate rock type could be derived from the mixing of a mantle derived parent with partially melted lower crustal rocks that are adjacent to the intrusion. Dense refractory material in the zone of partial melt, as well as in the parent magma, will tend to sink and lighter granitic fractions will tend to migrate upward. Granitic material may be formed from the partial melt zones above and below the magma and from differentiation of the original magma. These lighter granitic liquids could accumulate into large magmas in the lower crust and then migrate and collect to form plutons of batholithic proportions in the upper crust. In this two stage model of the origin of granitic batholiths along continental margins, the production of the intermediate granitic rocks may be intimately associated with the production and migration of magmas of andesitic to basaltic composition from active subduction zones at the continental margins.

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Thermal contraction joints in a spreading seafloor as origin of fracture zones

IF newly formed oceanic crust is moving away from a spreading axis, it can be expected to contract by cooling. Vertically, the cooling is expressed in the existence and cross-sectional shape of the mid ocean ridges, where there is a relationship between depth of the ocean and age of the crust[1]. A steady state solution for a model of a lithospheric plate which cools while moving[2-4] is compatible with data on topographical height.

Horizontally, in the direction of spreading, cooling will result in the formation of superficial horst and graben structures parallel to the median rift[5,6]. Also in the third direction—that of the ridge axis—internal stresses will arise which are much larger than the breaking strength of rock. If the thermal expansion coefficient λ, is 1.5×10^{-5} ° C^{-1}, and Young's modulus, E, is 1×10^6 kg cm^{-2}, a decrease in temperature of 1,000° C gives an internal tension of 15,000 kg cm^{-2}. Faulting can therefore be expected to occur perpendicular to the direction of the ridge axis, and I propose that fracture zones are the topographical expression of those faults (Fig. 1 and refs 7–10).

The hypothesis provides a model for an orthogonal median rift-fracture zone system[11]. The actual shape of the ridge as a whole will be dictated by other causes, such as the shape of the initial rift. That is exemplified by the parting of North America and South America from Africa. A change in spreading direction can also place constraints on the shape of the ridge. Straight ridges with fracture zones and no offset could be explained using this model: