Thermal control of basaltic fissure eruptions

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AS hot, basaltic magma rises in a newly opened dyke during an eruption, it flows through colder crustal rock. The flowing magma advects heat along the dyke; at the same time, heat is conducted out of the dyke into the colder surroundings. The loss of heat will lead initially to the channel becoming constricted by magma solidifying against the walls. The channel may then become completely blocked, which will end the eruption at that site before the supply of magma is exhausted. Alternatively, the continual supply of heat by the flowing magma may, after some time, exceed the losses into the country rock. In this case, the initial solidification is reversed, the walls of the channel are progressively melted and the dyke is widened until the supply diminishes. We present here a model that quantitatively delineates these two regimes. We also identify an intermediate regime in which parts of the surface fissure may become blocked and the eruption continues from isolated vents.

Basaltic eruptions display a fascinating variety of behaviours[1]. The low viscosity of such lavas allows relatively steady degassing and flow, as opposed to the explosive eruptions characteristic of more viscous magmas. 'Quiet' basaltic eruptions nonetheless evolve over time in a highly complex way. Typically, a linear system of fissures opens rapidly and erupts a continuous 'curtain of fire'. Small eruptions simply cease within half a day. In larger eruptions activity begins, after several hours, to concentrate at certain points along the fissure[2,3]. If eruption persists, it becomes localized at one or a few vents, building cones. Preferential localization of subsequent fissure eruptions to the same areas produces substantial shield volcanoes. In the extremely voluminous eruption at Laki in 1783, flow persisted at not one but dozens of vents along the fissure[4].

The initial formation of the dyke is the result of elastic processes in the surrounding country rock[5,6]. We concentrate here on the primarily thermodynamic problem of determining both the flow in the dyke and its evolution after the initial rock propagation and widening have taken place. Naive estimates of the solidification of magma against the walls of the dyke based solely on conduction[7], indicate that flow should effectively cease in a matter of days. A previous model[8] which included some advective effects came to much the same conclusion unless brecciation of the dyke wall and local widening of the conduit by removal of the fractured blocks leads to increased flow rates[9]. There are deficiencies in this model, however, most notably the neglect of both the latent heat of solidification and the effect of the thermal advection in the magma on the temperature profile of the solid. These omissions consistently overestimate the tendency of the dyke to become blocked. Our model includes both these effects and allows us to understand how meltback of the walls can actually occur.

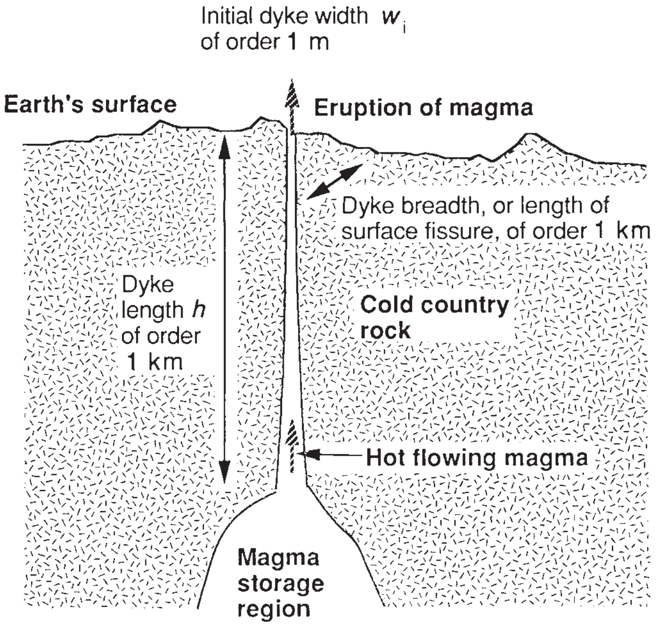

FIG. 1 Cross-section through a typical sheet-like planar dyke. Note that the dyke need not necessarily be vertical.

To obtain our results we analysed an initial-value problem for $\theta(x, z, t)$, the temperature in both the fluid and the solid with respect to locally cartesian coordinates z measured along the wall and x perpendicular to it, with $x=0$ at the wall. In constructing the equations, we assumed that the flow in the thermal boundary can be represented by a shear flow of magnitude γ. The governing equations then become

$$\theta_t - v\theta_x + \gamma x \theta_z = k\theta_{xx} \quad (x, t>0) \qquad (1)$$

$$\theta = T_m \ (z=0), \qquad \theta = T_w \ (x=0), \qquad \theta = T_m \ (t=0 \text{ or } x\to\infty) \qquad (2)$$

$$\theta_t - v\theta_x = k\theta_{xx} \quad (x<0, t>0) \qquad (3)$$

$$\theta = T_w \ (x=0), \qquad \theta = T_\infty \ (t=0 \text{ or } x \to -\infty) \qquad (4)$$

The velocity of migration of the interface between fluid and solid, v, is proportional to the difference in the conductive heat flux across the wall and is determined from

$$(L/c)v = -\kappa[\theta_x(0+, z, t) - \theta_x(0-, z, t)] \qquad (5)$$

The shear flow at the edges of the channel of width W is driven by a pressure difference ΔP and results in a volume flow rate $Q(t)$. These are related by

$$\Delta P = 12\mu Q(t)\int_0^h W^{-3}(z, t)\,dz \quad \text{and} \quad \gamma(z, t) = 6W^{-2}(z, t)Q(t) \qquad (6)$$

With either solidification or melting, the half-width gradually changes according to

$$W(z, t) = W_i - 2\int_0^t v(z, t')\,dt' \qquad (7)$$

The solutions to equations (1)-(7) were obtained numerically by integrating suitably simplified and non-dimensionalized versions which are formally valid for all time[10,11].

We consider, as depicted in Fig. 1, the dyke to be a two-dimensional channel the width of which is initially uniform but subsequently changes because of local solidification or melting. The newtonian flow is considered to be laminar and the driving pressure is taken to be fixed. Thus if the channel becomes constricted, the total flow rate decreases. Cooling of the magma in the dyke is typically confined to a thin boundary layer adjacent to each wall. The width of the boundary layer increases with length along the dyke and can attain a value of ~10 cm at the end of the dyke. The difference between the heat supplied to the wall from this boundary layer and that conducted into the surrounding solid determines the rate of solidification or melting, which is directly proportional to the magnitude of the latent heat. The temperature profile in the solid is determined numerically taking into account the effects of advection in the magma. A full derivation of the governing equations and details of their solution are given in refs 10 and 11.

The behaviour of the solutions is governed by three dimensionless parameters. The first is the Stefan number of the magma, $S_m = L/(c(T_m - T_w))$, where L and c are the latent and specific heats per unit mass, T_m is the temperature of the magma at the entrance to the dyke and T_w is the temperature at the wall, which is defined as the temperature at which the crystallizing magma ceases to flow. In fact, solidification and melting take place over a temperature range; this range is, however, typically small compared with the difference between T_w and T_∞, the far-field temperature in the crust, and hence can be neglected. The second parameter governing the behaviour of the thermal processes is the Stefan number of the surrounding crust $S_\infty = L/(c(T_w - T_\infty))$. This incorporates the approximation that the fluid magma, its solid product and the country rock have identical thermal properties. The final parameter is $B = (W_i^4/H^2)(\Delta P/\mu\kappa)$, where W_i is the initial width of the dyke of length H, ΔP is the driving pressure in the reservoir, μ is the mean value of the dynamic viscosity of the magma within the boundary layer and κ is the thermal diffusivity of both magma and country rock. The cube root of B is the ratio of the initial width of the dyke to the initial width, at the top of the dyke, of the thermal boundary layer, due to advection before any solidification or melting can take place. We assume that this ratio is large.

We also assume that $c = 730\ \mathrm{J\ kg^{-1}\ ^\circ C^{-1}}$, $L = 8\times10^5\ \mathrm{J\ kg^{-1}}$, $\kappa = 10^{-6}\ \mathrm{m^2\ s^{-1}}$, $T_m = 1{,}200\ ^\circ\mathrm{C}$, $T_w = 1{,}150\ ^\circ\mathrm{C}$ and $\mu = 100$ Pa s. (ref. 12), which reflect values appropriate for a typical basaltic magma. The driving pressure was evaluated from the lithostatic overburden[13] as $\Delta P/H = 2{,}000\ \mathrm{N\ m^{-3}}$, with H taken as either 2 or 5 km. We chose these values because the depth interval 2–5 km corresponds to the top and bottom of active magma reservoirs in Hawaii, Iceland and the East Pacific Rise. It also corresponds approximately to the upper surface and the keel of dykes in the respective rift zones of these localities. In a given volcanic system the remaining two variables T_∞ and W_i determine the course of successive eruptions. The value of T_∞ may vary between 0 and 1,150 °C (T_w) depending on the previous thermal history of the dyking zone. With these values, $S_m = 22$, S_∞ ranges upwards from 1.05 (which corresponds to cold country rock at 0 °C) and B can take on any positive value, though it must exceed unity for our boundary-layer model to be appropriate.

The important qualitative behaviour of our two-dimensional model is sketched in Fig. 2. This is augmented by the quantitative results depicted in Fig. 3. When the eruption commences magma initially solidifies against the walls of the dyke. In those dykes for which W_i and T_∞ are sufficiently small so that they fall below demarcation line AA′ for a 2-km dyke and BB′ for a 5-km dyke, solidification continues in the downstream portion of the channel. The surface fissure will eventually become blocked and the eruption through the dyke feeding the fissure will cease irrespective of the magma supply. If $T_\infty = 100\ ^\circ\mathrm{C}$, for a dyke 2 km long the critical initial width is 1.3 m, whereas for a dyke 5 km long the critical initial width is 1.7 m. The time taken for the dyke to become blocked as a function of its initial width is shown in Fig. 3*b* for these two dyke lengths.

If ΔP remains fixed, eruptions can only continue from dykes whose initial width exceeds the critical value, which is decreased if the dyke is surrounded by warmer country rock. This increase in T_∞ could be the result of pre-warming by previous dyking events or by hydrothermal circulation. For dyke widths beyond the critical value the thermal energy from the reservoir advected along the dyke melts back the walls and the channel remains open. Thereafter, the rate and duration of the eruption are controlled by the decay of the driving pressure ΔP and the possible closing of the dyke because of the resulting subsidence, which we have not included in our model. When ΔP has decreased sufficiently solidification will take place once more and the eruption fade away.

The concept of the two-dimensional widening of a dyke as a

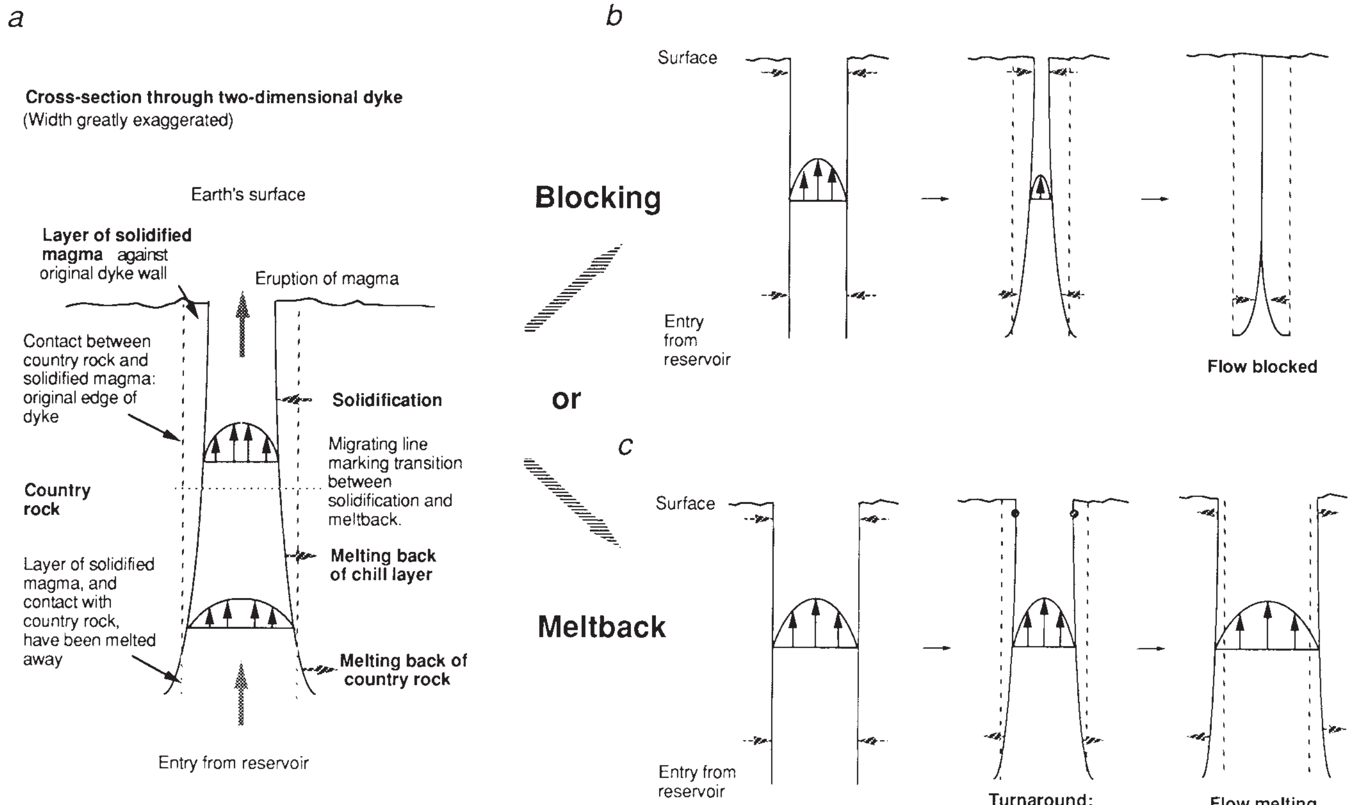

FIG. 2 The evolution of a two-dimensional dyke. *a*, State at an instant during initial development. *b*, Development of blocking. *c*, Development of turnaround and meltback.

result of melting is broadly consistent with some previous observations. Both the evolution of the Pu'u O'o series of en echelon dyke fissures in Hawaii (L. Wilson, personal communication) and the final width of 25 m for some fissure dykes in Iceland (M. R. Ryan, personal communication) seem to be explicable only if melting back of the original dyke walls is postulated.

Further, the two-dimensional analysis quantifies an important feedback mechanism. Decreasing the width of the dyke restricts the flow. This reduces the advected heat supply which leads to further solidification and reduction of the width. Correspondingly, widening the dyke by melting the walls increases the advected heat supply which results in further melting. This feedback mechanism has an important role in a more realistic three-dimensional model, in which the width of the dyke varies, as depicted in Fig. 4. Cross-flows between sections of different initial widths will form because flow through the narrow sections will encounter greater constrictions because of preferential solidification. The cross-flows further reduce the advected heat flux in the downstream portions of the narrow sections and increase it at the wider portions, thereby exaggerating initially small differences in the width of the dyke (Fig. 4*b*). Our model predicts that, as a result of this mechanism, many fissure eruptions of moderate size are localized at isolated vents over a period of days (Fig. 4*c*), which is consistent with observations on Hawaii and elsewhere (refs 2, 3, 14 and T. L. Wright, personal communication) and with a previous qualitative discussion[15]. Eruptions from these vents will persist until limited by other volcanic processes. Sustained eruptions result in the formation of cylindrical conduits that are much wider than the original dyke and are preserved as volcanic plugs. □

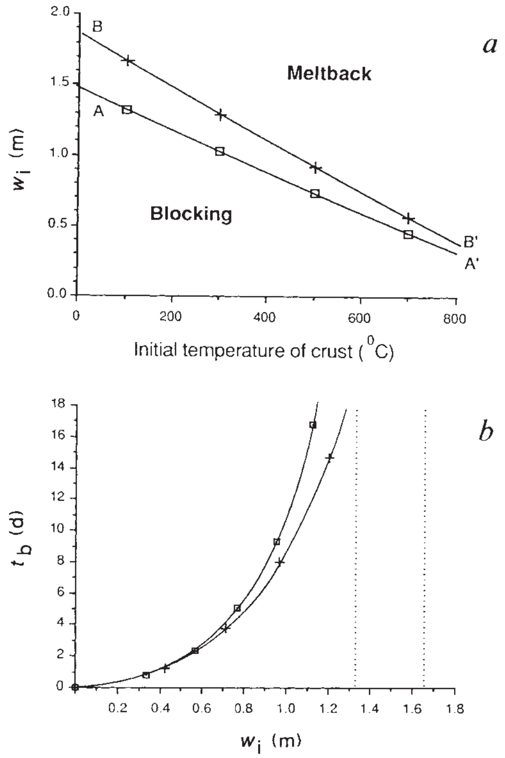

FIG. 3 Numerical results of model, for parameter values given in text. *a*, Regions of parameter space in which blocking or meltback are predicted for dyke length 2 km (squares) and 5 km (crosses). *b*, Time for dyke to become blocked as a function of initial dyke width, for $T_\infty = 100$ °C. The time to block becomes infinite for widths approaching the critical value for meltback to occur (dotted lines (left) $H = 2$ km; dotted lines (right) $H = 5$ km). Symbols as in *a*.

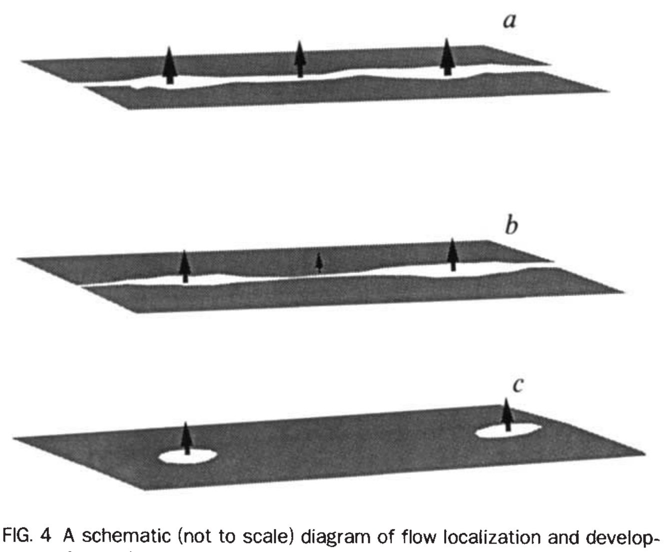

FIG. 4 A schematic (not to scale) diagram of flow localization and development of a surface fissure over time. *a*, initial irregularities of width (which may be small). *b*, faster solidification in narrow regions exaggerates the difference. *c*, flow through isolated vents melts back walls and flow is blocked elsewhere.

Received 6 June; accepted 7 November 1989.

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ACKNOWLEDGEMENTS. We thank R. C. Kerr, J. R. Lister, R. S. J. Sparks, A. W. Woods and M. G. Worster for helpful discussions and P. T. Delaney, A. R. McBirney, D. D. Pollard, L. Wilson and T. L. Wright for stimulating comments. P.M.B was supported by NERC studentship and H.E.H was supported by the B.P. Venture Research Unit.

Hydrothermal plumes along the North Fiji Basin spreading axis

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STEADY-STATE emanations of hydrothermal vent fluid along the mid-ocean ridges give rise to plumes rising 100–300 m above the sea floor, identified by anomalies of temperature, conductivity, chemistry and particles. Recently, a quite different type of hydrothermal plume created by a brief but massive release of high-temperature hydrothermal fluids was discovered by Baker *et al.* at a Juan de Fuca vent field[1–3]. This gigantic thermal plume was termed a 'megaplume'. In December 1987, we found a large

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