# The Generation of Granitic Magmas by Intrusion of Basalt into Continental Crust

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## ABSTRACT

When basalt magmas are emplaced into continental crust, melting and generation of silicic magma can be expected. The fluid dynamical and heat transfer processes at the roof of a basaltic sill in which the wall rock melts are investigated theoretically and also experimentally using waxes and aqueous solutions. At the roof, the low density melt forms a stable melt layer with negligible mixing with the underlying hot liquid. A quantitative theory for the roof melting case has been developed. When applied to basalt sills in hot crust, the theory predicts that basalt sills of thicknesses from 10 to 1500 m require only 1 to 270 y to solidify and would form voluminous overlying layers of convecting silicic magma. For example, for a 500 m sill with a crustal melting temperature of 850 °C, the thickness of the silicic magma layer generated ranges from 300 to 1000 m for country rock temperatures from 500 to 850 °C. The temperatures of the crustal melt layers at the time that the basalt solidifies are high (900–950 °C) so that the process can produce magmas representing large degrees of partial fusion of the crust. Melting occurs in the solid roof and the adjacent thermal boundary layer, while at the same time there is crystallization in the convecting interior. Thus the magmas formed can be highly porphyritic. Our calculations also indicate that such magmas can contain significant proportions of restite crystals. Much of the refractory components of the crust are dissolved and then re-precipitated to form genuine igneous phenocrysts. Normally zoned plagioclase feldspar phenocrysts with discrete calcic cores are commonly observed in many granitoids and silicic volcanic rocks. Such patterns would be expected in crustal melting, where *simultaneous* crystallization is an inevitable consequence of the fluid dynamics.

The time-scales for melting and crystallization in basalt-induced crustal melting ($10^2$–$10^3$ y) are very short compared to the lifetimes of large silicic magma systems ($>10^6$ y) or to the time-scale for thermal relaxation of the continental crust ($>10^7$ y). Several of the features of silicic igneous systems can be explained without requiring large, high-level, long-lived magma chambers. Cycles of mafic to increasingly large volumes of silicic magma with time are commonly observed in many systems. These can be interpreted as progressive heating of the crust until the source region is partially molten and basalt can no longer penetrate. Every input of basalt triggers rapid formation of silicic magma in the source region. This magma will freeze again in time-scales of order $10^2$–$10^3$ y unless it ascends to higher levels. Crystallization can occur in the source region during melting, and eruption of porphyritic magmas does not require a shallow magma chamber, although such chambers may develop as magma is intruded into high levels in the crust. For typical compositions of upper crustal rocks, the model predicts that dacitic volcanic rocks and granodiorite/tonalite plutons would be the dominant rock types and that these would ascend from the source region and form magmas ranging from those with high temperature and low crystal content to those with high crystal content and a significant proportion of restite.

## INTRODUCTION

One of the central questions in igneous petrology concerns the generation of silicic magmas. There is now convincing evidence that most of the large plutonic complexes of granite in the continental crust are the result of crustal anatexis (Pitcher, 1987). There is also

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widespread evidence that basaltic magma from the mantle is often intimately associated with the generation of silicic magmas (Hildreth, 1981). This association of mafic and silicic magmas can occur in orogenic belts above subduction zones, in continental hot-spots, and in regions of crustal extension. In plutonic complexes, mafic and intermediate igneous activity are recorded in contemporaneous dyke swarms, small satellite intrusions, and in mafic enclaves within the granites (Vernon, 1983; Pitcher, 1986, 1987). In silicic volcanic centres, evidence of basaltic magmatism is found in satellite lava fields and cinder cones, early lava shields and stratovolcano complexes prior to the main silicic volcanism (Lipman, 1984), and as mafic inclusions and bands within the silicic volcanic rocks (Smith, 1979; Bacon, 1986). Petrological and geochemical features of many silicic igneous rocks are also convincingly explained by admixture of a mantle-derived (mafic) component with a crustal melt. Regions of high temperature and low pressure metamorphism are commonly associated with granite plutonism and a plausible explanation of this association is that basalt is intruded into the crust, causing melting and high heat flow. Indeed basalt underplating of the crust is a currently popular idea to explain both large scale crustal melting and the strongly layered character of the lower crust. While there may be some silicic magmas that are generated by processes without the aid of basaltic input, such as tectonic thickening of radioactive crust (England & Thompson, 1984; Pitcher, 1987), this paper takes the position that in many cases the additional thermal energy of basalt is essential.

The continental crust is strongly layered in terms of its composition, density, and mechanical behaviour. The upper crust is cold and brittle whereas the lower crust is hotter, has a higher density, deforms in a ductile manner, and is commonly characterized by prominent horizontal layering. Basalt magma can be emplaced into the continental crust as dykes and sills and, in some cases where the rate of magma input is high, these intrusions can coalesce to form larger magma chambers.

Dyke emplacement does not seem an efficient way of generating large volumes of silicic magma, because dykes are usually small in width and much of the potential heat for melting will not be utilized if the mafic magma erupts. Sills provide a more promising situation in which *extensive* crustal melting can occur. Horizontal intrusions concentrate their heat at a particular level in the crust and do not dissipate their heat over a large depth range. Sills are intrinsically more efficient than dykes in this respect. Dykes may play an important role in heating up the crust to initiate melting. However, once a region of the crust has become hot, ductile, and partially molten, conditions for dyke propagation become less favourable. A layer or region of partially molten crust provides an effective density barrier and we suggest that basalt magma reaching such a level will spread out as horizontal intrusions. An additional factor which promotes sill formation in the lower parts of the crust is its strongly layered character providing a structural environment in which horizontal intrusions are favoured. For these reasons this paper is concerned principally with the heat transfer and fluid dynamics of sills intruded into hot continental crust.

We consider the cooling and crystallization of basaltic sills emplaced into the continental crust. In particular, we emphasize the situation where the roof of the sill is composed of rock which has a fusion temperature that is lower than the magma temperature and the roof rock consequently melts. This is likely to be the normal situation where basalt intrudes into the typical rock types of mature and ancient middle and upper crust which are already at high temperature. However, the concepts developed in this paper are also likely to be applicable to conditions in immature continental crust such as in island arcs, to more refractory lower crust and to lower crust formed by slightly older or even contemporaneous episodes of basalt underplating. In each of these latter cases, lithologies which have relatively low fusion temperatures can form by differentiation processes and can be remelted by further intrusion of basalt. Thus the model is not confined to the origin of granites, but should be relevant to the origins of intermediate rocks such as tonalites and evolved alkaline rocks such as syenite.

We present experimental studies on the melting of the roof of a sill. We develop a quantitative model of the melting process at the roof, which describes the rates at which a new layer of roof melt forms and the rates at which the underlying liquid layer solidifies. We discuss possible mechanisms by which the melts can be mixed together and also their implications for magma genesis within the continental crust. A companion paper (Huppert & Sparks, 1988*a*) describes the melting of the roof of a chamber from a detailed fluid mechanical point of view.

Throughout this paper the magma will be considered to be Newtonian. Although magma in reality can be non-Newtonian, especially when it is rich in crystals (McBirney & Murase, 1984) its nonlinear rheological properties and the consequences of its non-Newtonian rheology are poorly understood. Two effects may be evident: there may exist a yield strength, so that for a sufficiently low applied stress the magma will not move; and the nonlinear viscosity may alter the heat flux transferred by a convecting magma. Because of the relatively large values of the Rayleigh number that result in most of our calculations, we anticipate that the yield strength will be exceeded by quite a margin. The alterations in the heat flux are at the moment difficult to anticipate and we suggest that the reader views our quantitative results as an indication of the calculated quantity rather than as a precise value. It may be possible to examine non-Newtonian effects with greater insight in the future, but a Newtonian description illuminates many of the fundamental effects and is a necessary first step in order to form the basis for any comparison.

## EXPERIMENTAL STUDIES

The geological problem is to consider how a sill of basaltic magma will melt its roof when emplaced in crustal silicic rocks with low fusion temperatures. We have investigated some of the essential physical features of this problem using polyethylene glycol (PEG) waxes to simulate the wallrocks and aqueous solutions to simulate the magma.

The experiments consisted of placing a wax roof, with a thickness of 15 cm, at the top of a perspex container 20 x 20 x 40 cm high. A hot aqueous solution of $NaNO_3$ was introduced at the base of the container until it was completely filled. The melting temperature of the PEG 1000 wax used is 37–40 °C and the initial temperature of the hot solution was around 70 °C. The density of the molten wax decreases fairly linearly between $1{\cdot}11\ \mathrm{g\,cm^{-3}}$ at 40 °C to $1{\cdot}09\ \mathrm{g\,cm^{-3}}$ at 70 °C. The density of the hot fluid could be varied by altering its concentration. In the present study the density of the aqueous solution exceeded the density of the molten wax. In the experiments the melt formed a separate layer between the roof and the hot fluid. The layer was initially quiescent and transferred heat by conduction, though subsequently convection also occurred (Fig. 1). Qualitatively, the experiment displays simple physical principles which describe the melting of a roof. If the melt is light it forms a discrete layer separated from underlying fluid by a sharp interface. There is negligible mixing of fluids between the two layers.

Experiments and some of the qualitative observations described here have been obtained by Campbell & Turner (1987) using aqueous solutions and their solid equivalents. Their experiments also involved saturated solutions which crystallized as they melted the solid boundaries. The addition of crystallization to some of their experiments did not modify the basic features described here.

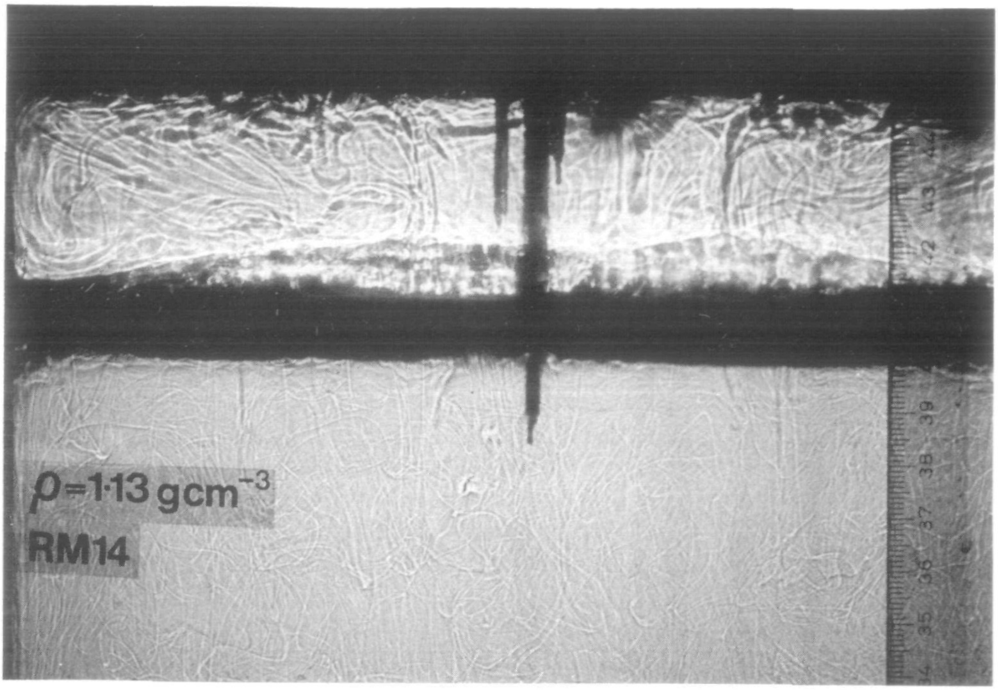

FIG. 1. The photograph shows a layer of molten PEG wax beneath a melting roof. The molten wax is separated by a sharp interface from the underlying hot, NaNO$_3$ solution The interface appears as a dark band on the photograph due to refraction of light, but is in fact knife sharp

Melting at a roof can also occur when the melt generated is heavier than the underlying solution. In this case the melt sinks immediately into the solution and mixes with it. Experimental and theoretical studies of this case are given in Huppert & Sparks (1988*a*), but are not considered further here as the situation of crustal rocks generating melt which is denser than basalt is not likely to be common. Melting can also occur at the floor of a sill. Experimental studies of floor melting are presented by Huppert & Sparks (1988*b*). In this case a theoretical treatment has yet to be developed, because it is complicated by the opposing compositional and thermal effects on density at a floor where light melt is generated. Although there are important geological applications, the discussion of this case must await further research.

Finally, the wax in the experiments has a single melting temperature and thus behaves like a eutectic composition rock. In natural situations the additional complexities of a melting interval will have to be considered. In addition basalt will initially chill against country rock. However, the thermal energy in the convecting basaltic layer will quite quickly melt the chill and we hence omit further discussion of it from this paper. A very brief discussion, which presents the time scale to melt the chill, for a similar situation appears in Huppert (1986) and a more extensive description is currently in preparation (Huppert, 1988).

## THEORY OF A MELTING ROOF

### *General considerations*

A theoretical description is outlined for melting of a roof when a stable layer of low density, silicic magma is generated. A detailed theoretical analysis of the general case of roof

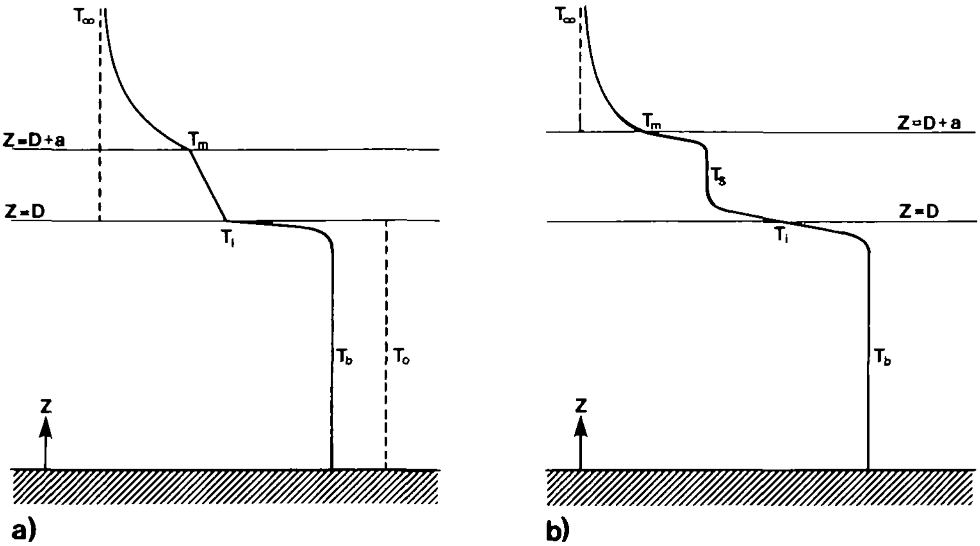

FIG. 2. Boundary conditions of a melting roof region showing temperature variations with height (a) Initial stage in which the melt layer is stable and heat is transferred by conduction from the underlying hot convecting fluid to the melting roof. (b) Later state in which the melt layer convects and has a uniform internal temperature.

melting is presented elsewhere (Huppert & Sparks, 1988*a*). Only the essential results of the analysis, modified for the geological application, will be given here.

We consider a layer of hot basaltic magma with an initial thickness $D$, and temperature $T_b(0)$, emplaced beneath a solid roof and we assume that the solid roof above it has a very large height and has an initial uniform temperature $T_0$. The roof melts at a temperature $T_m$ which is attained at the contact with the underlying basaltic layer after a short time. Thereafter a stable layer of silicic magma forms and grows with time (Fig. 2). Initially the silicic layer is too thin to convect and heat is transferred by conduction across the layer (Fig. 2a). The temperature at the interface between the silicic magma and the basaltic layer is $T_i(t)$. It is assumed that the interface temperature $T_i$ is too high for a chill to form and this will be justified *a posteriori*. Eventually the silicic magma layer increases in thickness to the stage where the critical thermal Rayleigh number is exceeded and convection begins. From then on the silicic magma layer is considered to become thermally well-mixed within its interior (Fig. 2b), and can be characterized by a single temperature $T_s(t)$. The thickness of the melt layer will be denoted by $a(t)$. The basaltic layer, which cools and crystallizes as it loses heat, eventually becomes so crystal-rich that it effectively becomes a solid and convection stops. However, the overlying silicic layer continues to convect and to melt its roof. Finally, the silicic layer itself becomes so crystal-rich and viscous that convection also ceases.

The requirements of a theoretical model are to predict the temperature of the interface between the magma layers, $T_i(t)$, the temperatures of the two magma layers, $T_b(t)$ and $T_s(t)$, and the thickness of the melt layer $a(t)$, as functions of time. It is also pertinent to calculate the time at which convection begins in the silicic magma layer, the time convection ceases in the basaltic layer and the time that convection ceases in the silicic magma layer.

With the assumption of active convection in the basaltic magma, the heat flux to the silicic magma above is given by

$$F_b = \rho_b c_b J_b (T_b - T_i)^{4/3}, \qquad (1)$$

where $\rho_b$ is the density and $c_b$ the specific heat of the basalt and $J_b$ is defined by

$$J_b = 0{\cdot}1\,(\alpha_b g \kappa_b^2/\nu_b)^{1/3}, \tag{2}$$

where $\alpha_b$ is the coefficient of thermal expansion of the basalt, $g$ the acceleration due to gravity, $\kappa_b$ the thermal diffusivity of the basalt, and $\nu_b$ the kinematic viscosity of the basalt. Since the silicic magma layer thickness, $a$, is always significantly less than the length scale of thermal diffusion $(\kappa_s t)^{1/2}$, where $\kappa_s$ is the thermal diffusivity of the silicic magma, we can use the linear conductive profile to describe the temperature in the silicic magma before it begins to convect. The conductive flux through the silicic magma is then given by

$$F_s = k_s(T_i - T_m)/a, \tag{3}$$

where $k_s$ is the thermal conductivity of the silicic magma. The conservation of heat in the basaltic layer of thickness $D$, across the interface between the two magmas and at the solid–melt interface at the roof, then becomes:

$$\rho_b c_b D \frac{\mathrm{d}T_b}{\mathrm{d}t} = -F_b + \rho_b L_b D x_b'(T_b)\frac{\mathrm{d}T_b}{\mathrm{d}t}, \tag{4}$$

where $x_b'(T)$ is the derivative of the crystal content, $x_b$, with respect to temperature, and $L_b$ is the latent heat of crystallization of the basalt;

$$F_b = F_s; \tag{5}$$

and

$$\frac{\mathrm{d}a}{\mathrm{d}t} = k_s H_s^{-1}(T_i - T_m)/a, \tag{6}$$

where

$$H_s = \rho_s[c_s(T_m - T_0) + L_s] \tag{7}$$

and $\rho_s$ is the density of the solid, $c_s$ its specific heat, and $L_s$ the heat of fusion. $H_s$ is the heat needed to raise unit volume of solid crust from its far-field temperature to its melting temperature and also to melt it. The initial conditions

$$T_b = T_b(0), \quad T_i = T_m, \quad \text{and} \quad a = 0 \quad (t = 0) \tag{8a,b,c}$$

complete the specification of the problem.

Substituting (3)–(5) into (6), and using (8a,b) in taking the first integral, we obtain

$$a = H_s^{-1}\{\rho_b c_b D[T_b(0) - T_b] + \rho_b L_b D x_b(T_b)\}. \tag{9}$$

Substituting (1), (3), and (9) into (5), we obtain the following implicit relationship between $T_b$ and $T_i$

$$T_i = T_m + \rho_b^2 c_b^2 D J_b H_s^{-1} k_s^{-1}\,[T_b(0) - T_b + L_b c_b^{-1} x_b(T_b)](T_b - T_i)^{4/3}. \tag{10}$$

Finally (1) and (4) can be written

$$\frac{\mathrm{d}T_b}{\mathrm{d}t} = -(J_b/D)(T_b - T_i)^{4/3}/[1 - L_b c_b^{-1} x_b'(T)]. \tag{11}$$

Equations (8)–(11) describe the heat transfer and melting process while the silicic magma layer is stable to convection and transfers heat by conduction. The solution and derivation of the equations are discussed in more detail in Huppert & Sparks (1988*a*). They cease to be a valid description when the Rayleigh number of the melt layer exceeds a critical value of order $10^3$, where the Rayleigh number is defined as

$$Ra_s = \alpha_s g(T_i - T_m)a^3/\kappa_s \nu_s. \tag{12}$$

Convection in the melt layer is then initiated and both (3) and (6) need to be altered. It is possible, however, that $R_s$, which is small both initially and after a long time, never exceeds the critical value.

Generally, however, convection in the silicic magma layer will set in and this is demonstrated below. Initially the convection will be weak, but as the layer thickness increases it will build up to such a strength that a turbulent layer of uniform temperature forms for which the transfer of heat can be described by the four-thirds relation. We now analyse this situation.

Consider the situation sketched in Fig. 2b. The temperature of the lower, hot basaltic layer is denoted by $T_b(t)$ and that of the silicic magma layer of thickness, $a(t)$, by $T_s(t)$. The heat flux from the basaltic layer is given by equation (1) while that from the silicic melt to the roof is given by

$$F_s = \rho_s c_s J_s (T_s - T_m)^{4/3}. \tag{13}$$

The heat flux into the silicic layer has to be equal to the heat flux out of the basaltic layer and thus $F_b$ can also be defined as

$$F_b = \rho_s c_s J_s (T_i - T_s)^{4/3}.$$

Thus

$$\rho_b c_b J_b (T_b - T_i)^{4/3} = \rho_s c_s J_s (T_i - T_s)^{4/3}, \tag{14}$$

from which we deduce that

$$T_i = (T_b + yT_s)/(1 + y), \tag{15}$$

where

$$y = (\rho_s c_s J_s / \rho_b c_b J_b)^{3/4}. \tag{16}$$

Conservation of heat in the basaltic and silicic melt layers requires that

$$\frac{\mathrm{d}T_b}{\mathrm{d}t} = -(J_b/D)(T_b - T_i)^{4/3}/[1 - L_b c_b^{-1} x_b'(T_b)] \tag{17}$$

$$\frac{\mathrm{d}T_s}{\mathrm{d}t} = (J_s/a)\left\{[(T_i - T_s)^{4/3} - (T_s - T_m)^{4/3}] - \frac{\mathrm{d}a}{\mathrm{d}t}[T_s - T_m - L_s c_s^{-1} x_s(T_s)]\right\}\Big/ \\ \times [1 - L_s c_s^{-1} x_s'(T_s)], \tag{18}$$

where $x_s(T_s)$ is the crystal content of the silicic magma layer and $x_s'(T_s)$ is its derivative with respect to temperature. Conservation of heat at the silicic solid/melt interface becomes

$$L_s \frac{\mathrm{d}a}{\mathrm{d}t} = c_s J_s (T_s - T_m)^{4/3} + c_s \kappa_s \frac{\partial \theta}{\partial z}(a, t), \tag{19}$$

where $\theta(z, t)$ is the (conductive) temperature in the solid country rock with respect to a fixed, vertically upwards $z$-axis. This temperature profile is governed by the diffusion equation

$$\frac{\partial \theta}{\partial t} = \kappa_s \frac{\partial^2 \theta}{\partial z^2} \tag{20}$$

$$\theta(z, 0) = T_0 \tag{21}$$

$$\theta(a, t) = T_m \tag{22}$$

$$\theta \to T_0 \quad (z \to \infty). \tag{23}$$

If the roof melts at a steady rate, the solution of (19)–(23) can be expressed as (Huppert, 1986)

$$\frac{\mathrm{d}a}{\mathrm{d}t} = F_s/H_s. \tag{24}$$

We determined numerical solutions of (15)–(18) in addition to either (19)–(23) or merely (24). The differences were negligible.

The final phase occurs when the basaltic magma layer cools to a temperature at which there are so many crystals present that the viscosity becomes very large and convection in the basalt ceases. However, the silicic magma will continue to convect, and will generate further melt at the roof. The silicic magma will continue to cool and crystallize until it also becomes so crystal-rich that convection ceases altogether. The amount of heat lost by the convecting silicic magma melting its roof is very large compared with the heat gained from the underlying, solidified basalt layer by conduction. Consequently, as a first approximation, the heat flux from the basalt can be neglected. The equations for temperature and thickness variations are then obtained by neglecting (17) and the term $(T_l - T_s)^{4/3}$ in (18). At this stage the conduction of heat into the roof becomes increasingly important as the convection weakens and the conductive terms dominate eventually.

A previous study of the melting of the roof of a magma chamber was presented by Irvine (1970) who modelled the process in terms of thermal conduction. He demonstrated that substantial volumes of silicic magma could be generated by this mechanism. The major difference in the new treatment is the incorporation of convection which dramatically increases the rate of melting and influences the thermal and fluid dynamical history of the silicic melt layer.

## *Modifications and assumptions for geological application*

For application of the roof melting theory to crustal melting by basalt, some additional relationships are required, as well as discussion of under what circumstances the theory can apply in nature.

### *Melting temperature of crustal rocks*

When crustal rocks progressively melt they can be regarded as partially molten solids with mechanical strength when the fraction of melt is low (Shaw, 1980). At this stage the crystalline phases form an interconnected framework. However, beyond some critical melt fraction the connectedness of the crystalline framework is destroyed and the rock is said to have been converted into magma. Experimental and theoretical studies (van der Molen & Paterson, 1979; Shaw, 1980; March, 1981) indicate that there is typically a very large variation in viscosity by up to ten orders of magnitude for small changes of melt fraction and temperature in the vicinity of the critical melt fraction. Various estimates from laboratory and theoretical studies indicate that between 30 and 50% melt is a reasonable range for the critical melt fraction (Marsh, 1981; Wickham, 1987). We assume that the temperature at which the critical melt fraction is attained is a sensible value to take as the effective fusion temperature of the crust roof rock in applying the theory.

Temperatures at which crustal rocks reach the critical melt fraction will vary according to the composition of the rock and the availability of volatiles, principally water. Upper crustal rocks contain an intrinsic water content, in the form of hydrous minerals such as mica and hornblende, which vary from a fraction of a percent to 2% depending on the rock type. The breakdown of these minerals can be the cause of substantial melting over a narrow temperature interval (Wyllie, 1984). On the whole, most crustal rocks have a low intrinsic water content (typically less than 1%) so their melting is close to the anhydrous case. On the other hand water may be added to the region that is melting. Degassing of basaltic magma could add substantial amounts of $CO_2$ and $H_2O$ to the overlying silicic magmas. Circulating hydrous metamorphic fluids could also be added to increase the water content during melting (Wickham, 1987).

Since nature may provide situations varying from melting of completely dry lithologies to melting of water-saturated rock, we have chosen four different cases to cover the spectrum of possibilities. For most of the calculations (the standard case) we assume a granodiorite composition, since this is thought to be the average of the upper crust (Taylor & McLennan, 1985). Experimental studies (Wyllie, 1984) at pressures appropriate to the mid-crust suggest that a fusion temperature of 850 °C is a reasonable estimate for a water content of 2%. For the second 'wet' case we assume a water content of 6% which is an appropriate value for water-saturated granite which would be completely molten at 850 °C and 5 kb pressure. The main influence of water is to decrease viscosity substantially. For the third 'dry' case, we assume no intrinsic water and set the effective fusion temperature at 950 °C. These calculations would be suitable for considering the melting of relatively anhydrous rock or more refractory lithologies. In the dry case the viscosities are somewhat higher due to the absence of water, but this is somewhat offset by the high temperatures of the magmas. Finally, the fourth case is for a granite (*sensu stricto*) which displays eutectic melting behaviour at 850 °C.

*Relationships between temperature, crystal content and viscosity*

The detailed phase relationships of basalt magmas and crustal melts are complicated and depend on composition, total pressure and volatile partial pressures. The crystallinity is generally a nonlinear function of temperature between solidus and liquidus. A parameterization of one magma or rock type under specified conditions would be difficult to apply to another rock or magma or to different conditions. At this stage we consider that the broad features of crustal melting are best examined with the simplest possible parameterization. In actual fact there is very little information available in the petrological literature on how crystal content varies with temperature in silicic systems. We have chosen the following empirical formulae as representative, based on experimental information such as that presented by Wyllie (1984).

$$x_b = 7200\,T^{-1} - 6 \qquad (1091 < T < 1200) \qquad (25)$$

and

$$x_s = 0{\cdot}65(1000 - T)/150 \qquad (850 < T < 1000) \qquad (26a)$$

$$x_s = 0{\cdot}65(1100 - T)/150 \qquad (950 < T < 1100), \qquad (26b)$$

where $x_b$ and $x_s$ are the fractional crystal contents of the basalt and silicic magmas and $T$ is the temperature in degrees centigrade. Equation (26a) represents the standard and wet cases and (26b) represents the completely dry case.

In most of the calculations presented in this paper we assume that the crystals formed in both the basaltic and silicic magmas are small enough to remain suspended in the turbulently convecting magma (Sparks *et al.*, 1984). They thus contribute to an increase in the viscosity of the magmas as the temperature decreases. The empirical relationships that describe the variation of viscosity with temperature and crystal content are based on the work of Shaw (1963, 1972, 1980) and Marsh (1981), and are

$$\nu_b = 10^3(1-1{\cdot}67x_b)^{-2{\cdot}5}\,\mathrm{cm^2\,s^{-1}} \quad (1091 < T < 1200) \tag{27}$$

and

$$\nu_s = \gamma_0 e^{1{\cdot}85\times10^4/(T+273)}(1-1{\cdot}53\,x_s)^{-2{\cdot}5}\,\mathrm{cm^2\,s^{-1}} \quad (850 < T < 1000), \tag{28}$$

where $\nu_b$ and $\nu_s$ are the kinematic viscosities of the basalt and silicic magma, $\gamma_0$ is a pre-exponential constant which has a value of 62 for the completely dry case, 0·62 for the case with 2% $H_2O$, and $0{\cdot}62 \times 10^{-3}$ for the wet case with 6% $H_2O$.

If the country rock is composed of granite (*sensu stricto*) with eutectic behaviour, then crystallization occurs at a single temperature. This case has been examined by modifying equation (28) so that no crystals are present above 850 °C

$$\nu_s = 0{\cdot}62\,e^{1{\cdot}85\times10^4/(T+273)}\,\mathrm{cm^2\,s^{-1}} \quad (T > 850). \tag{29}$$

Table 1 lists values of the various physical parameters used in the calculations that follow.

TABLE 1

*Value of physical parameters used in calculations*

| | |
|---|---|
| $\alpha_s$, $\alpha_b$ | $5 \times 10^{-5}\,\mathrm{K^{-1}}$ |
| $g$ | $981\,\mathrm{cm\,s^{-2}}$ |
| $\kappa_b$, $\kappa_s$ | $8 \times 10^{-3}\,\mathrm{cm^2\,s^{-1}}$ |
| $\nu_b$ | $10^3\,\mathrm{cm^2\,s^{-1}}$ |
| $\rho_s$ | $2.3\,\mathrm{gm\,cm^{-3}}$ |
| $\rho_b$ | $2.7\,\mathrm{gm\,cm^{-3}}$ |
| $c_s$, $c_b$ | $0{\cdot}32\,\mathrm{cal\,gm^{-1}\,K^{-1}}$ |
| $L_b$ | $100\,\mathrm{cal\,gm^{-1}}$ |
| $L_s$ | $70\,\mathrm{cal\,gm^{-1}}$ |

## APPLICATION TO CRUSTAL MELTING

### *The standard case*

We first consider a typical calculation to illustrate the general features which are displayed by all our results. We consider a 500-m sill of basalt at a temperature of 1200 °C emplaced into crust with a fusion temperature of 850 °C and an initial temperature of 500 °C. Initially the interface at the roof will be brought to an intermediate temperature of somewhat over 850 °C. Any chilled basaltic crust, initially formed against the roof, will melt before the overlying roof rocks melt and this dense crust will be mixed back into the convecting layer of basalt. The interface temperature will thus increase to near the basalt temperature. A layer of magma will develop between the interface at 1200 °C and the 850 °C isotherm. Initially this magma layer will be stable and conduct heat, but when the Rayleigh number exceeds about 2000 convection will begin. With this critical Rayleigh number we calculate that the magma layer will begin to convect after 7 days when its thickness will be only 1·2 m. The Rayleigh number exceeds $10^6$, and the model becomes strictly valid, after 1 y when the melt layer thickness is approximately 7 m.

There are at least four reasons for believing that our calculation somewhat underestimates the time-scale and thickness of the layer when convection is initiated. First, there is a strong viscosity gradient across the silicic magma layer reflecting the substantial change in temperature and crystal content across the layer. The thickness of the thermal boundary layer needed to initiate convection may well be somewhat greater if there is a strong viscosity gradient (Richter *et al.*, 1983; Jaupart & Parsons, 1985). Second, it is probable that the crystal-rich parts of the melting region have non-Newtonian properties. In particular a yield strength would inhibit convective instabilities in the thermal boundary layer. Available data on yield strengths in crystal-rich magmas (McBirney & Murase, 1984) do not suggest that yield strength will have any important influence on the criteria developed here. Third, many crustal rocks are likely to have a strongly anisotropic layered structure (as in many migmatites) and the occurrence of continuous layers of refractory lithologies in the zone of melting could inhibit melting for periods. Fourth, the formation of a chilled crust of basalt may temporarily inhibit heat transfer until it breaks away. Although we consider that such a crust is inherently unstable, its behaviour may be complicated and would be impossible to model in detail. However, the important point to emphasize is that the time-scale is very short compared to the overall evolution of the system and that only a thin crustal melt layer is required for convection to begin.

When both the basalt and silicic magma layers are convecting three temperatures characterize the system. Figure 3 shows the variation of the interior temperatures of the convecting magma layers and the interface temperature as functions of time. The silicic magma layer decreases in temperature continuously with time, because melt generated at the roof is being incorporated continuously. The evolution of the silicic magma must thus involve simultaneous melting and crystallization which at first sight appears a contradiction. However, melting is confined to the thermal boundary layer adjacent to the solid roof. Any parcel of fluid in this thermal boundary layer must experience a temporal increase in temperature causing melting to occur. Once a particular region of the thermal boundary layer becomes unstable it will be mixed into the underlying convecting magma where it will attain the mean internal temperature. However, the interior decreases in temperature with

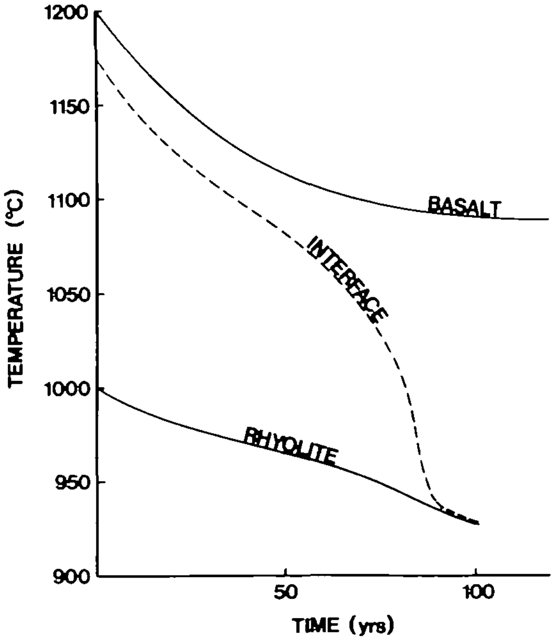

FIG. 3. Variation of the temperatures of the basalt and silicic magma layer with time for a 500 m sill emplaced into crust at 500 °C. The interface temperature between the two magma layers is also indicated.

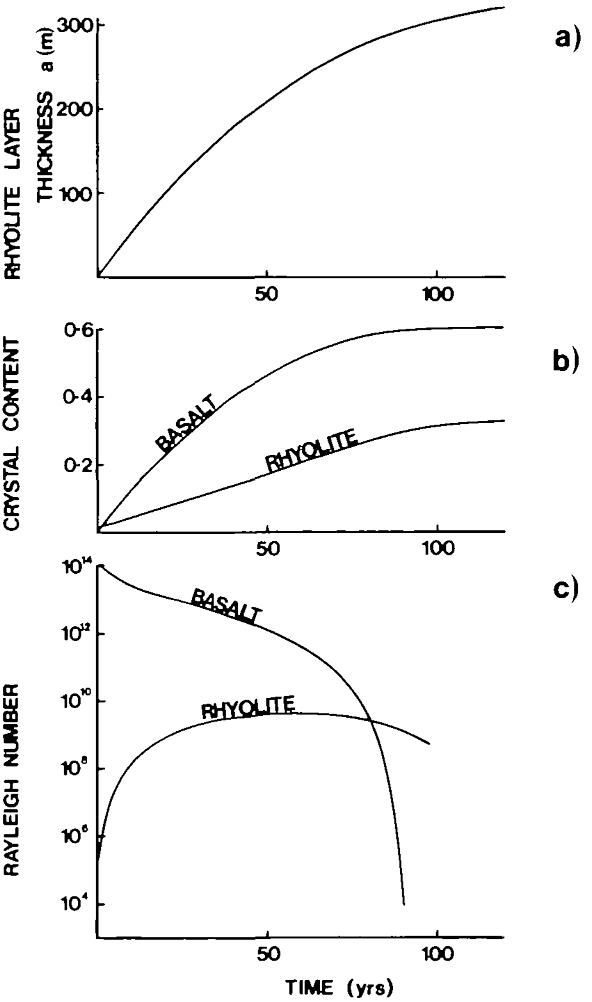

FIG. 4. Variations of (a) the silicic magma layer thickness, (b) the crystal contents of the magmas and (c) the Rayleigh number of the magma layers with time for a 500 m thick sill emplaced into crust at 500 °C.

time and is thus a region of cooling and crystallization (Figure 4). Possible petrological consequences of this convecting structure are discussed below.

The temperature at the interface between the magma layers is initially much closer to the temperature of the basalt than of the silicic magma. This result reflects the difference in viscosity between the magmas. The higher viscosity magma requires a considerably larger temperature difference across the layer than the lower viscosity magma to sustain the same heat flux. Consequently the interface temperature is much closer to the lower viscosity, basaltic magma than to the silicic magma. After approximately 45 y, however, the interface temperature falls below the temperature at which the basaltic magma contains an excess of 60% crystals. After this time the interface temperature rapidly decreases towards the temperature of the silicic magma layer. This occurs because as the basalt layer becomes highly crystalline it becomes more viscous than the silicic magma layer. After approximately 90 y the basalt magma becomes so crystal rich and viscous that convection in it ceases. The cross-over in viscosity of the magma layers is also reflected in the variation of Rayleigh

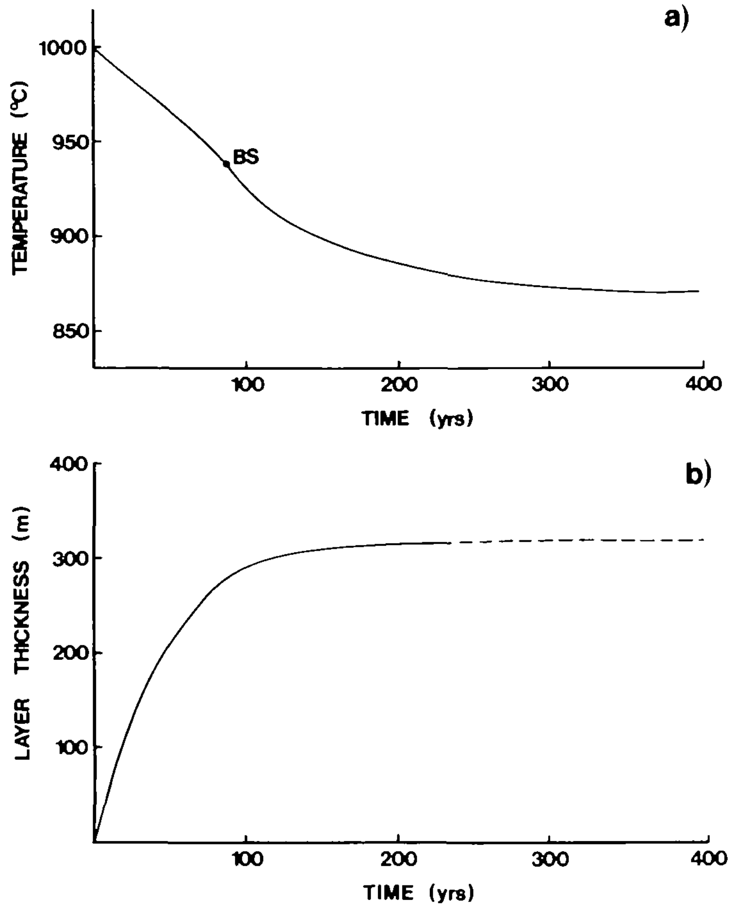

FIG 5 Variations of (a) temperature, and (b) thickness of the silicic magma layer with time for a 500 m thick sill emplaced into crust at 500 °C. The point BS marks when the basalt ceases to convect, because it is too crystal rich

number with time (Fig. 4c). The Rayleigh number of the basalt layer decreases with time as crystallization proceeds and viscosity increases. The Rayleigh number of the silicic magma layer initially increases as the layer increases in thickness, reaching a maximum after approximately 65 y.

Once convection has ceased in the basaltic layer, after 90 y, the silicic magma still has a temperature of 934 °C and a Rayleigh number of $1{\cdot}1 \times 10^9$, with a large temperature difference between the melting roof and the interior. The silicic magma layer will continue to melt its roof, to increase in thickness, to decrease in temperature and to increase in crystal content. Figure 5 shows some of these variations with time. Eventually the silicic magma layer will become so crystal-rich and viscous that the Rayleigh number will fall beneath the critical value and this layer will once again be a partially molten solid rather than magma. The time for the Rayleigh number to fall beneath $10^3$ is about 1000 y as the crystallinity approaches 65%. However, this estimate is quite sensitive to slight changes in the empirical constants used in expressions defining relationships between crystallinity, temperature and viscosity (equations (26)–(28)); and these empirical constants are only known approximately. Furthermore, some extra heat will be conducted away from the underlying basalt layer and this has not been considered. A better indication of the time-scales involved is to choose some lower crystallinity where variations with time are still quite significant. Most silicic magmas erupt with crystal contents below 50%. The time taken to come to 50% crystallinity is 215 y, showing that most of the crystallization and cooling in the silicic magma takes place on a very short time-scale. The thickness of the silicic layer at this stage is 315 m. Little further growth will occur after this time because most of the heat convected out of the silicic layer is subsequently used to heat the roof above the melting front (see below).

Further cooling and crystallization of the non-convecting, partially molten silicic layer will occur by conduction over a much longer time period. We estimate that this layer would take at least $10^4$ y to cool down to the ambient temperature of 500 °C. Thus the emplacement of the basalt sill creates a long-lived region of hot, partially molten crust. If another sill of basalt is emplaced within $10^4$ y then a larger layer of silicic magma will be generated, because the crust is hotter and already partially molten.

Finally we consider the amount of heat conducted into the roof ahead of the melting boundary. The distance, $\delta$, over which there exists a conductive temperature profile can be defined as

$$\delta = \frac{1}{(T_m - T_0)} \int_{a(t)}^{\infty} (T_b - T_0)\, dz. \tag{30}$$

This distance is plotted as a function of time in Fig. 6 for the case of a 500 m sill. Much beyond this distance the temperature is ambient at 500 °C. The proportion of the thermal energy of the convecting silicic layer which is transferred into the overlying region by conduction is indicated. While the layer is still convecting, the conductive layer is thin (a few metres) and the amount of heat lost to this region represents a small proportion of the total heat budget. However, once the basalt solidifies, an increasing proportion of the heat convected out of the silicic layer is conducted away by the country rock rather than melting it. Concurrently the thickness of the conductive profile increases and the roof melting eventually ceases. As indicated in Fig. 5 after about 250 y the silicic layer has ceased to grow.

## *Variations in the main-parameters*

This section explores how variations in the initial conditions influence the evolution of the magma layers.

### *Basalt layer thickness*

The effect of changing the thickness of the basalt layer can be illustrated by calculating the time taken to reach 60% crystallinity in the basalt and by calculating the thickness and temperature of the silicic magma layer at that time. The results are listed in Table 2 which shows that the time-scale of solidification ($\tau_1$) increases with the basalt layer thickness, which reflects the increasing heat content. However, even for a 1·5 km sill the time is still only 272 y, confirming that the melting of the crust has a strong cooling effect on basalt magma.

The calculated temperatures of the silicic magma at $\tau_1$ for different basalt layer thicknesses are virtually identical at approximately 935 °C showing that the silicic magma layer will in all situations be heated up to a high temperature and that the process is closer to total fusion of the crust than to partial melting. The time-scale ($\tau_2$) for reaching 50% crystallinity in the silicic magma layer is also listed in Table 2, and shows that the time-scale for most of the crystallization in the silicic magma layer is also short. The implication is that when silicic magmas are surrounded by rocks of similar composition and melting relations they

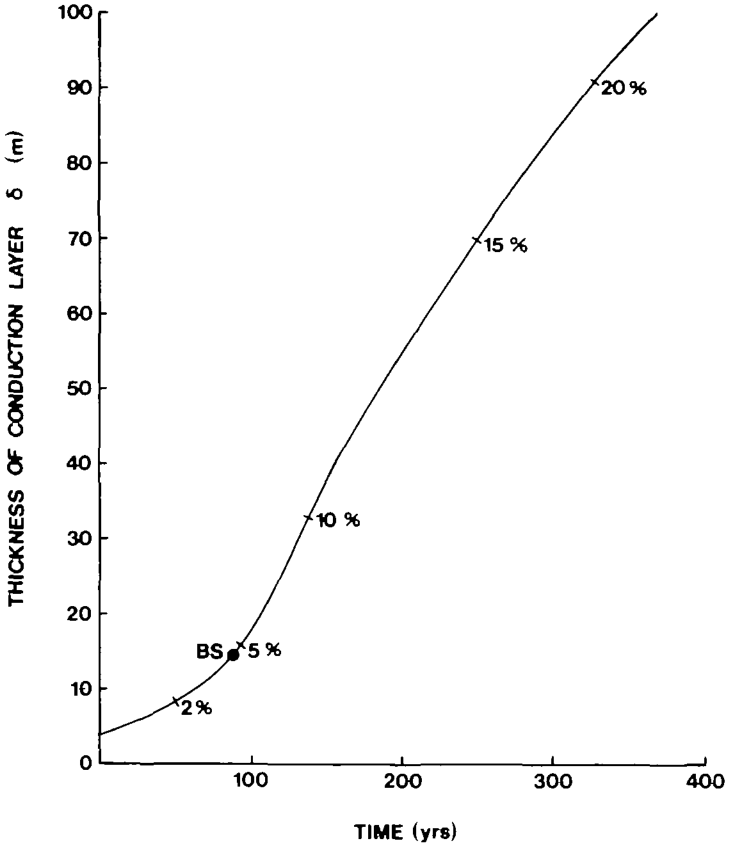

FIG. 6. Variation of the thickness, $\delta$, of the conduction layer in overlying roof rocks as a function of time. The percentage values indicate the amount of heat that is conducted into the roof as a fraction of that in the silicic magma layer.

## TABLE 2

*The table lists the time ($\tau_1$) at which the basalt layer becomes 60% crystalline and ceases convection for different basalt layer thicknesses. The silicic magma layer thickness and temperature at $\tau_1$ are also listed. The time $\tau_2$ is that at which the silicic magma layer becomes 50% crystalline. The country rock temperature is 500 °C.*

| *Basalt layer thickness (m)* | $\tau_1$(*y*) | a($\tau_1$)*m* | T($\tau_1$)°*C* | $\tau_2$(*y*) |
|---|---|---|---|---|
| 10 | 1·6 | 5·2 | 943 | 3·0 |
| 50 | 8·6 | 27 | 937 | 17·5 |
| 150 | 26·5 | 84 | 935 | 58 |
| 500 | 89·7 | 289 | 934 | 215 |
| 1500 | 271·8 | 878 | 933 | 650 |

cannot exist as magmas for long periods of time. They cool themselves off rapidly by melting their boundaries.

If the sills are too small, the model is not appropriate, because the Rayleigh number of the silicic layer does not reach the critical value. The smallest sill for which calculations are presented is 10 m and we estimate that sills smaller than 3 or 4 m in thickness cannot generate a convecting silicic magma.

### *Country rock temperature*

The influence of country rock temperature on the evolution is illustrated in Table 3 for a 500 m sill. As this temperature increases the amount of heat required to melt a given mass of crust decreases, and so the silicic magma layer thickness which evolves increases. It can reach 1·2 km at 50% crystallinity if the crust is already partially molten at a temperature of 850 °C. However, the time-scale and variations of temperature and crystallinity of the silicic magma layer are negligibly affected by changing the country rock temperature.

Table 3 suggests some of the effects which will arise if there are multiple injections of basalt into the crust. If repeated injections are made on a time-scale that is short compared to the overall conductive cooling time-scale ($> 10^4$ y) then each successive injection will form a greater amount of crustal melt. For example if the crust warms up by earlier basaltic intrusions from 500 °C to being in a partially molten state at 850 °C the amount of silicic magma generated increases by a factor of four for a given volume of injected basalt.

We emphasize that the model discussed here is unlikely to be appropriate if the country rock is much colder than 500 °C, because the contact temperature can never reach the effective fusion temperature of the country rock. Thus we would not expect this model to be appropriate in cold crust where a chilled margin in the basalt sill will effectively insulate the country rock from reaching high temperatures. The model is basically applicable to deeper levels of the crust that are already hot and to upper levels of the crust where the rocks have already been pre-heated by earlier intrusions.

### *Lithology and water content*

The rocks of the continental crust vary considerably in their melting and crystallization behaviour and in their viscosity. Fusion temperature and the water content are likely to be the most important variables, because these parameters control the boundary conditions at the roof and the viscosity of the silicic magma layer. Various possible cases are considered and some typical results are listed in Table 4. A wet magma with a water content of 6% and effective melting temperature of 850 °C has a considerably lower viscosity (equation (28)). The wet water-saturated magma consequently convects more readily and melts the roof and cools more rapidly. Thus it takes a shorter time to reach the condition where the basalt

TABLE 3

*The table lists the time ($\tau_1$) at which the basalt layer becomes 60% crystalline and ceases convection for different country rock temperatures in the case of a 500 m basalt sill. The silicic magma layer thickness, $a(\tau_1)$ and its temperature, $T(\tau_1)$ are listed. The time $\tau_2$ is that at which the silicic magma layer becomes 50% crystalline*

| *Country rock temperature °C* | $\tau_1(y)$ | $a(\tau_1)$ | $T(\tau_1)$ | $\tau_2(y)$ |
|---|---|---|---|---|
| 500 | 89·4 | 294·8 | 934 | 230 |
| 750 | 90·2 | 498·0 | 939 | 385 |
| 850 | 90·6 | 688·0 | 943 | 600 |
| 850* | 91·0 | 986·0 | 941 | 795 |

*850 assumes that the crust is already 35% partially molten with a reduced heat of fusion ($L_s = 42$ cal gm$^{-1}$)

TABLE 4

*The table lists the time ($\tau_1$) at which the basalt layer becomes 60% crystalline and ceases convection. The table compares the results for the dry case, the eutectic case, the standard case, and finally the case where the silicic magma layer is always water-saturated at 6% $H_2O$. The silicic magma layer thickness and temperature at $\tau_1$ are also listed. The time $\tau_2$ is that at which the silicic magma layer becomes 50% crystalline. Results for different country rock temperature are presented*

| *Country rock temperature °C* | $\tau_1 (y)$ | $a(\tau_1) m$ | $T(\tau_1)$°C | $\tau_2 (y)$ |
|---|---|---|---|---|
| 500 (dry) | 658 | 197 | 1010 | 805 |
| 500 (standard) | 89·7 | 289 | 934 | 215 |
| 500 (eutectic) | 76·0 | 276 | 910 | 195 |
| 500 (wet) | 25·1 | 337 | 894 | 32 |
| 750 (wet) | 25·4 | 629 | 903 | 45 |
| 850 (wet) | 25·6 | 917 | 908 | 65 |
| * 850 | 25·7 | 1365 | 909 | 85 |

*850 as described in Table 3.

solidifies (Table 4). At this time the temperature of the magma is lower, which can again be attributed to its lower viscosity. The time for the magma to reach 50% crystallinity is likewise much shorter than the dry case. We conclude that if the magmas produced by melting have a lower viscosity than the standard case then the melting process takes place more rapidly, but cooler magmas are generated at any given stage.

On the other hand, for melting an anhydrous lithology with an effective fusion temperature of 950 °C (the completely dry case) the viscosity is considerably greater (equation (28)). The time to generate the melt layer and solidify the basalt is increased and very high temperature magmas are generated.

Although the results for the extreme cases of completely dry and water-saturated depart from the standard case, they represent only relatively small changes in the time-scales involved or in the temperature and crystallinity histories of the magmas. We conclude that irrespective of the lithology of the crust, the water content of the magmas, the thickness of the basalt sill, and temperature of the country rock, the time-scale for solidification of basaltic sills in the crust and melting and crystallization of the adjacent crust is only tens or a few hundred years.

If the crustal rocks were granite (*sensu stricto*) with a minimum melt composition and a single 'eutectic' melting temperature, the time-scales or amount of melt generated would not be significantly affected. This case has been considered using equation (29) and the results are given in Table 4 as the eutectic case. At temperatures of 850 °C or greater the rhyolitic magma would be entirely molten or even superheated. In comparison to the standard case, the time-scale $\tau_1$ for melting the magma, the temperature at time $\tau_1$ and the magma layer thickness are all slightly less. This is expected as more heat is required to melt the country rock totally and the crystal-free magma has lower viscosity at a given temperature. Thus a granodiorite composition wallrock would tend to form porphyritic dacites whereas truly minimum melt granitic wall rocks would tend to produce crystal-poor or crystal-free rhyolitic magmas.

Finally, we consider the case where the wall rocks are entirely refractory and do not melt. If it is assumed that the basalt convects freely and does not solidify against the roof, the time to reach 60% crystallinity is 700 y and would be longer if a roof chill formed. These results show that the cooling rate of basalt sills typically increases by about an order of magnitude when its roof melts. This is discussed in greater detail in Huppert & Sparks (1988a), particularly in the Appendix.

### *Other complicating factors*

The roof melting model does not incorporate all the factors and complexities of nature. We can identify a number of effects which will modify the evolution of the system, but at the moment we only make qualitative statements on their influence.

The crustal rocks could be strongly heterogeneous with interbedded layers of refractory rock with high melting temperatures as is commonly the case in gneisses and regional metamorphic rocks. If the boundary between the melting roof and silicic magma layer reaches a thick refractory layer further melting could be inhibited or even stopped. In migmatites mafic refractory layers typically disrupt into blocks which may resist melting and become enclaves within the silicic magma layer. Some of these dense enclaves could be sufficiently large and dense to fall through the interface into the basalt layer where they may be partially or wholly resorbed or undergo further chemical reactions with a basalt. Such a process would also complicate the thermal and compositional evolution of both magma layers, as well as having interesting geochemical and petrological consequences.

During the evolution of the basalt layer the interface temperature can reach a value where a crystal-rich crust may develop between the layers through which heat must be conducted. Although such a layer would slow heat transfer down the two magma layers could still convect vigorously and the rates would still be controlled by convection.

## *Simultaneous crystallization and melting*

The physical structure and dynamical behaviour of the melting roof region is illustrated in Fig. 7. The effects expected in this region represent a conceptual change in thinking about the relationship between melting and crystallization in magmas. An axiom of many petrological models is that melting occurs in one place (the source) and crystallization occurs at a later time usually in another place (a magma chamber). This is not, however, what happens in crustal melting as described here, where melting and crystallization of the crust occur simultaneously in the source region. The roof region consists of three zones: a region of partial melting, a thermal boundary layer and an internal hot convecting region. The behaviour predicted in each of these regions is now discussed in turn.

In the partially molten roof the temperature gradient decreases from the effective fusion temperature to the ambient country rock temperature over a distance which is determined by the rate of melting of the roof and the conduction of heat. The situation is profoundly different to a magma chamber where crystallization occurs at the margins. In this latter case, the heat loss from the magma chamber and the thermal gradient in the country rock are entirely determined by conduction in the country rock. In a melting roof the heat loss from the magma body is determined by the rate of melting at the roof and therefore by the boundary condition for melting. There is only a very weak dependence due to conductive heat loss through the country rock. The heat leaked into the roof zone is typically only a few per cent of the total heat transferred from the basalt to the silicic magma. The thermal gradient in the partially molten roof is in fact determined by the melting conditions at the roof.

Beneath the roof there is a thermal boundary layer in which further melting must occur. The thermal boundary is a dynamical region where there is a very steep temperature

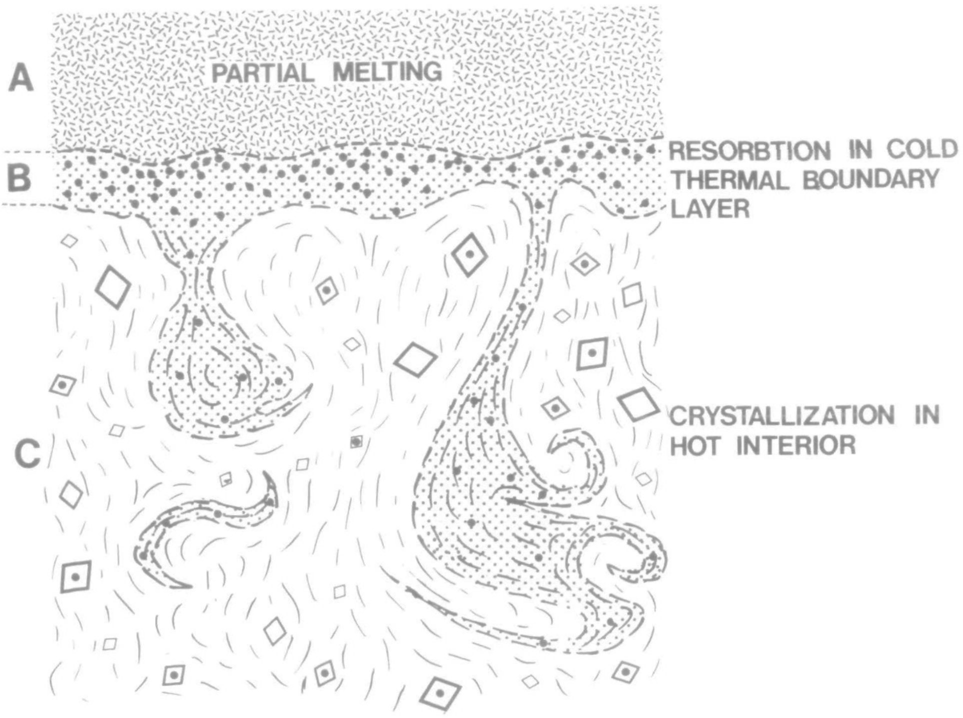

FIG. 7. Schematic diagram illustrating thermal and fluid dynamical structure of melting roof. Partial melting occurs in a roof-zone (A). The thermal boundary (B) is a region of heating and strong thermal gradients in which crystals (restite) are resorbed. The thermal boundary randomly detaches to form plumes which move downwards and mix into the hot interior. The convecting interior (C) is a region of small temperature variations where crystallization occurs due to mixing in of cold magma from above.

gradient from the melting temperature at 850 °C to the interior temperature. Where the thermal boundary layer becomes unstable locally a plume of magma descends into the underlying hot convecting magma. The thermal boundary and the detaching plumes must also be regions of melting since any fluid parcel must be heated from an initial value of 850 °C to the internal magma temperature. Restitic crystals will be resorbed in this region.

The interior of the convecting magma layer is the hottest region, but paradoxically is being cooled. Therefore, crystallization must also be occurring in the interior. An important consideration in this process is the ratio of crystals which form by crystallization in the hot interior to relict crystals derived from the roof. The former are igneous phenocrysts and the latter are restite crystals. At any given temperature between the liquidus and the effective solidification temperature (65% crystallinity) crystals from the roof rocks will not be totally melted, but will be mixed in as a restite component. The amount of restite in the crustal melt layer can be calculated from our numerical results (Table 5).

The percentage of restite decreases with time as crystallization in the magma becomes more important than melting at the roof. For crystal contents between 20 and 50% the proportion of restite ranges from 0·45 to 0·3. These calculations only provide a broad indication of the proportions since at any stage the magma may ascend and crystallize further at shallow depth in which case the restite component will be smaller. As an example consider a magma which at 10% crystallinity contains approximately 5·5% restite and 4·5% phenocrysts. If this magma ascended and crystallized to 50% in a magma chamber

TABLE 5

*The table lists calculations of the total percentage of crystals, the percentage of restite and temperature for various times for the silicic melt layer. The calculations are for a 500 m basalt sill and country rock temperature of 500 °C. However, calculations for different sill thicknesses and different country rock temperatures lead to very similar proportions of restite at a given total crystal content. When the silicic magma approaches 50% crystallinity approximately one third of the crystals are restite, defined as solid crystals from the roof that have been mixed into the melt layer without melting.*

| *Time (y)* | *Total crystals (%)* | *Restite (%)* | *Temperature (°C)* |
|---|---|---|---|
| 30 | 11·1 | 6·0 | 974 |
| 60 | 18·6 | 9·5 | 957 |
| 90 | 29·7 | 12·2 | 931 |
| 150 | 45·4 | 15·3 | 895 |
| 200 | 50·1 | 16·6 | 884 |

there would be 44·5% phenocrysts and still 5·5% restite. Also the melting relationships of some lithologies may be strongly nonlinear in contrast to the simple linear parameters used here, so that the proportions of restite may well vary even more widely than indicated in these calculations. The curves probably give an indication of the maximum amount of restite. The calculations support the view of Chappell (1984) and Wyborn & Chappell (1986) that restite can be a significant component of silicic magmas. However, the proportion of restite in the form of unmelted crystals is significantly less in this model than in that of Wyborn & Chappell (1986), who appear to regard the bulk of the crystals as restite. Because the silicic layer goes through a high temperature evolution a significant fraction of the refractory components of the crust will be melted and then re-precipitated as the layer cools and crystallizes.

A specific application of these ideas and calculations can be made to the interpretation of plagioclase feldspars in silicic magmas. Plagioclase is a particularly useful mineral as it strongly resists re-equilibration and preserves details of magmatic histories in zoning patterns. If the country rock consists of plagioclases with a range of anorthite contents then these will be partially dissolved but will also provide a nucleus for new plagioclase growth in the convecting and cooling interior. We would expect to observe a wide range of plagioclase cores with resorbed outline mantled by normally zoned rims. The cores could be both more calcic or more sodic than the rims if there is a spectrum of rock types in the melting country rock. Temperature fluctuations in the convecting interior could be sufficient to cause oscillatory zoning (Fig. 7). This interpretation is essentially that given by Wyborn & Chappell (1986) and is consistent with observations of dacitic volcanic rocks in the Andes (Francis *et al.*, 1988). Hybridism of crustal magmas with basalt is a closely related phenomenon and can also produce mixed feldspar populations and these zoning patterns. Magma mixing and restite/phenocryst mixing are *both* effects which can be anticipated in the source region of granites.

## GEOLOGICAL IMPLICATIONS

Before developing a general discussion of the implications of the model, we emphasize that the model is for a single intrusive event. In reality, a silicic magma system probably develops by large numbers of individual intrusive events over a prolonged period. However, we believe that the phenomena described for single events captures the essence of the physics for a simple case. More complex models which consider either the progressive intrusion of magma or the cumulative effects of large numbers of intrusive events are clearly worth developing in the future.

### *Current models of silicic magmatism*

There are a number of diverse concepts in the literature on processes involved in the formation of silicic magma systems and the origin of granitoid plutons. Perspectives on current and past ideas are provided by Lipman (1984) and Pitcher (1987). One possible view is that silicic magmatism involves partial melting of crustal source rocks, and ascent of these melts to higher crustal levels where long-lived magma chambers are formed. In the conceptual model envisaged by Smith (1979) and Hildreth (1981), crystallization and differentiation occur in pluton-sized shallow magma chambers and compositional zoning is a common consequence. In these magma chambers it is argued that a zoned silicic cap develops beneath a dominant mafic volume, and the involvement of mafic magma in the development of such systems is regarded as crucial (Hildreth, 1981).

Not all silicic volcanic systems appear to be strongly zoned and some are characterised by eruption of large volumes of rather homogeneous and usually crystal-rich magma. These include the Fish Canyon Tuff (Stormer & Whitney, 1985; Whitney & Stormer, 1985), some volcanic units in the Lachlan Fold Belt of southeast Australia (Wyborn & Chappell, 1986) and the Cerro Galan ignimbrites, in northwest Argentina (Francis *et al.*, 1988) Whitney & Stormer (1985) have proposed the controversial idea that crystallization in the Fish Canyon Tuff occurred in a deep chamber at high pressure before ascent to a high level chamber. Objections to this idea are partly based on questioning the validity of the geobarometric calculations (Grunder & Boden, 1987), but also appear to be based on the opinion that crystallization predominantly occurs in shallow chambers.

Yet another view of silicic magmatism is provided by Chappell (1984) and Wyborn & Chappell (1986) based on studies of S and I type granitoid plutons and associated volcanic rocks in the Lachlan fold belt of southeast Australia. According to them, silicic magmas represent mixtures of restite crystals and partial melt which rise from the source region and thus can be close to the crustal source composition. In some cases mixtures of restite and melt ascend to form homogeneous plutons and crystal-rich volcanic units. In other cases partial melt separates to form crystal-poor magmas which ascend and crystallize along the margins of the chamber to form zoned plutons and zoned volcanic units. Vernon (1983) has criticized the restite hypothesis in its simplest form pointing out that there is involvement of basaltic magma in at least I-type plutons. The mafic enclaves in many of these plutons are more convincingly interpreted as originating by mixing of basaltic magma than of restite.

The crystallization and cooling of shallow silicic magma chambers have been studied by theoretical calculations and experiments. Heat loss is limited by conduction through the margins of the chamber and cooling times are in the range $10^4$ to over $10^5$ y for pluton-sized bodies (Spera, 1979). Crystallization is thus assumed to be a slow process. Brandeis & Jaupart (1986) have argued that in silicic magmas crystallization occurs predominantly at the margins of the chamber because of high magma viscosity. They calculated that the characteristic time-scales for crystal nucleation and growth are short compared to the characteristic time-scales for detachment of the thermal boundary layer. Most of the boundary layer solidifies in the margin. This argument would suggest that internal convection is weak and that little crystallization can occur in the interior. Crystallization at the margin can also generate low density compositional boundary layer flows and this is currently the most popular hypothesis for causing compositional zoning (McBirney *et al.*, 1985).

There are thus a diversity of ideas on how silicic magmas are generated and differentiate. Some of the concepts appear to be in conflict with one another. We suggest that none of these concepts have a monopoly on the truth and that there is merit in them all. However, the results reported introduce new concepts on the time-scales and mechanisms of melting crust and crystallization in silicic magmas. We present an alternative model based on our crustal melting model and which incorporates several of the concepts mentioned in this short review.

### *General evolution*

In normal continental crust with a geothermal gradient of 20 °C km$^{-1}$ rocks in the crust will have temperatures well below 450 °C (Sclater *et al.*, 1980). When an episode of basaltic magmatism is initiated, the crust should often be cold enough to allow some of the basalt to penetrate to the surface. As our calculations indicate, rather little melt will be generated while the crustal rocks are cold and the chilling effect along the margins of dykes and sills will inhibit rapid heat transfer and melting.

However, if the emplacement of basaltic intrusions adds heat to the crust faster than conduction along the geotherm can extract heat, the interior of the crust will heat up and the geotherm will steepen. The heating effect will be accentuated if most intrusions are concentrated at favourable depths governed by density, the mechanical properties of the crust and the stress regime (Ryan, 1987). The focus of intrusion, for example, might be either the Moho or in the middle crust at the transition from dense lower crust to lower density upper crust. As the rocks at the focus of intrusion approach their melting temperature the probability of substantial melting associated with emplacement of a basalt intrusion increases. As argued above, if the country rock temperature exceeds 500 °C the contact temperature for a 1200 °C sill will be of order 850 °C and partial melting will often be sufficient to form magma. As convection breaks up any chilled crust then the interface temperature approaches within a few tens of degrees of the basalt and there is a rapid transfer of heat by convection between basalt and crust.

Once extensive partial melting of the crust has been initiated, we suggest that it becomes difficult or impossible for basalt to penetrate through to the surface. Essentially, basic dykes cannot propagate through a layer of ductile, partially molten crust. When this stage is reached, the heating of the crustal interior must accelerate and the formation of horizontal intrusions beneath the region of partial melting is favoured. While some basalt can reach the surface only a part of the heat is used. However, once the crust becomes too hot and ductile all the basalt will be trapped and so the heat available for melting will approach a maximum. Each individual input of basalt increases the heat content of the crust. Again, if the crust becomes hotter with time, the volume of crustal melting associated with each successive basaltic input must also increase with time (Table 3).

Lipman (1984) has emphasized a common sequence of events in large silicic caldera complexes, which are initiated with basic to intermediate lava shields and then evolve to more silicic explosive eruptions. Such systems often develop to a final stage of a small number of very large volume silicic ignimbrite eruptions and the formation of calderas. Mafic to felsic sequences are also common in plutonic systems (Presnall & Bateman, 1983). Such sequences can be interpreted in terms of progressive heating of the crust. The shield stage represents cold crust which allows basaltic differentiates and/or hybrid andesites (between crust and basalt) to reach the surface (Fig. 8a). As the heat content of the crust

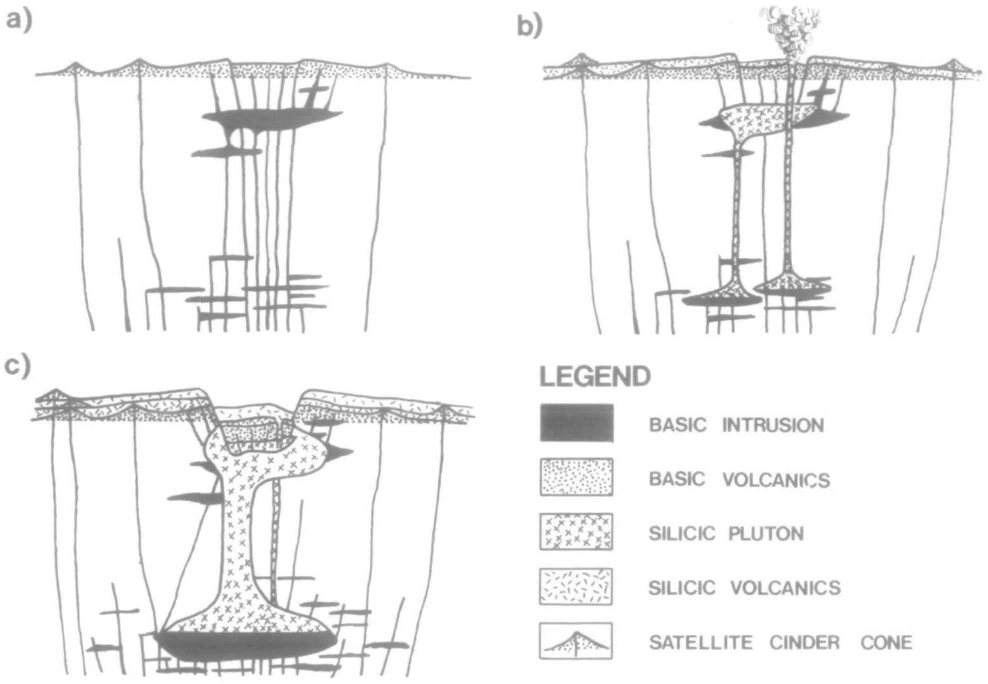

FIG. 8. Evolutionary scheme for a silicic magma system formed by emplacement of basalt into the crust. (a) Early stage where crust is cold and most basalt reaches surface to form volcanic shield or cinder cone field. (b) Crust reaches high temperature so that melting is initiated. Basalt is now mostly trapped within the crust and the silicic magma produced by episodes of intrusion either erupts directly at the surface or forms shallow intrusions. (c) When a large region of the crust is close to melting, large magma bodies can be generated which ascend to the surface causing major ignimbrite eruptions, caldera collapse, and large plutonic units. Basalt can still reach the surface in peripheral regions.

builds up beneath the volcanic system and efficiency of trapping basalt in the hot crust increases, the system evolves predominantly to erupting crustal silicic magmas (Fig. 8b). The volumes of silicic magma should increase with time reaching the climax of large volume ignimbrite eruptions and high level pluton emplacement (Fig. 8c).

There are a number of important differences between this melting model and previous ideas. First there is no necessity to involve shallow long-lived magma chambers at all, although there is no reason to suggest that they cannot form as a consequence of magma generation. Crustal melting and crystallization occur on a very short time-scale ($10^2$–$10^3$ y) compared to the lifetime of a large silicic magmatic system ($10^6$ y). Thus there may be no permanent magma chamber and much of the time the root zone of a silicic system may be partially molten hot crust. Every time a batch of basalt arrives in this root zone a melting event is triggered and this magma could rise straight to the surface or be emplaced at shallow levels to either differentiate, to generate more evolved magma, or to freeze as plutons. (Fig. 8b).

Another important difference is that extensive and rapid crystallization can occur in the source region. A fundamental feature of the model is that melting occurs in the thermal boundary layer and simultaneously crystallization must occur in the interior of the melt layer *as it forms*. The idea that melting occurs in the source, but crystallization must occur after magma ascends to a shallow level in a cooling environment is invalidated. Objections to deep crystallization are removed. The model predicts that large volumes of crystal-rich magma could be generated in the source region in very short time periods. Brandeis & Jaupart (1986) have identified important theoretical difficulties with forming crystals in the interior of solidifying magma chambers. We suggest that phenocryst-rich silicic magmas can form in the source region during melting. High level zoned plutons and volcanic units can form by marginal crystallization of crystal-poor magmas that ascend from the source region. Again this view is close to that advanced by Wyborn & Chappell (1986).

Our calculations suggest that crustal melt layers will initially be at high temperatures. For the standard case the crustal melt is between 900 and 950 °C when the basalt solidifies. For completely dry lithologies even higher temperature silicic magmas could form (Table 4). Thus the model can produce magmas which represent large degrees of partial fusion of the crust. The silicic magmas could thus represent the source composition as argued by Wyborn & Chappell (1986). Previously many petrologists have objected to the restite hypothesis because many crystals in silicic rocks look like phenocrysts grown from a melt and not like metamorphic relicts. We suggest that the results in this paper rationalize this apparent problem. Because the crust is largely melted in the initial stages even refractory components are dissolved. Thus the 'restite' component is melted and then reprecipitated. This effect can be observed in Table 5 where the proportion of restite decreases with crystallization. The views of Wyborn & Chappell are thus given a sound physical basis. Many crystals are of genuinely igneous origin, but, from a geochemist's viewpoint, represent refractory components of the crust.

### *Other implications*

The model envisaged in this paper has some other possible consequences for the evolution of silicic-mafic systems. We do not have the space to explore all these in detail, but offer the following as ideas which are worthy of further work.

(1) The paper has emphasized melting of granodioritic material because this is essentially the average composition of the upper crust. If minimum-melt granitic compositions are melted, then superheated, crystal-poor rhyolitic magmas will have formed at temperatures of 900–950 °C when the basalt solidifies (Table 4). The tendency for rhyolites to be crystal-poor or erupted as obsidians finds a simple explanation. Such minimum melt eutectic compositions are either entirely liquid or entirely solid.

(2) The case for development of compositional zoning in shallow-level magma chambers by fractional crystallization coupled with other effects is overwhelming (Hildreth, 1981). Indeed there is nothing in our models which precludes shallow magma chambers or extensive crystallization within them. Crystal-poor silicic magmas can rise to shallow levels before they have time to crystallize at depth (Fig. 8b). However, the view that all zoned ignimbrites are formed in this way must be questioned. If a heterogeneous crust is being melted it is plausible that zoned systems can be generated in the source region during melting. The components of the crust with the lowest melting temperatures produce melts with lowest density and there is a strong tendency for light fluid to rise to the top, to be the last compositions to solidify and the first to melt on further influx of basalt. While experiments or theory for melting heterogeneous material have not yet been carried out, zoning is a probable outcome.

(3) The model suggests that the cooling history of silicic magmas can depend critically on whether or not melting of the surrounding wallrocks occurs. In the hot source region silicic magmas can cool quite rapidly if surrounded by materials of the same composition and melting relations. For example the silicic magmas effectively resolidify in periods of only a few hundred years by continuing to melt their surroundings. However, in the cold upper crust where heat is extracted slowly by conduction through the boundaries and where

crystallization rather than melting occurs, the cooling rate may be much slower, allowing opportunities for extensive differentiation.

(4) When roof melting is dominant in heterogeneous crustal rocks, refractory dense lithologies may resist melting in the silicic magma layer. An important process of chemical transfer may be the foundering of the refractory blocks through the silicic magma into the underlying basalt. The geochemical consequences of removal of some refractory components from the silicic magma and selective contamination by relatively refractory components of the crust into basaltic magmas should also be assessed.

(5) Another important effect may relate to the role of water in the basalt magma and in the crust. Crustal lithologies have quite low intrinsic water contents and so melting may commonly be under dry conditions generating quite hot magmas. However, some basalt magmas, particularly those associated with subduction zones, contain significant water which is further concentrated during differentiation. Since water has such a large effect on density, it is likely that the silicic differentiates of basalt intrusions can be lower in density than the dry silicic crustal melts. The underlying basalt intrusions could generate water-rich differentiated magmas which could mix with denser overlying crustal melts. Some of these volatiles could also be transferred to overlying crustal melts.

(6) The concepts developed here are not confined to the case of melting old pre-existing crustal rocks. If basalt is repeatedly intruded into the same region during an episode of underplating, later intrusions are very likely to encounter the more silicic or felsic differentiated products of earlier intrusive events. Repeated remelting of such lithologies could generate felsic magmas such as trachytes, phonolites, and alkali rhyolites from basalt magmas of alkaline affinity as well as silicic magmas of tholeiitic or calcalkaline affinity.

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