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The fluid dynamics of crustal melting by injection of basaltic sills

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ABSTRACT: When basaltic magma is emplaced into continental crust, melting and generation of granitic magma can occur. We present experimental and theoretical investigations of the fluid dynamical and heat transfer processes at the roof and floor of a basaltic sill in which the wall rocks melt. At the floor, relatively low density crustal melt rises and mixes into the overlying magma, which would form hybrid andesitic magma. Below the roof the low-density melt forms a stable layer with negligible mixing between it and the underlying hotter, denser magma. Our calculations applied to basaltic sills in hot crust predict that sills from 10–1500 m thick require only 2–200 years to solidify, during which time large volumes of overlying layers of convecting silicic magma are formed. These time scales are very short compared with the lifetimes of large silicic magma systems of around 10^6 years, and also with the time scale of 10^7 years for thermal relaxation of the continental crust. An important feature of the process is that crystallisation and melting occur simultaneously, though in different spots of the source region. The granitic magmas formed are thus a mixture of igneous phenocrysts and lesser amounts of restite crystals. Several features of either plutonic or volcanic silicic systems can be explained without requiring large, high-level, long-lived magma chambers.

KEY WORDS: crystallisation, granites, hybridisation, intrusions, magma chambers, melting of continental crust, restite.

The continental crust is generally well below its melting temperature. In tectonically active regions, however, temperatures can be raised sufficiently to generate substantial amounts of granitic magma. The basaltic magmas that are generated in the mantle at subduction zones, in mantle plumes or during extension of the continental lithosphere probably provide the major mechanisms for raising the temperature of the crust and forming granites. The heat transfer between basaltic intrusions and the continental crust must also result in differentiation of basaltic magmas to generate intermediate and silicic magmas of mantle derivation. Complex physical and chemical interactions between these various magmas may also occur to generate granitic magmas of hybrid character.

This paper highlights some of the heat and mass transfer processes that might occur when basaltic magma is injected into the crust. The problem is considered from a fluid dynamical point of view. We present an analysis of the intrusion of a basaltic sill into a deep, hot region of the crust which is assumed to be close to its melting temperature. This is the simplest case that could be considered and brings out some key features of the melting process. The geological situation where large numbers of intrusions may be emplaced into a geologically-complicated crust over a long period of time is undoubtedly much more complex. However, it is necessary in our view to start by understanding simple physical situations, before attempting to develop more elaborate models. This paper is a companion to a fluid dynamical study of roof melting (Huppert & Sparks 1988a) and a more geologically-oriented discussion of melting at the roof of a sill (Huppert & Sparks 1988b). In Section 1 we present some experimental results on melting horizontal boundaries. In Section 2 we summarise theoretical models of melting at these boundaries. In Section 3 we present some applications of the theory to the geological problem. In Section 4 we discuss the implications for the origin of granites.

1. Experimental studies

Experiments were carried out in a perspex tank measuring $20 \times 20 \times 50$ cm high. A 15 cm-thick roof or floor of polyethylene glycol (PEG) was moulded and placed in the tank and a hot aqueous solution of either NaCl or $NaNO_3$ was then added until it filled the tank and was in contact with the solid PEG roof or floor. PEG is a water-soluble wax with a melting temperature of between 37 and 40°C. The aqueous solutions were emplaced at typically 70°C and the PEG would begin to melt on contact. The density of the aqueous solutions could be varied by changing the content of dissolved NaCl or $NaNO_3$. The kinematic viscosity of the PEG at 45°C is $0{\cdot}78\ cm^2\,s^{-1}$, which is 130 times larger than that of the aqueous solution. In this paper we restrict our discussion to the cases for which the molten wax is less dense than the aqueous solutions. This is the situation of greatest geological interest since silicic magmas are substantially lower in density than basalt. Huppert and Sparks (1988a) describe experiments where the molten wax is heavier than the aqueous solution, but this situation is not thought to be commonplace in geological systems.

Figure 1 shows the result of an experiment where the roof generates a low density melt. A layer of molten wax is observed to form and increase in thickness with time. Sluggish convection currents are observed in the viscous wax layer and more vigorous turbulent convection is observed in the aqueous solution, which has much lower viscosity. There is a sharp interface separating the two convecting layers, with negligible mixing between them. The temperature of the underlying aqueous solution decreases with time as heat

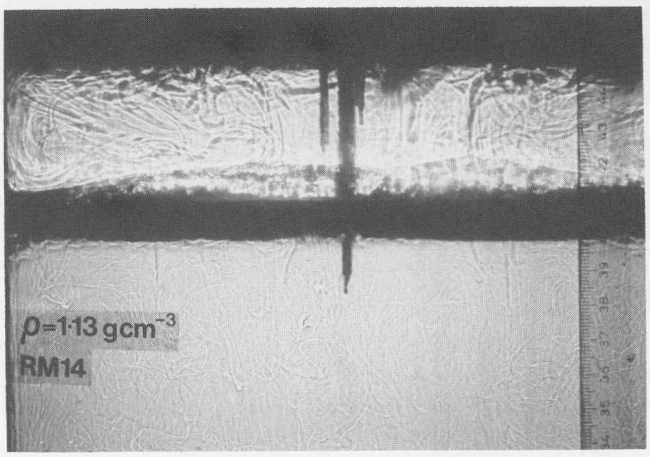

Figure 1 A photograph of a layer of molten PEG wax beneath a solid roof; the molten wax (from approximately 41 to 45 cm) is separated by a sharp interface from the underlying hot aqueous solution; interface artificially thickened due to an incorrect alignment of the illumination.

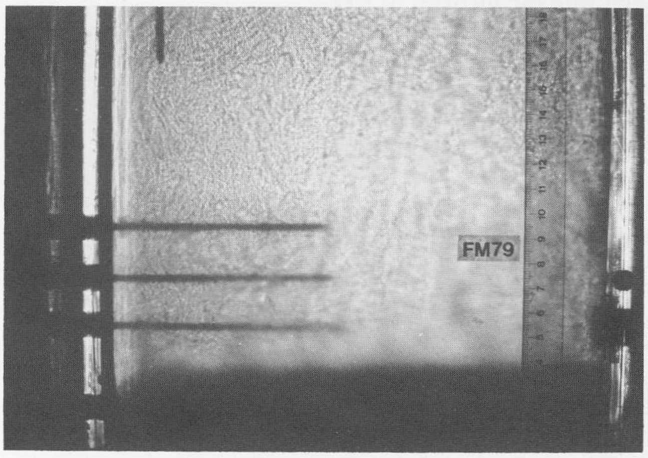

Figure 2 A solid floor melting into an aqueous solution whose density exceeds that of the melt; compositionally-driven, turbulent convection in the aqueous solution tends to mix in the molten wax.

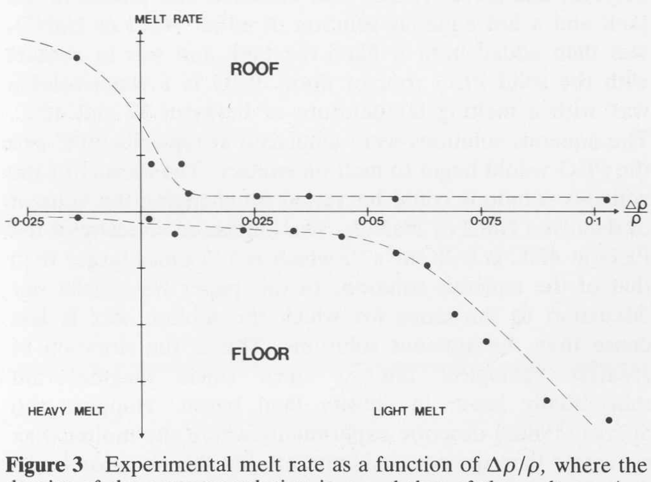

Figure 3 Experimental melt rate as a function of $\Delta\rho/\rho$, where the density of the aqueous solution is ρ and that of the melt $\rho - \Delta\rho$; positive values of $\Delta\rho$ represent light melt; upper portion of diagram refers to melting the roof and lower portion to melting the floor; dashed curves have been inserted by eye to highlight the experimental data represented by ●; for light melts with $\Delta\rho/\rho$ not too small, melt rate of the roof is independent of $\Delta\rho/\rho$, as predicted by the theoretical analysis of Huppert and Sparks (1988a, b); melt rate of the floor, on the other hand, is a fairly strongly-increasing function of $\Delta\rho/\rho$; recall that a theoretical foundation for melting a floor does not yet exist and so these results cannot yet be quantitatively transferred to melting of magma chamber walls.

is transferred to the wax. The experiments demonstrate the processes of melting the roof by convective heat transfer from the underlying fluid and melt layers.

The case of the melting roof has already been presented by Huppert and Sparks (1988a, b). In this paper we describe some new experiments on melting the floor where a light melt is generated. Figure 2 shows the principal effects observed. A thin layer of melt is first observed to form between the solid PEG floor and overlying aqueous solution. Delicate plume structures are observed. After a few minutes thin plumes of wax begin to ascend into the overlying aqueous solution. These experiments indicate that melting the floor causes efficient mixing of the fluids and melt in contrast to the situation of roof melting.

We found that the rate of melting at the floor depends on the density difference between the PEG melt and overlying fluid. Figure 3 plots the average melting rate as the aqueous solution cools from 65 to 55°C. The melting rate is fairly constant up to a density ratio of 6×10^{-2} but increases rapidly for larger values. In our experiments, at larger density contrasts the melting rate at the floor is substantially greater than at the roof for comparable temperature and density differences. How this quantitative result translates to geological systems is not yet clear.

Campbell and Turner (1987) have also carried out experiments on melting roofs using aqueous solutions and their crystalline solids. In some of their experiments the aqueous fluid layer was saturated and crystallised as heat was transferred to the roof. Otherwise their results are comparable to those described here for the melting of a roof.

2. Theory

A quantitative theory for the melting of the roof where light melt is generated has been developed by Huppert and Sparks (1988a). The key features are summarised here. Figure 4 shows a schematic profile through the melting region, using notation described in Table 1. The melt layer formed is too thin to convect during the early stages and so heat is transferred by conduction across the melt layer and in the overlying solid roof. The lower layer convects vigorously. Eventually the melt layer is sufficiently thick that convection is initiated and the internal motion results in a nearly uniform temperature across the layer, except at the upper and lower boundaries. In the experiments the first stage lasted a few minutes only. Huppert and Sparks (1988b) have calculated that in typical geological situations the first stage lasts only a few weeks since a melt layer of granite has to be merely a few metres thick for active

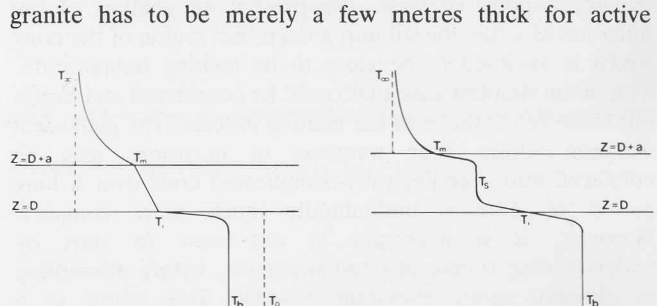

Figure 4 Sketch of the initial and subsequent geometry and mean temperature profiles when a solid roof melts to produce a molten material whose density is less than the underlying liquid.

Table 1 Main physical parameters involved in melting of a roof, with some typical values

Symbol		
a	Thickness of silicic melt layer	
c_p	Specific heat of basalt magma	0·32 cal gm^{-1} K^{-1}
c_s	Specific heat of silicic magma	0·32 cal gm^{-1} K^{-1}
D	Thickness of basalt magma layer	
g	Gravitational acceleration	981 cm s^{-2}
L_b	Heat of fusion (crystallisation) of basalt	100 cal gm^{-1}
L_s	Heat of fusion (crystallisation) of crust	70 cal gm^{-1}
T_b	Interior temperature of basalt layer	
T_M	Effective melting temperature of crust	750–1100°C
T_s	Interior temperature of silicic magma layer	
T_∞	Country rock temperature	
α_s, α_b	Coefficient of thermal expansion	5 × 10^{-5} K^{-1}
ρ_b	Density of basalt magma	2·7 gm cm^{-3}
ρ_s	Density of silicic magma	2·2–2·5 gm cm^{-3}
μ_b	Viscosity of basalt magma	10^2–10^3 poise
μ_s	Viscosity of silicic magma	10^5–10^6 poise

convection to commence. It is only the second convective stage that is of geological importance.

Table 1 lists and defines the major parameters and typical values for the geological case. As would be expected, the relevant parameters are the physical properties of the fluid and melt layers, the dimensions of the layers and the initial temperatures of the solid roof and fluid layer. The problem is solved by balancing the heat fluxes across the fluid layer, the melt layer and the roof. Empirical relationships are also needed to describe the variation of viscosity and crystal content in the layers as functions of temperature. Appropriate relationships are developed and discussed in detail in Huppert and Sparks (1988b). The theoretical analysis results in a set of differential equations which are solved numerically.

Here we draw out some of the key physical effects and constraints which are implicit in the theoretical model. The model assumes that both the lower fluid layer and upper melt layer are convecting at Rayleigh numbers exceeding 10^6, beyond which the well-established "four-thirds" relationship for heat transfer is valid (Turner 1973). This relationship states that the heat flux across a convecting layer is proportional to the temperature difference to the four-thirds power. The geological situation also involves a layer of granite magma of high viscosity overlying basalt of much lower viscosity. Since heat fluxes are inversely proportional to the one-third power of the viscosity, it can be anticipated that a much larger temperature difference will exist across a very viscous fluid than across a low viscosity fluid if the heat fluxes are comparable. The theoretical model also assumes that any crystals in the layers are sufficiently small that they are suspended by the convection and contribute to the viscosity of the fluid.

Our model is substantially different from that of Irvine (1970), who considered that the melting of the roof of a magma chamber took place by conductive processes. He also demonstrated that substantial volumes of silicic magma could be generated, but our incorporation of convective motions substantially increases the rate of melting and changes the predictions for the fluid dynamical and thermal evolution of the silicic melt layer. In addition, we have considered effects due to melting of the floor of the chamber and the resultant formation of well-mixed hybrid magmas.

In transferring our ideas to the geological situation, some further restrictions need to be envoked. If the overlying rocks are too cold, then the basalt can chill against the roof and the rocks may never reach a sufficiently high temperature to melt. We have suggested (Huppert & Sparks 1988b) that the country rock temperature must be greater than 500°C for melting to develop rapidly. Thus the model is applicable to melting of warm regions of the crust and may not be directly applicable to basalt sills emplaced at high levels. Another restriction is that a chilled boundary should not form at the interface between the basalt and granite. We have argued that this is usually the case (Huppert & Sparks 1988b). The main idea is that turbulent heat transfers are generally sufficiently large that they melt any chill which forms on the initial interaction between melt and cold boundary (Huppert in prep.). A final simplification is to consider that the crust has a single effective melting temperature. This is a reasonable assumption, because there are dramatic changes in physical behaviour at degrees of partial melting of about 35–40% (Wickham 1987b) where viscosity decreases by several orders of magnitude in a small temperature interval. When crustal rocks are melted they reach a "critical melt fraction" at a fairly well-defined temperature at which there is a change from partially molten rock to crystal-rich magma. We select this temperature as the effective fusion temperature of the crust.

A theoretical model for floor melting has not yet been developed. The problem is complicated by the fact that the overlying fluid is being cooled and is therefore thermally stable to convection, whereas the release of melt causes it to become unstable. The experiments indicate that, with large density differences, compositional convection can become dominant. However, the nature of the interactions between the thermal and compositional effects have yet to be quantified and so extrapolation to geological conditions is not yet possible.

3. Results

From the theory of a melting roof we have calculated the variation of temperature and crystal content of the silicic and basaltic magma layers with time. We also calculated the interface temperature and the thickness of the silicic magma layer with time. Figure 5 shows graphs of these various parameters with time for the case of 150 m-thick sill emplaced into a crust with initial temperature of 500°C and effective melting temperature of 850°C. These values were chosen to simulate the fusion of hot crust with the average (granodioritic) composition of the upper crust. This effective fusion temperature assumes a water content of 2% and is derived from experimental data presented in Wyllie (1984). From plots in Figure 5 and complementary results presented in Huppert & Sparks (1988b), we can draw a number of important conclusions about the processes of magma generation.

(i) Magma generation initially occurs with almost *total* fusion of the crust, rather than merely partial fusion. In the early stages, the silicic magma layer is at temperatures close to the liquidus of the crust, so that a crystal-poor dacitic or rhyolitic magma would be generated. In heterogeneous crust with refractory layers (as in gneiss terrains) large blocks of refractory lithologies could resist melting and might even sink into the underlying basaltic layer, but even in this case the magmas produced would have compositions corresponding to large degrees of partial fusion.

(ii) The time scales of melting are remarkably short if the basalt is emplaced in discrete intrusive events. In the illustrated case the basalt stops convecting when its crystal content reaches 65% at 1090°C after twenty-seven years. At this stage an 84 m-thick layer of silicic magma has formed. These short time-scales are not particularly sensitive to variations in the main parameters. Variations in the basalt sill thickness from 10 m to 1500 m, in country rock temperature from 500°C to 850°C, in country rock melting relationships, and in water content of the granitic magma from 0% to 6% result in time-scales for melting which range from a few years to a few hundred years (Huppert & Sparks 1988b). The important point is that these time-scales are much shorter than the tens of thousands to millions of years associated with the overall development of granitic magma systems. This is because the process involves heat transfers which are convective (rather than conductive). Substantial volumes of silicic magma could be generated in a hot crust during an episode of basalt intrusion.

(iii) The melting process *always* develops so that the granite magma layer cools as it grows in thickness. The granite layer magma must therefore crystallise at the same time as melting into the layer takes place from the roof. We propose that many of the crystals in granites and their volcanic equivalents are formed in the source region during magma generation. This idea implies that it is not necessary to involve a high-level magma chamber to account for the formation of phenocrysts. These notions also have implications for whether crystals in granites have restitic or igneous origins, which is currently a controversial issue (Chappell *et al.* 1987; Wall *et al.* 1987). According to Chappell *et al.* (1987), many granites represent the mobilisation of a mixture of partial melt and restite crystals from a source region. Our results suggest a modification of these ideas. Since the granitic magma layer has initially a high temperature, much of the refractory material will be dissolved, but is then reprecipitated in the later stages as the granite layer grows and cools. It is possible to estimate the proportion of phenocrysts and restite if the phase relations are well-constrained. In the illustrative calculations of Huppert and Sparks (1988b) the final percentages of total crystals have 70% phenocrysts and 30% restite. In the physical sense the majority of crystals are thus genuinely igneous in origin, but in terms of their influence on the geochemistry they represent refractory components. The substantial crystallisation in the source region makes the model envisaged comparable with the ideas of Wyborn and Chappell (1986) and Whitney and Stormer (1986).

The simultaneous crystallisation is a consequence of the fundamental fluid dynamical structure of a melting boundary and is depicted in Figure 6. The thermal boundary layer is a region in which any parcel of fluid is being continually heated, and so must be a region of melting and crystal dissolution. The thermal boundary layer thickens in a local

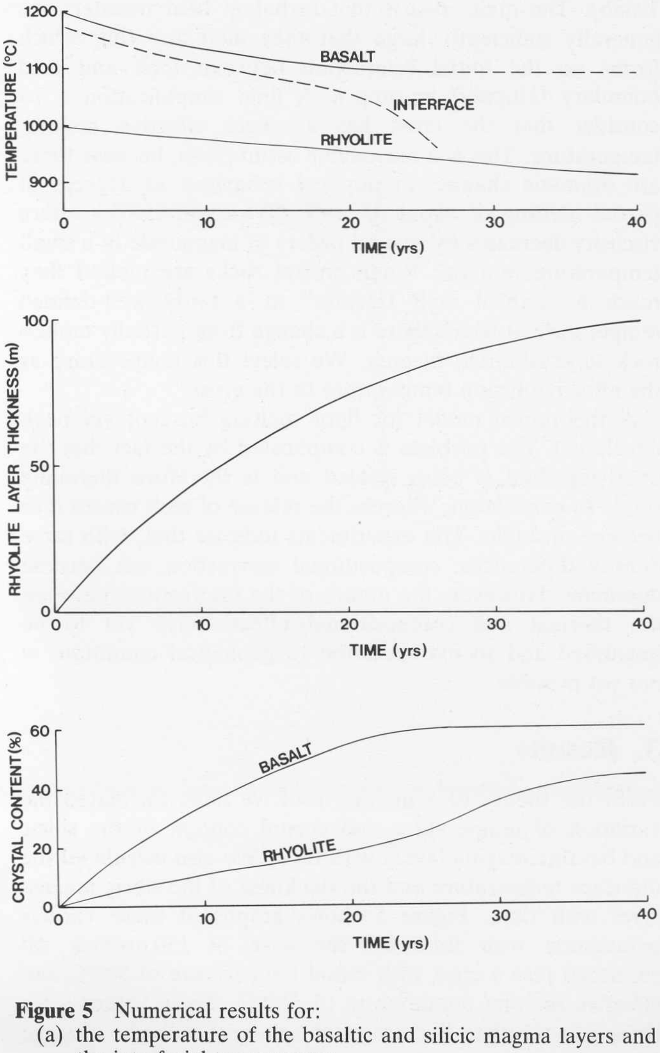

Figure 5 Numerical results for:
(a) the temperature of the basaltic and silicic magma layers and the interfacial temperature;
(b) the silicic magma layer thickness; and
(c) the crystal contents of the magma layers as functions of time.
Results are for a 100 m-thick basaltic layer initially at 1200°C emplaced into continental crust of initial temperature 500°C which melts at 850°C and begins to solidify at 1000°C.

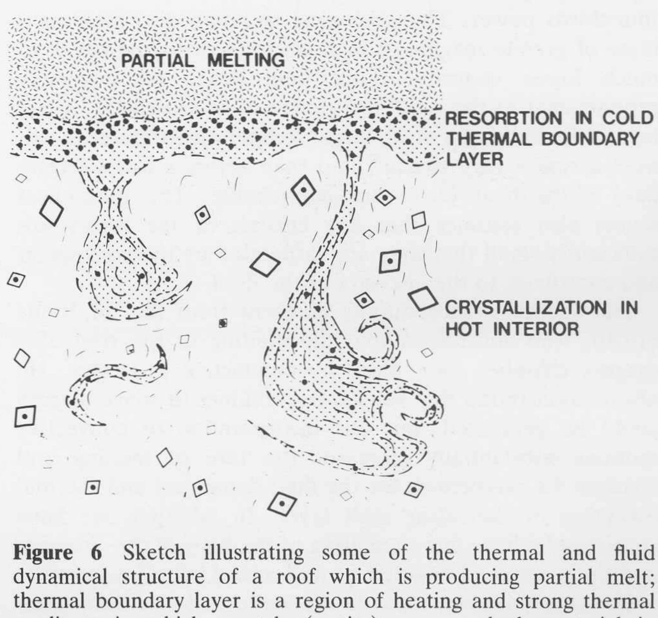

Figure 6 Sketch illustrating some of the thermal and fluid dynamical structure of a roof which is producing partial melt; thermal boundary layer is a region of heating and strong thermal gradients in which crystals (restite) are resorbed; material in thermal boundary layer randomly detaches to form plumes which descend and mix into hot interior; in turbulently convecting interior spatial temperature variations are relatively small and crystallisation occurs due to incorporation of cold magma from above.

region with time until it reaches a critical thickness, where it becomes unstable. A plume of the boundary layer detaches and is stirred into the underlying hotter magma. The interior is thus always the hottest place, but is continually cooled by incorporation of the colder thermal boundary layer. As implied in Figure 6 the resorbing (restite) crystals from the thermal boundary layer can act as nuclei for crystal growth in the convecting interior. Francis *et al.* (1988) suggest that the zoning patterns of plagioclases in silicic ignimbrites of the Andes can be explained by this mechanism. Dacitic lavas and ignimbrites of the Cerro Galan Caldera system contain two distinct populations of plagioclase core compositions and thick oscillatory zoned rims. The cores could be restite and the zoned rims crystallisation from magma, although there are a number of alternative interpretations (see Wall *et al.* 1987).

(iv) After solidification of the basalt a granitic magma layer will continue to melt its roof, grow in thickness and crystallise if it is surrounded by wall-rocks of similar composition. We calculate that the time-scale for re-solidifying the granitic layer is typically a few hundred to a few thousand years. Paradoxically, a layer of granitic magma remaining in its hot source region may solidify much faster than the source magma emplaced in the upper crust against cold or refractory rocks.

(v) The process involves the rapid differentiation of basaltic magmas and allows a number of opportunities for hybridism with the continental crustal magmas. A basaltic magma that crystallises by 60–65% during its existence as a layer of convecting magma would contain a substantial amount of residual interstitial liquid of intermediate to silicic composition which would be available for segregation from the basic crystal mush and mixing with overlying silicic crustal melts or for remobilisation in subsequent intrusive events. Water-rich basaltic magma, for example, could well generate a silicic residual melt which has a lower density than dry silicic melts of the crust. Melting of the floor of a sill provides another mechanism of hybridism between crustal and mantle magmas, as is illustrated in the experiments concerned with melting the floor.

4. Broader implications

The evolution of a large silicic magma system is complex and involves the interplay of many factors. The rate of mafic magma input from the mantle, the tectonic stress regime and the structure and detailed geology of the crust come to mind as important, but by no means the only, factors in controlling the generation of silicic magmas. Our calculations are based on a single intrusive event, whereas actually these systems probably develop by a large number of intrusive and tectonic events over hundreds of thousands to millions of years. However, we suggest that important features shown by the simple models would remain, however complex the actual natural system becomes.

Figure 7 shows schematic diagrams illustrating the possible development of a large granite plutonic complex

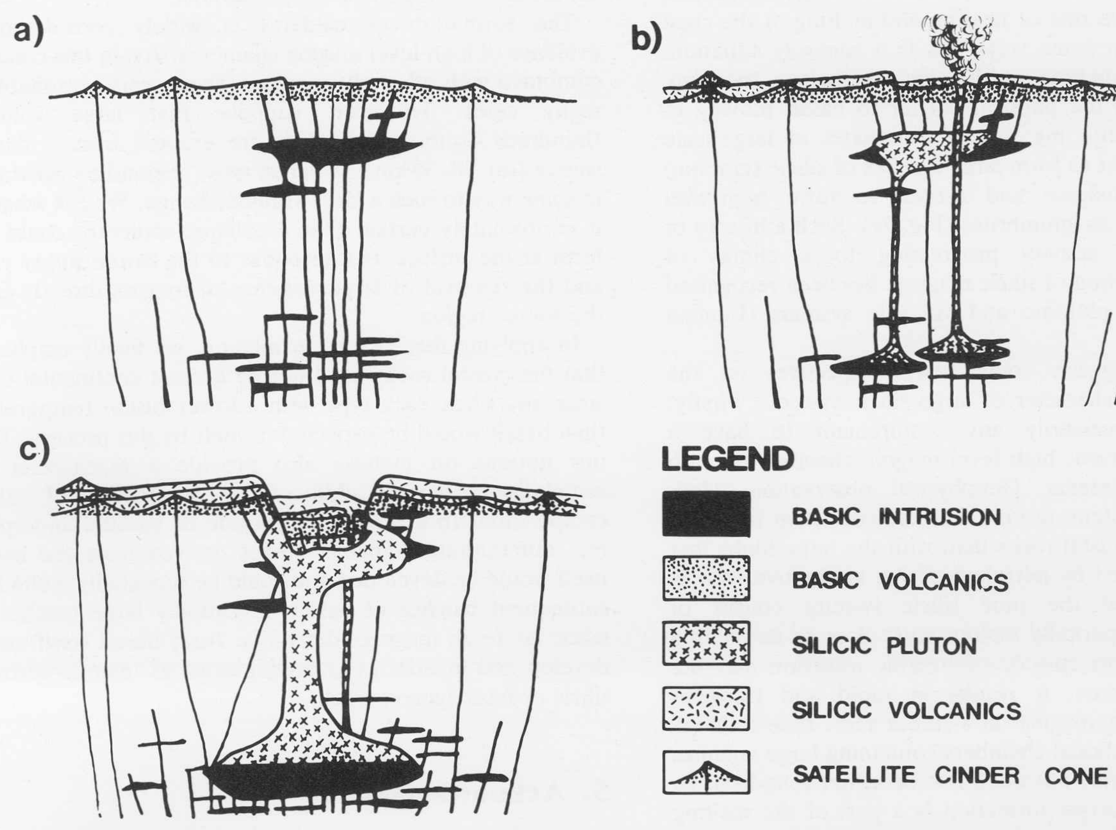

Figure 7 Evolutionary scheme for a silicic magma system formed by emplacement of basalt into the crust.
(a) At an early stage when crust is cold; most of the basalt reaches surface to form either a volcanic shield or a cinder cone field;
(b) later on, the crust reaches a sufficiently high temperature that melting is initiated; basalt is now mostly trapped within crust and silicic magma produced by episodes of intrusion either erupts directly at surface or forms shallow intrusions;
(c) when large region of crust is close to melting, large magma bodies can be generated which ascend to surface, causing major ignimbrite eruptions, caldera collapse and large plutonic units; basalt may still reach surface in peripheral regions.

and associated caldera complex. In Figure 7a the continental crust is invaded by dykes and sills of basaltic magma, generated by some major tectonic process such as subduction, lithospheric extension or a mantle plume. In the early stages the continental crust has a normal geothermal gradient so the temperature everywhere is well below the fusion temperature of the rocks. At this stage, the crust is sufficiently cold to allow dykes to reach the surface and some eruptions of basalt and derivatives intermediate to silicic magmas can occur. Opportunities for minor melting and crustal assimilation also might occur in magma chambers that develop, or during ascent in dykes (Huppert & Sparks 1985). It can be presumed that some (unknown) proportion of the basaltic magma is intruded and begins to heat up the crust. This will be most effective if there is a focus of the intrusion, for example at places where there is a prominent change in physical properties of the crust (Ryan 1987). The Moho and the transition from denser, more ductile, lower crustal rocks to lower density, cold, brittle upper crust are levels particularly susceptible to acting as physical magma traps.

In Figure 7b the level of intense intrusion approaches the fusion temperature and large intrusive events should begin to generate batches of silicic magma by the mechanism described here. Additionally, mixing of crustal and mantle magmas to form hybrid intermediate to silicic magmas can occur if physical conditions (notably density relations) are appropriate. Once a region of the crust becomes hot and partially molten it creates a significant physical barrier to the penetration of basaltic magma to the surface. We suggest that at this stage most of the basaltic magma becomes trapped and so the rate of heating and melting of the crust will accelerate. In some ways this is a runaway situation, because as more melting and heating takes place, the more effective becomes the physical barrier to basalt moving to the surface. Finally, the system culminates in large-scale melting of the crust to form large batches of silicic (granitic) magma which coalesce and ascend to form high-level plutons and erupt as ignimbrites (Fig. 7c). Such a history of prolonged mafic activity progressing to a climax of large-scale production of silicic magmas has been recognised in a number of plutonic and volcanic systems (Lipman 1984).

Our model suggests some new perspectives on the development and character of large silicic systems. Firstly, there is not necessarily any requirement to have a permanent, long-lived, high level magma chamber beneath silicic volcanic systems. Geophysical observations (Iyer 1984) on active systems are in fact more consistent with hot, partially molten crustal rocks than with the large blobs that have been imagined by petrologists. An alternative view is that for much of the time silicic systems consist of anomalously hot, partially molten crust close to its melting point. Whenever an episode of basaltic intrusion into the source region occurs, it results in rapid and transient generation of silicic magma on a rather short time-scale (*c.* 10^2 years). Conventional chambers containing large volumes of magma may form, but need not be either long-lived or usual. Since phenocryst formation is a part of the melting process, these magmas could ascend directly to the surface without requiring an intermediate high-level chamber.

This alternative view is not inconsistent with the conventional high-level, long-lived chamber model. It is indeed likely that batches of magma generated at depth can intrude at high-level and develop into magma chambers which have an important influence on the character of the erupted rocks. However, it is possible that too much emphasis has been placed on the high-level processes in explaining what is observed. Many granites appear to be intruded at high-level as crystal mushes (Pitcher 1986) implying that much crystallisation occurs at deeper levels. Whitney and Stormer (1986) have also interpreted the petrology of the Fish Canyon Tuff in terms of deeper-level crystallisation. Similarly, the ideas of Wyborn and Chappell (1986) place more emphasis on processes in the source region. The fluid dynamical modelling provides another set of arguments for believing that processes which occur at depth have an important influence.

Two common features of silicic systems are often cited as strong evidence for high-level magma chambers: compositional zoning and caldera formation. We do not accept that zoning necessarily provides conclusive evidence of high-level chambers. We suspect that melting of heterogeneous source rocks could also generate zoned magma bodies in the source region during melting. Examples of zoned magmatic systems due to melting are in fact known from migmatite complexes. For example Wickham (1987a) has shown that leucogranite intrusions emplaced within and above regional migmatites zone downwards into biotite granite and then to diorite in the Trois Seigneurs Massif in the Pyrenees. Geochemically and petrographically the leucogranites are equivalent to crystal-poor high silica rhyolites and the biotite granites and diorites are comparable to crystal-rich dacitic magmas. We do not dispute that the high-level processes for generating zoned magmas are important (Hildreth 1981), but suggest that the very stimulating hypotheses that have been developed from the study of zoned ignimbrites should not come to be regarded as rigid dogma.

The formation of calderas is widely considered as evidence of high-level magma chambers. Again this concept, combined with other observations, seems very reasonable in many cases. However, suppose that large volumes (hundreds or thousands km^3) are erupted from a deeper source (say 20–30 km): the crust must respond by deforming in some way to such a large volume change. We ask whether it is absolutely certain that a collapse structure could not form at the surface as a response to the brittle upper crust and the removal of large volumes of magma directly from the source region.

In applying these ideas on melting, we finally emphasise that the crustal rocks need not be ancient continental crust since *any* silicic rock type with a lower fusion temperature than basalt would be expected to melt by this process. Thus our notions on melting also provide a mechanism for remobilising recent additions to the crust of silicic composition. In a prolonged episode of basaltic underplating, differentiated igneous bodies derived from the basalt itself would be developed and could be repeatedly melted by subsequent batches of basalt. Eventually large batches of silicic or felsic magmas derivative from basalt itself could develop and invade the overlying crust as 'mantle-derived' silicic or felsic igneous rock.

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