THEORETICAL MODELS OF MAGMA ASCENT THROUGH A GEOTHERMAL BOREHOLE

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ABSTRACT

As drilling technology improves and shallower resources become tapped, interest is growing in deeper, hotter geothermal resources. In volcanic settings, the possibility of encountering a magmatic intrusion increases with each metre of borehole. Such intrusions may not be resolved by surface geophysical surveys, particularly if they are small, deep or sit beneath other bodies that shadow their presence. Further, if the intrusion contains a mobile fraction of partial melt, this could enter the borehole and begin to ascend.

Magma has been encountered at least three times during geothermal drilling: in Iceland, Kenya and Hawai'i. In the latter instance, dacitic magma at a depth of 2.5 km entered the borehole and ascended 5.5 m over several minutes before stopping. Several international projects have proposed to drill shallow magma bodies or their thermal aureoles for scientific inquiry, energy generation or hazard mitigation. However, the possibility of uncontrolled magma ascent within a wellbore has not been quantitatively addressed.

Here, we present an analysis of the competing time scales controlling the thermal survival of a magma column ascending the borehole. Rise time is most rapid for low viscosity magmas (like basalts) under large overpressure that are encountered at shallow depths in large-diameter boreholes. This magma must reach the surface before it is stalled by crystallization – either through radial cooling or growth of a crystal plug that exerts a drag on the borehole wall. Magma rise is also rapidly accelerated in the presence of a low-viscosity outer annulus of degassed volatiles, although it is not clear if this configuration could persist stably.

Investigating a typical parameter range, we show that rhyolite magmas are too viscous to feasibly ascend wellbores, which is consistent with the experience at Hawai'i and Iceland. In exceptional circumstances, low-viscosity basaltic magmas could reach the surface before freezing, such as occurred in Námafjall geothermal field in 1977 during the Krafla Fires.

1. INTRODUCTION

As conventional geothermal systems are increasingly developed, the search for new high-temperature resources has turned to greater depths. For instance, the Iceland Deep Drilling Project (IDDP) has explored the use of deep geothermal wells to access supercritical temperatures of 450-600°C at 4-5 km depth in conventional geothermal fields (Friðleifsson and Elders, 2015). The concept has also been proposed in unconventional geothermal settings, such as at Avachinsky Volcano in Kamchatka (Fedotov et al., 2007), and at Newberry Volcano in Oregon (Cladouhos et al. 2018). Drilling to magma would offer unique insights to the scientific community (e.g., the proposed Krafla Magma Testbed; Hólmgeirsson et al., 2018) and has even been suggested as a mechanism to prevent future volcanic eruptions (Wilcox et al., 2017).

There have been several notable instances in which geothermal drilling has encountered magma at shallow depths. In 2005, drilling of KS-13 at Puna geothermal field, Hawai'i, encountered 1050°C dacitic magma at a depth of 2.5 km (Teplow et al., 2009). Magma flowed into the borehole (25.88 cm diameter) and ascended 5.5 m in several minutes before coming to a halt. Teplow et al. (2009) concluded that radial cooling would not have been rapid enough to crystallise and arrest the magma. Instead, they suggested a crystallised plug of magma must have formed in contact with the drilling fluid.

In 2009, drilling of IDDP-1 at Krafla geothermal field, Iceland, encountered >900°C rhyolite magma (Friðleifsson et al., 2015). Drilling logs show that, on the second side track, the drill string became stuck at a depth of 2.1 km. It was freed and pulled back 9.5 m up the hole and then, three minutes later, re-lowered a depth of 2 m where it became stuck again (Pálsson et al., 2014). One explanation is that an encounter with magma caused the first stoppage and, when the string was pulled up, it flowed 7.5 m up the 12 1/4" borehole where the second encounter occurred. The well was eventually completed and flowed steam at 450°C. It was lost two years later due to casing failure after a valve failed and the well had to be killed with cold water.

This was not the first instance of magma flow in boreholes in the Krafla region. In 1977, during the Krafla Fires (a period of rifting and an eruption between 1975 and 1984), basaltic magma flowed horizontally into a 1138 m borehole in the Námafjall geothermal field (Larsen et al., 1979). In a 20-minute period, 1.2 m^3 of magma was erupted at an estimated rate of 25 kg s^{-1}. From an estimated column height of 20 to 40 m, the magma was inferred to have travelled at 20 to 30 m s^{-1} and ascended the (average) 6" diameter borehole in 0.5 to 1 min.

Finally, a study of drill cuttings from wells MW-04 and MW-06 at Menengai geothermal system, Kenya, revealed fresh glassy, quenched cuttings from depths of 2 km (Mbia et al., 2015), indicating the drilling had encountered magma.

A further field study relevant to basaltic magma transport in boreholes is the 1981 drilling campaign of the Kilauea Iki lava lake (Helz and Wright, 1983), originally erupted in 1959. Here, drilling of the crystallized carapace to the crust-melt interface at depths up to 100 m produced multiple observations of "ooze-back", a phenomenon where the hole was refilled with melt. Ascent rates shortly

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after drilling were as high as 1.4 m/min in the 3.5" holes, a bit slower than the 1.8 and 2.5 m/min inferred at Puna and Krafla, respectively.

Although the scientific and economic arguments for exploiting magma are compelling, there are risks that must be addressed. The goal of this paper is to explore models of borehole magma ascent, which would be a requirement for any uncontrolled extrusion or explosive event. Our approach is to develop analytical and semi-analytical models that describe important timescales in the system, specifically magma ascent and crystallisation. Then, using known or likely parameters for different settings, we assess the 'thermal survival' of rising magma columns and compare these to observations.

2. MAGMA ASCENT THROUGH A WELLBORE

Consider a magma body encountered from above by drilling (Figure 1). If the magma is mobile and overpressured, it will ascend the borehole under viscous flow conditions – potentially lubricated by annular flow of volatiles – and, if unimpeded, will reach the surface in time, t_{ascent}. While in the borehole, the magma will lose heat to its surroundings and the drilling fluid. If it loses enough heat, it will freeze and form a plug within borehole, a process that occurs in time, t_{freeze} or t_{plug} (depending on the model). In the analysis that follows, we will develop simple models for t_{ascent}, t_{freeze} and t_{plug}, and compare their relative magnitudes under probable geothermal conditions.

For simplicity, we have not considered the possibility of explosive interactions with borehole fluids, interactions between magma and the borehole casing or cement, or degassing of the magma as it ascends and depressurizes.

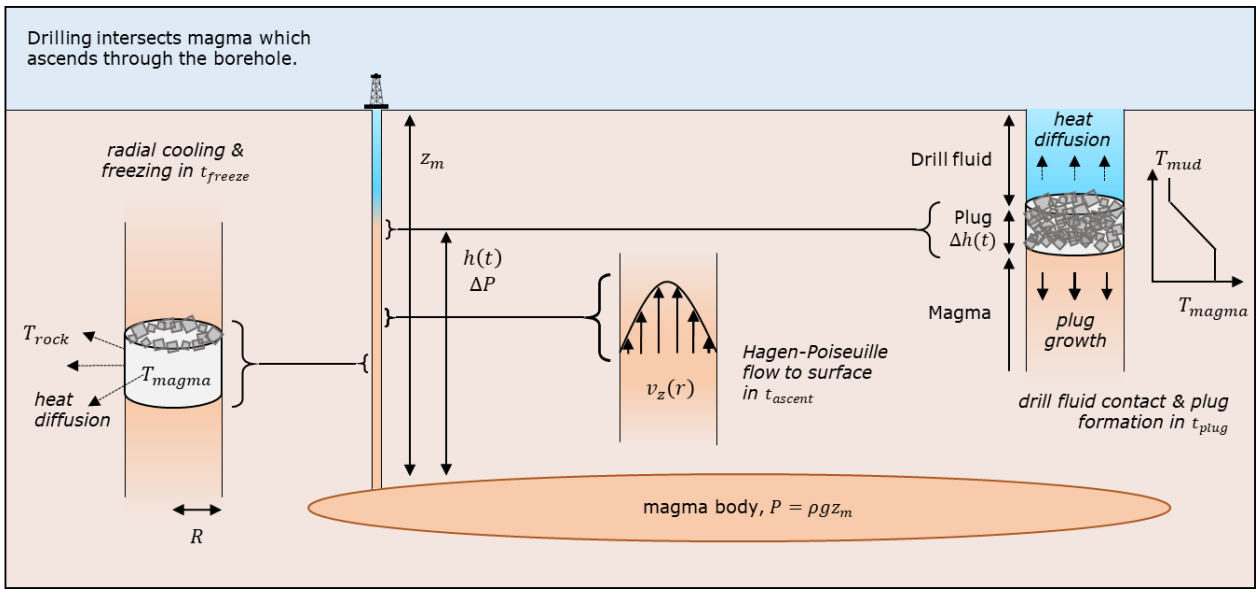

Figure 1: Schematic illustration of borehole transport and freezing processes relevant to magma ascent. Symbols and equations are referred to in the text.

2.1. Viscous flow through a cylindrical pipe

The incompressible Navier-Stokes equation describes unsteady flows of incompressible, viscous fluids:

$$\rho\left(\frac{\partial \boldsymbol{v}}{\partial t} + (\boldsymbol{v}\cdot\nabla)\boldsymbol{v}\right) = -\nabla P + \mu\nabla^2\boldsymbol{v}, \tag{1}$$

where $\boldsymbol{v}$ is the fluid velocity vector, $P = p - \rho g z$, is the hydrodynamic pressure, p is fluid pressure, g is gravity, and ρ and μ and fluid density and viscosity, respectively. Conservation of mass is expressed by the continuity equation: $\nabla\cdot\boldsymbol{v} = 0$.

For a well-developed flow along a vertical wellbore of radius, R, an axisymmetric geometry can be adopted with radial and vertical coordinates r and z, and the velocity vector written, $\boldsymbol{v} = (0, v_z)$. From the continuity equation it follows that $\partial v_z/\partial z = 0$. Assuming inertial effects are small, (1) simplifies to:

$$\frac{\partial P}{\partial z} = \frac{\mu}{r}\frac{\partial}{\partial r}\left(r\frac{\partial v_z}{\partial r}\right). \tag{2}$$

Eq. (2) is solved under standard fluid mechanical assumptions: no slip at the borehole wall ($v_z(r = R) = 0$), P is a function of z only, and the flow velocity is finite at $r = 0$. We define $\partial P/\partial z$ as the pressure drop ΔP from the bottom of the borehole to the ascending magma front at height, $h(t)$.

Integrating Eq. (2) twice with respect to r, we obtain the quadratic solution for pipe flow, with the column height, h, given implicitly as the integral of v_z:

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$$\underbrace{v_z = \frac{R^2 - r^2}{4\mu}\frac{\Delta P}{h}}_{\text{velocity profile}}, \qquad \underbrace{\pi R^2 h(t)}_{\substack{\text{volume of}\\ \text{magma}\\ \text{column}}} = \underbrace{\int_0^t 2\pi \int_0^R r v_z(r,t')dr\,dt'}_{\text{cumulative flow at front}}, \tag{3}$$

where the magma front is constrained by balancing the volume of the magma column with the area and time integrated flow producing the column. By making the substitution for h in (3), integrating over r, and then differentiating with respect to t to eliminate the outer integral, we obtain a differential equation for h that can be solved with initial conditions $h(t = 0) = 0$. Setting h to magma depth, z_m, gives the time to reach the surface, t_{ascent}:

$$v_z = \frac{R^2 - r^2}{R}\sqrt{\frac{\Delta P}{4\mu t}}, \qquad h(t) = R\sqrt{\frac{\Delta P t}{4\mu}}, \qquad t_{ascent} = \frac{4\mu z_m^2}{R^2 \Delta P}. \tag{4}$$

Thus, magma ascent is most rapid for large diameter boreholes that access shallow, low viscosity magma under large overpressure. The maximum overpressure due to buoyancy will occur as magma first enters the borehole and is given by the difference between the lithostatic load and the mud weight (approximately hydrostatic) in the wellbore at depth, z_m, i.e., $\Delta P_{max} = (\rho_{OB} - \rho_{mud})gz_m$, where ρ_{OB} is the overburden density. For typical depths (2-2.5 km), rock density (ρ_{OB} = 2600 kg m^{-3}) and mud weight (ρ_{mud} = 1050 kg m^{-3}), ΔP ranges between 30 and 34 MPa. However, as magma ascends the well, ΔP will be reduced under the weight of the magma column. The minimum overpressure due to buoyancy, when magma reaches the surface, is $\Delta P_{min} = (\rho_{OB} - \rho_{magma})gz_m$, where density difference between magma and country rock is now explicit. For $\rho_{OB} - \rho_{magma} \sim$ 100 kg m^{-3}, ΔP_{min} is between 2 and 2.5 MPa. This calculation does not account for magma overpressured above lithostatic, which would decrease t_{ascent}, nor local depletion of the magma pressure due to mass flow.

The calculation above assumes that flow in the pipe is laminar. Onset of turbulent flow in a pipe occurs when the Reynolds number, $Re = 2R\,\bar{v_z}\,\rho/\mu$, exceeds ~ 2300. For typical parameter values used here, the flow will transition from initially turbulent to laminar in, at most, about one second.

For a standard 9-5/8" diameter geothermal borehole intersecting a rhyolite (μ = 10^7 Pa s) at 2 km depth with average overpressure 40 MPa during transport, t_{ascent} is about eight years. For the same well intersecting a basalt (μ = 10^2 Pa s) at 2.5 km depth with average overpressure 4 MPa, t_{ascent} is about half a day.

2.2. Alternative models with annular flow

We consider an extension to the above flow model where the rising magma is surrounded by an annulus of fluid with different viscosity at the edge of the borehole. This could arise in two situations. First, if the magma degasses due to decreasing pressure as it rises, an annulus of supercritical water with relatively low viscosity could form at the edge of the borehole. Second, crystallization of magma as it rises could result in an annulus of crystallized or relatively high viscosity magma forming at the edge of the borehole. The model presented in Section 2.1 can be extended to account for either a low viscosity or a high viscosity annulus next to the borehole wall. In both cases, the velocity profile must be continuous across the two layers and the shear stress must be equal at the interface between the rising magma and the surrounding annulus. With these conditions, the velocity profiles in the magma and the annulus are given by:

$$v_z = F\left(\frac{r^2}{R^2} + \alpha\mathcal{R}^2 - \alpha - \mathcal{R}^2\right), \qquad v_a = F\alpha\left(\frac{r^2}{R^2} - 1\right). \tag{5}$$

In Eq. (5), F is the pressure forcing given by:

$$F = \frac{-\Delta P R^2}{4\mu h}, \tag{6}$$

α is the ratio of the viscosity of the magma to that of the fluid in the annulus, and $\mathcal{R}$ is the fraction of the borehole radius occupied by rising magma:

$$\mathcal{R} = \frac{(R - \delta)}{R}, \tag{7}$$

with δ being the thickness of the annulus.

The velocity given in Eq. (5) can be compared to that from Eq. (3) in each of the two described situations to determine the importance of the annulus formation. For the first situation, we use an annulus viscosity of 10^{-5} Pa s based on superheated steam at temperatures similar to the magma temperatures at abovementioned geothermal fields (i.e. 900-1000°C). For the situation of a crystallizing annulus, we consider the annulus viscosity to be much smaller than that of the magma and assume that $\alpha = 0$. Fig. 2 shows the multiplicative factor that results from accounting for an annulus of either low (A) or high (B) viscosity fluid. This model suggests a thin annulus of gas can allow the magma to rise at much higher velocities than if the magma directly contacted the borehole. In contrast, a crystallizing annulus will reduce the rise velocity in a manner that is equivalent to the borehole reducing in size.

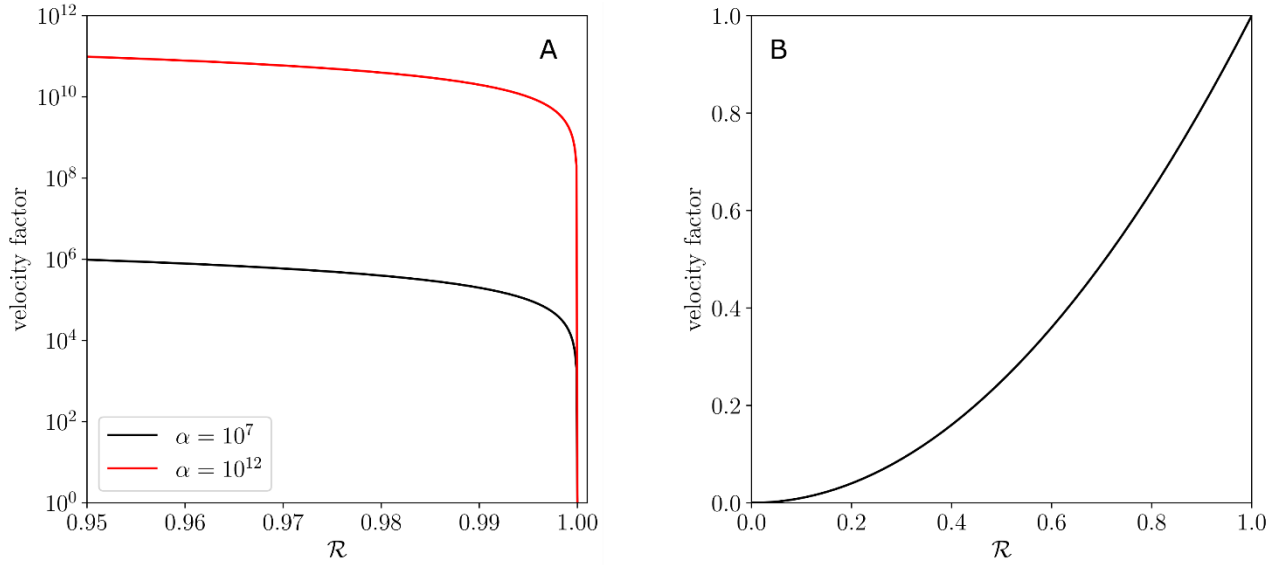

Figure 2: The multiplicative factor for the magma velocity at the centre of a borehole for the situation of a low viscosity gas annulus (A) and a high viscosity solid or crystalizing annulus (B). Note the different x-axis scales and the logarithmic scale on A. $\alpha = 10^7$ and $\alpha = 10^{12}$ correspond to basalt and rhyolite respectively.

2.3. Heat loss through the borehole wall

Magma in the borehole will be at a much higher temperature than its surroundings, which are assumed to have been locally cooled by the drilling fluid. In the first instance, we have assumed that heat loss by the magma across the borehole wall is conductive, although if the rock is sufficiently permeable then local heating could induce groundwater flows that enhance the rate of heat loss. For magma at its eutectic, heat loss will cause crystallization while maintaining a constant temperature. The axisymmetric solution for temperature evolution at $r > R$ for an infinite domain at initial temperature, T_{rock}, internally bounded by a cylindrical surface at $r = R$ at temperature, T_{magma}, is given by Carslaw and Jaeger § 13.5.6 (1959):

$$T = T_{magma} + \frac{2\Delta T}{\pi} \int_0^\infty e^{-\kappa u^2 t} \frac{J_0(ur)Y_0(uR) - Y_0(ur)J_0(uR)}{J_0^2(uR) + Y_0^2(uR)} \frac{du}{u}, \tag{8}$$

where $\Delta T = T_{magma} - T_{rock}$, $\kappa = K/\rho c$ is the thermal diffusivity of the rock, K, ρ and c are the thermal conductivity, density and heat capacity of the rock, and J_0 and Y_0 are zeroth order Bessel functions of the first and second kind (Fig. 3A). Note, we have assumed that rock and magma density are sufficiently similar to not warrant distinction here.

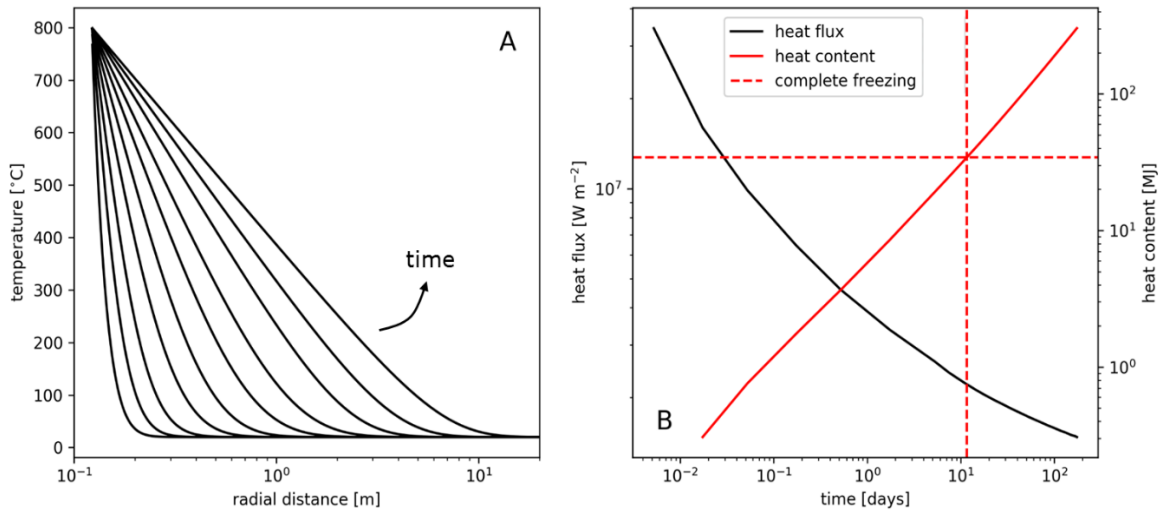

Figure 3: Temperature and heat flux evolution around a cylindrical borehole at fixed temperature. (A) Radial temperature profiles at different times, as given by Eq. (8). (B) Radial heat flux at the wellbore (black; Eq. 9) and total heat loss (red; Eq. 10) with time. The heat loss and time required to freeze a plug of rhyolite within a 9-5/8" wellbore is indicated by red dashed lines (Eq. 11).

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Of interest here is the heat flux at R, which is given (Carslaw and Jaeger, 1959, § 13.5.9):

$$q = \frac{4\Delta TK}{R\pi^2}\int_0^\infty e^{-\kappa u^2 t}\frac{du}{u[J_0^2(uR)+Y_0^2(uR)]}. \tag{9}$$

If the magma is instantaneously emplaced in the wellbore at t = 0, then at time t, the heat lost is given by the integral of Eq. (9):

$$Q = 2\pi\int_0^t q(t')\,dt' = \frac{8\Delta TK}{R\pi\kappa}\int_0^\infty (1-e^{-\kappa u^2 t})\frac{du}{u^3[J_0^2(uR)+Y_0^2(uR)]} \tag{10}$$

Equating this to the latent heat of crystallization for the cylindrical magma volume in the borehole, $Q_{freeze} = L_m\rho\pi R^2$, gives an expression for the time for borehole magma to completely freeze over:

$$\int_0^\infty \frac{(1-e^{-\kappa u^2 t_{freeze}})}{u^3[J_0^2(uR)+Y_0^2(uR)]}du = \frac{L_m\pi^2R^3}{8c\Delta T}. \tag{11}$$

Eq. (11) can be solved for t_{freeze} either by numerical quadrature or simulation of the discretized energy conservation equation. For magma at 800°C with latent heat of crystallization, L_m = 2.9×10^5 J kg^{-1} (Caricchi and Blundy, 2015) in a 9-5/8" diameter borehole in contact with rock cooled to 20°C with specific heat 10^3 J kg^{-1} K^{-1}, t_{freeze} is about 10 days (Fig. 3B).

2.4. Plug formation in the borehole

Teplow et al. (2009) suggest that a more likely mechanism for choking wellbore flow is the formation of a crystallized plug at the top of the ascending magma column. Such a plug would form due to direct contact of the magma with cool borehole fluid, and magma ascent would halt when the frictional resistance provided by the plug with the borehole wall exceeds the magma overpressure. An analytical model can be developed to quantify these processes.

The heat extraction, Q, to freeze a cylindrical plug of height, Δh, from magma at its eutectic is:

$$Q(t) = \pi R^2\rho L_m\,\Delta h(t). \tag{12}$$

If there is ample mixing of the borehole fluid, then one can assume that 1D linear heat flow through the plug, q, is controlled to first order by an equilibrated conductive temperature gradient. This yields an integrodifferential equation that can be solved for plug growth, $\Delta h(t)$:

$$q(t) = K\frac{\Delta T}{\Delta h(t)}, \qquad Q(t) = \pi R^2\int_0^t q(t')dt', \qquad \Delta h(0) = 0, \qquad \Delta h = \sqrt{\frac{2K\Delta T t}{\rho L_m}}, \tag{13}$$

where $\Delta T = T_{magma} - T_{mud}$. The frictional strength, $F_{friction}$, due to contact of the plug with the borehole wall will increase with the growing contact area:

$$F_{friction} = \underbrace{2\pi R\Delta h}_{\substack{\text{contact}\\ \text{area}}} \times \underbrace{f}_{\substack{\text{friction}\\ \text{coef.}}} \times \underbrace{\rho_{mud}\,g\,z_m}_{\substack{\text{normal}\\ \text{stress}}}, \tag{14}$$

where f is the friction coefficient and we have assumed that normal stress in the plug is in equilibrium with the overlying mud weight at the time of freezing, i.e., the isotropic magma pressure is "frozen" in place. This model does not account for density and volume changes due to the phase transition, i.e., as magma freezes, it contracts and will exert a smaller normal stress on the borehole. In addition, vertical advection of the plug would reduce the axial stress applied by the overlying mud column, further reducing the radial normal stress.

The frictional strength is compared against the maximum borehole overpressure the plug must resist if it is to arrest flow, which is given by the difference between magma overpressure below and mud weight above:

$$F_{magma} = \pi R^2(\rho_{OB} - \rho_{mud})gz_m. \tag{15}$$

Equating (14) and (15) and substituting (13) for Δh yields an expression for the time to freeze a large enough borehole plug that can arrest magma ascent:

$$t_{plug} = \left(\frac{R(\rho_{OB}-\rho_{mud})}{f\rho_{mud}}\right)^2\frac{\rho L_m}{8K\Delta T}. \tag{16}$$

For a magma at 1050°C with L_m = 2.9×10^5 J kg^{-1} and K = 2.5 Wm^{-1}K^{-1}, in contact with 20°C drilling mud with density 1.05 g cm^{-3} in a 9-5/8" diameter borehole with friction coefficient of 0.6, t_{plug} is about 1 hour. This is about an order of magnitude too large to account for magma arrest within "several minutes" as observed by Teplow et al. (2009) for magma ascent in KS-13.

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3. DISCUSSION

The characteristic time scales, t_{ascent}, t_{freeze} and t_{plug} depend on the borehole radius as well as other parameters of the rising magma. In particular, viscosity has an outsized impact on whether the rising magma reaches the surface, as this parameter varies over many orders of magnitude (from 10^1-10^2 Pa s for basalts to 10^6-10^7 Pa s for rhyolites). To understand the range of possible outcomes, we have computed t_{ascent} for basaltic and rhyolitic magmas across a range of magma encounter depths (2-2.5 km), magmatic overpressures (4-40 MPa) and borehole radii (Fig. 3). This estimate excludes the effect of annular volatile flow because we don't know at what depth this would occur, if at all. For comparison, we have also calculated t_{freeze} for a range of L_m (200-500 kJ kg^{-1}), thermal conductivities (1-5 Wm^{-1}K^{-1}), magma temperatures (800-1000°C) and borehole radii (Fig. 4), and t_{plug} for the same properties plus a range of friction coefficients (0.3-0.8) and mud weights (1.0-1.1 g cm).

The time scale analysis indicates that geothermal wells from slim hole to big bore (Garg and Combs, 1997; Bush and Siega, 2010) are not large enough for viscous rhyolite magmas to ascend (decades to millennia) before freezing (hours to months) or plug formation (minutes to days) occurs. Instead, a borehole would need to have a radius of 7.5 m or more for t_{plug} to reliably exceed t_{ascent} and, even then, it would take weeks to reach the surface. In fact, a natural analog to these 'boreholes' is the conduits that feed rhyolite dome extrusions. For instance, during the 1995-97 Soufrière Hills eruption in Montserrat, a lava dome was extruded from an estimated 30 m diameter conduit (Melnik and Sparks, 1999) at an average rate of 2.1 m^3/s (Sparks et al., 1998), which corresponds to t_{ascent} of about 20 days from the 5 km deep magma chamber.

The implications for encountering basaltic magma while drilling are quite different. As its viscosity would be many orders of magnitude lower than rhyolite, borehole ascent would be a lot more rapid. In particular, for large diameter wellbores, magma ascent could occur on time scales of hours, which is comparable to the time required to freeze a plug.

To date, basaltic magmas have not been encountered during drilling and their occurrence at shallow (drillable) depths is probably rare. In any event, were a basaltic magma to be inadvertently drilled, borehole transport may not be as simple as the description given above. First, magma need not completely solidify for flow to become arrested, as its viscosity will tend to increase with crystallinity, greatly slowing the rate of ascent. On the other hand, as the plug solidifies, it will tend to contract, reducing borehole friction. Other processes, such as lubrication by along the borehole wall by superheated drilling fluid or volatiles exsolved from the magma, could also reduce the frictional resistance of the plug. Addressing these processes would likely require sophisticated numerical simulators.

3.1. Comparison against prior magma encounters

Magma encounters at Krafla and Puna provide an opportunity to compare the model with data. Although both columns arrested after a few meters, full wellbore ascent times can be estimated by: (i) substituting observations of h, t and R into the middle of Eq. (4) to constrain the ratio $\Delta P/\mu$, and then (ii) substituting this value into the righthand of Eq. (4) alongside known z_m. For Krafla, the estimated ascent time was 160 days, and for Puna it was 230 days. Taking $\Delta P = \Delta P_{max}$, which is appropriate for flow near to magma entry to the borehole, the corresponding magma viscosities are 6.2 and 5.2×10^5 Pa s.

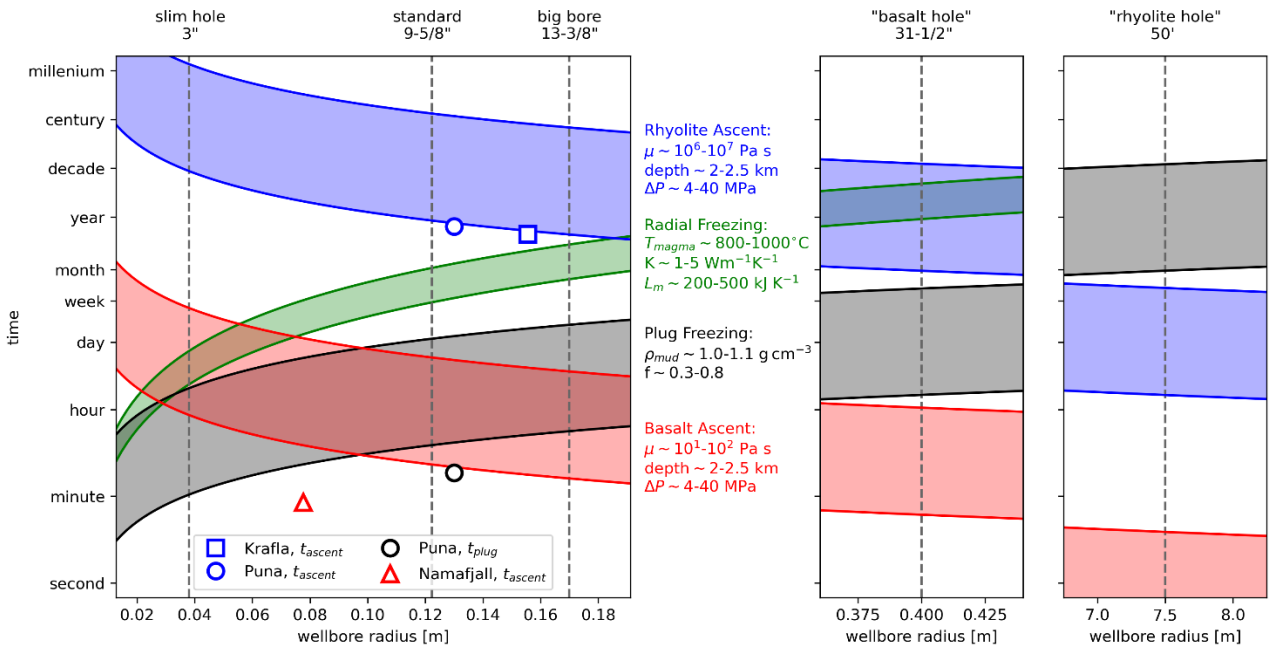

Figure 4: Comparison of t_{ascent} for rhyolitic (blue) and basaltic (red) magmas across a plausible parameter range, along with t_{freeze} (green) assuming a radial heat loss model, and t_{plug} (black) assuming frictional plug formation for the same magmas. Both ascent and freezing times are strong functions of wellbore radius, and a typical range of geothermal wellbores are shown (left, vertical dashed lines). Eruption is assumed where both green and black curves plot above the red or blue curves. The two righthand plots show the wellbore radii required to "erupt" basalts and rhyolites.

Of course, neither magma column actually rose to the surface, and we have assumed this is due to borehole cooling that created a solid plug (because $t_{plug} < t_{freeze}$ for feasible borehole sizes, Fig. 4). For Puna, the magma had arrested after "a few minutes", however, corresponding estimates of t_{plug} range between 12 minutes and 30 hours. More rapid plug freezing would require smaller values of L_m, which might occur if plug formation is a glass transition (Webb, 1997) rather than crystallization. An alternative explanation is that there is a rapid reduction of overpressure driving the magma column upward (i.e., a reduction of F_{magma}) as magma enters the well. If we conceptualise the magma body as a reservoir where melt percolates through a crystal matrix, then such a pressure reduction would be similar to that which occurs around groundwater bore as a result of pumping (mass removal).

Inferred magma ascent during the Námafjall eruption (0.5 to 1 min) is much more than the 30 mins to 1 day expected for basalts in 6 in boreholes. However, the inferred ascent rate was based on observations of the eruption column and this only constrains magma velocity in the upper section of the borehole. If significant acceleration occurred during ascent, e.g., due to fragmentation or degassing processes, the ascent rate would be overestimated. Fault movement and earthquake activity in the Námafjall area was noted up to one hour before the eruption (Larsen et al., 1979) and it is possible that magma entry to the borehole occurred at this time. Alternatively, faster than expected rise could occur if a lubricating volatile annulus formed (Fig. 2).

3.2. Future improvements

As a first attempt to constrain relevant time scales, the models introduced above are intentionally simplified. They could be improved by incorporating more complex physics such as allowing ΔP to vary linearly between ΔP_{max} and ΔP_{min} as the magma column rises or, allowing ΔP to decrease due to "magma reservoir drawdown". Furthermore, time scales t_{ascent} and t_{plug} are computed independently, whereas the ascending magma column will be sensitive to flow and freezing processes simultaneously. Thus, an improved model could develop a magma trajectory equation (modifying the middle Eq. (4)) that incorporates an evolving overpressure, viscous drag, and plug drag.

4. CONCLUSION

We have developed models to describe the physics of magma ascent and arrest in a borehole, and to characterize time scales of the system. Magma ascent is conceptualized as a balance between viscous drag and driving overpressure, which yields the Hagen Poisueille equation for pipe flow. Magma viscosity is shown to be the primary constraint on magma ascent, with rise times of years to centuries for high viscosity rhyolites and as low as minutes to hours for low viscosity basalts. However, these could be dramatically reduced in the event that a low-viscosity annulus of degassed volatiles forms.

Erupting magma from a borehole is a thermal survival problem, with both radial cooling and plug formation through drill fluid contact tending to freeze and arrest the rising column. Plug formation was formulated as a steady heat flow and crystallization problem and, for feasible parameter ranges, is nearly always more rapid that the ascent time scale. Only for the very particular case that drilling encounters a shallow, pressurized basalt, does the prospect of an eruption become feasible.

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REFERENCES

Bush, J., and Siega, C.: Big Bore Well Drilling in New Zealand – A Case Study, *Proceedings*, World Geothermal Congress, Bali, Indonesia (2010).

Caricchi, L., and Blundy, J.: Experimental petrology of monotonous intermediate magmas. *Geological Society Special Publications* **422**, (2015), 105-130.

Carslaw, H. S., and Jaeger, J. C.: Conduction of heat in solids. *Oxford: Clarendon Press*, (1959), 2nd ed.

Cladouhos, T. T., Petty, S., Bonneville, A., Schultz, A., and Sorlie, C. F.: Super Hot EGS and the Newberry Deep Drilling Project, *Proceedings*, 43rd Workshop on Geothermal Reservoir Engineering, Stanford, California (2018).

Elders, W. A. et al.: Origin of a rhyolite that intruded a geothermal well while drilling at the Krafla volcano, Iceland. *Geology*, **39**, (2011), 231-234.

Fedotov, S. A., Sugrobov, V. M., Utkin, I. S., and Utkina, L. I.: On the Possibility of Using Heat Stored in the Magma Chamber of the Avachinsky Volcano and the Surrounding Rock for Heat and Power Supply. *Journal of Volcanology and Seismology*, **1**, (2007), 28-41.

Friðleifsson, G. O., and Elders, W. A.: The Iceland Deep Drilling Project: a search for deep unconventional geothermal resources. *Geothermics*, **34**, (2005), 269-285.

Friðleifsson, G. O., Pálsson, B., Albertsson, A. L., Stefánsson, B., Gunnlaugsson, E., Ketilsson, J., and Gíslason, Þ.: IDDP-1 Drilled Into Magma – World's First Magma-EGS System Created, *Proceedings*, World Geothermal Congress, Melbourne, Australia (2015).

Garg, S. K., and Combs, J.: Use of slim holes with liquid feedzones for geothermal reservoir assessment. *Geothermics*, **26**, (1997), 153-178.

Helz, R. T., and Wright, T.: Drilling report and core logs for the 1981 drilling of Kilauea Iki lava lake, Kilauea volcano, Hawai'i,with comparative notes on earlier (1967-1979) drilling experiences. *USGS Open-file Report 83-326*, (1983).

ISSN 2703-4275

Hólmgeirsson, S. et al.: Krafla Magma Testbed (KMT): Engineering challenges of drilling into magma and extracting its energy. *GRC Transactions*, **42**,

Larsen, G., Grönvold, K., and Thorarinsson, S.: Volcanic eruption through a geothermal borehole at Namafjall, Iceland. *Nature*, **278**, (1979), 707-710.

Mbia, P. K., Mortensen, A. K., Oskarsson, N., and Hardarson, B. S.: Sub-Surface Geology, Petrology, Hydrothermal Alteration of the Menengai Geothermal Field, Kenya: Case Study of Wells MW-02, MW-04, MW-06 and MW-07, *Proceedings*, World Geothermal Congress, Melbourne, Australia (2015).

Melnik, O. and Sparks, R.S.J.: Nonlinear dynamics of lava dome extrusion. *Nature*, **402**, (1999), 37-41.

Pálsson, B., Hólmgeirsson, S., Guðmundsson, Á., Boásson, H. Á., Ingason, K., Sverrisson, H., and Thórhallsson, S.: Drilling of the well IDDP-1. *Geothermics*, **49**, (2014), 23-30.

Rubin, A. M.: Propagation of magma-filled cracks. *Annual Review of Earth and Planetary Science*, **23**, (1995), 287-336.

Sparks, R.S.J., Young, S.R., Barclay, J., Calder, E.S., Cole, P., Darroux, B., Davies, M.A., Druitt, T.H., Harford, C., Herd, R. and James, M.: Magma production and growth of the lava dome of the Soufriere Hills Volcano, Montserrat, West Indies: November 1995 to December 1997. *Geophysical Research Letters*, **25**, (1998), 3421-3424.

Teplow, W., Marsh, B., Hulen, J., Spielman, P., Kaleikini, M., Fitch, D., and Rickard, W.: Dacite Melt at the Puna Geothermal Venture Wellfield, Big Island of Hawai'i. *GRC Transactions*, **33**, (2009), 989-994.

Webb, S.: Silicate Melts: Relaxation, Rheology and the Glass Transition. *Reviews of Geophysics*, **35**, (1997), 191-218.

Wilcox, B. H., Mitchell, K. L., Parcheta, C. E., Schwandner, F., and Lopes, R. M.: Defending human civilization from supervolcanic eruptions. *California Institute of Technology Jet Propulsion Lab Report*, (2017).

ISSN 2703-4275