Dike Channelization and Solidification: Time Scale Controls on the Geometry and Placement of Magma Migration Pathways

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Abstract We investigate the conditions under which magma prefers to migrate through the crust via a dike or a conduit geometry. We performed a series of analogue experiments, repeatedly injecting warm, liquid gelatin, into a cold, solid gelatin medium and allowing the structure to evolve with time. We varied the liquid flux and the time interval between discrete injections of gelatin. The time interval controls the geometry of the migration, in that long intervals allow the intrusions to solidify, favoring the propagation of new dikes. Short time intervals allow the magma to channelize into a conduit. These times are characterized by the Fourier number (Fo), a ratio of time and thermal diffusion to dike thickness, so that long times scales have Fo > 10² and short time scales have Fo < 10¹. Between these time scales, a transitional behavior exists, in which new dikes nest inside of previous dikes. The flux controls the distance a dike can propagate before solidifying, in that high fluxes favor continual propagation, whereas low fluxes favor dike arrest due to solidification. For vertically propagating dikes, this indicates whether or not a dike can erupt. A transitional behavior exist, in which dikes may erupt at the surface in an unstable, on-and-off fashion. We supplemented the experimental findings with a 2-D numerical model of thermal conduction to characterize the temperature gradient in the crust as a function of intrusion recurrence frequency. For very infrequent intrusions (Fo > 10⁴ to 10⁵) all thermal energy is lost, while more frequent intrusions allow heat to build up nearby.

Plain Language Summary When magma ascends through the Earth's crust to the surface, it tends to do so via cracks, which propagate upward with the magma inside. After these magmatic “dikes” erupt, they quickly start to channelize into a centralized vent, transforming a long, fissure eruption into a focused lava fountain. Underground, such a dike begins to partially solidify, forming a cylindrical conduit where the magma flows fastest and solidifying elsewhere. After the eruption stops, a new eruption can follow this same conduit if it happens quickly enough, preventing the conduit from solidifying and the pathway from closing. Otherwise, if the conduit solidifies, new magma needs to make another dike to be able to ascend through the crust. This time interval between magma ascent events likely affects volcano formation over a long time. If magma erupts frequently, it tends to follow the same path repeatedly, allowing lava to eventually build up a large, centralized volcano. If it is very infrequent, magma needs to make new dikes with new pathways each time, building a group of smaller, separate volcanic cones.

1. Introduction

Fissure eruptions have been documented to evolve with time, from initially elongated fissures to centralized vents (Jones et al., 2017; Keating et al., 2008; Wylie et al., 1999). The change of geometry from planar to cylindrical reflects the easiest form of magma transport; a planar form is most efficient for propagation through a brittle, elastic medium; a cylindrical form is the most thermally efficient (Keating et al., 2008). Dikes that supply such eruptions may also evolve with time at depth, as they have a high ratio of surface area to volume and therefore are very vulnerable to cooling and, eventually, solidification. Cooling is heterogeneous along the dike surface, in that regions that become cool also become more viscous, encouraging flow toward the hotter regions of the dike, which therefore remain warm and active. As they transition to cylindrical geometries, the surface area and heat loss drop, allowing thermally efficient magma transport to the surface (Fukushima et al., 2010). For example, an eruption on Heimaey, Iceland, began as a 1.5-km-long fissure, which focused to a single, central vent over a period of a day (Wylie et al., 1999). Similarly, exposed
shallow plumbing systems at monogenetic volcanoes in the southwestern United States display conduit structures, which likely formed in part due to excavation, then infill, of the surrounding crust (Keating et al., 2008).

There is an element of time necessary to achieve such a geometry, since channelization results from the prolonged effusion of magma. Dikes that do not develop a focused vent remain planar during solidification. In some cases, dikes even become narrower at their centers, as they evacuate magma and deflate, while their lateral fringes solidify and preserve their initial, active state (Daniels et al., 2012). The necessary time scale for flow channelization is not immediately clear; Keating et al. (2008) suggest that the size of the eruption may be important, in that small eruptions can focus on a scale of hours, while larger eruptions do so more gradually.

1.1. Hot Spot Volcano Settings

The channelization of fissure eruptions into focused vents has been documented in a number of settings, primarily characterized by basaltic intrusions. Fissure eruptions commonly occur in the rift zones of hot spot volcanoes like Kilauea and Piton de la Fournaise, in which dike intrusions tend to be linked to deformation of the larger edifices and therefore preferentially occur in certain regions (Jones et al., 2017; Michon et al., 2015). Propagating dikes (and after the fact, solidified swarms) have been imaged via geophysical data at both of these volcanoes and, in some cases, have been observed to evolve into centralized vents (Fukushima et al., 2010; Jones et al., 2017; Leslie et al., 2004; Owen et al., 2000). External forces (i.e., local and regional stresses) clearly guide dikes to intrude in such zones periodically, but magma migration may also take the form of long-lived channels, which must exist to allow structures like Pu‘u O‘o vent to persist (Patrick et al., 2019). These small rift zones can therefore support both persistent magma flow and more voluminous, intermittent dike propagation events.

1.2. Volcanic Field Settings

If hot spot volcanoes represent semicontinuously active settings, then their opposite would be infrequently erupting volcanic fields. Such regions can have eruption recurrence frequencies of one eruption per $10^4$ to $10^5$ years but can periodically increase in activity to one eruption per $10^3$ years or less (Valentine & Connor, 2015). Since these events are so infrequent, cooling dominates the system and intrusions have time to thoroughly solidify. Subsequent intrusions must create new paths via dike propagation. It has been shown that dike swarms have natural spacing between dikes, as each one induces a compressive stress on the surrounding crust, which inhibits later dikes from propagating too close; the final spacing between dikes in is energetically optimized (Bunger et al., 2013). As such, the infrequent timing of events causes dikes (and their eruptive products) to be spatially separated. It should be noted that volcanic fields have relatively small long-term outputs of lava and are, at their most voluminous, similar to arc-related stratovolcanoes ($10^2$ km$^3$/Myr) but can be orders of magnitude less (Valentine & Connor, 2015). Even though this type of setting is characterized by low magma output, with long repose times between events, a single event can display dike flow channelization, similar to the other examples we previously introduced (Keating et al., 2008). Volcanic fields, along with hot spot volcano rift zones, demonstrate that the geometry of magma transport depends on the magma flux, the duration of the injection and time interval between injections. These parameters dictate if a dike can develop into a conduit-like pathway and, depending on the repose time, whether a subsequent pulse needs to propagate a new dike.

1.3. Experimental Studies

Experimental studies on dike solidification can be subdivided into two categories: the ability of a dike or sill to propagate into a cold medium (Chanceaux & Menand, 2014; Delaney & Pollard, 1982; Fialko & Rubin, 1998; Rubin, 1993; Taisne & Taït, 2011) and the stability of an erupting dike (Bruce & Huppert, 1989; Wylie & Lister, 1995). A propagating dike is subject to cooling and solidification as a function of the temperature difference with the host crust, the magma flow rate, the dike geometry, and the magma’s latent heat (Delaney & Pollard, 1982; Fialko & Rubin, 1998). When the magma is very hot and the flow rate is high, the dike geometry evolves by thermal erosion, in which the host rock melts. When the magma is relatively cool and flow rate is low, solidification can proceed. Dikes gain surface area as they propagate and therefore become increasingly susceptible to heat loss via conduction. They therefore have an inherent limit on how
far they can propagate before solidifying, which is termed the “thermal entry length” (Delaney & Pollard, 1982; Einarsson & Brandsdóttir, 1978; Fialko & Rubin, 1998).

Taisne and Tait (2011) show through analogue experiments that, while solidification can stop propagation, further influx of magma into the center of the dike can cause it to inflate and resume propagation. They injected paraffin wax into gelatin and observed that the dike’s chilled margins provided sufficient insulation to maintain flow in the dike’s core. The propagation dynamic varied between no propagation, step-wise propagation, and continuous propagation, as a function of the temperature and the dimensionless flux, \( \Phi \). This flux parameter is a ratio of heating via influx and cooling via conduction:

\[
\Phi = \frac{QH}{\alpha LB},
\]

for volumetric flux, \( Q \), thermal diffusivity, \( \alpha \), and dike vertical length, horizontal breadth, and thickness, respectively, \( L, B, \) and \( H \). The propagation of a dike becomes less affected by thermal conditions (i.e., continuous propagation) for very hot magmas or very high fluxes, while low fluxes of cold magmas may not yield any propagation at all. For transitional conditions, dikes undergo step-wise propagation, in which an arrested dike needs to inflate and build pressure to resume propagation.

Chanceaux and Menand (2014) made similar experiments focusing on sill formation. They show that a dike, arriving at the interface between two layers of the crust, can deflect into a sill under certain temperature and flux conditions. These conditions coincide with step-wise propagation described by Taisne and Tait (2011), whereas dikes in the “continuous propagation” regime tend to pass through the interface and continue propagating vertically. Considering these two studies together from a thermal perspective, it is clear that dikes that are very hot and have a very high flux propagate in a fast and stable way, whereas magma closer to the solidus, flowing at a lower flux, is prone to unstable, step-wise propagation or solidification.

For a dike that can propagate to the surface, its form of eruption is inherently unstable, in that magma flow has a tendency to focus from a sheet of lava into one or a few vents over time. Wylie et al. (1999) show through numerical modeling that a magma with temperature contrasts within a dike, which are the dominant mechanism causing the flow channelization into a discrete vent. The viscosity contrasts cause the flow to channelize on a time scale of ~10 hr, consistent with the time scale observed in the 1973 fissure eruption on Heimaey, Iceland.

The size of an erupting vent evolves as a function of magma flux; if the flux is “high,” the vent walls will melt and the dike widens until the supply is exhausted, but if the supply is “low,” the magma will solidify and block the vent (Bruce & Huppert, 1989). In a 2-D framework, Bruce and Huppert (1989) numerically modeled this fluctuation and found it to be a function of the dike’s initial width. Over some critical size on the order of meters, a basaltic dike driven by a constant overpressure can continue to migrate and erupt until its source has emptied.

1.4. Our Contribution

We present below a series of analogue experiments investigating how dikes solidify and evolve with subsequent influxes of magma. We repeatedly injected hot, liquid gelatin into a cold, solid gelatin medium, varying the flow rate and the time between discrete injections. We then identify the conditions under which dikes can channelize into conduit-like structures, how they remain stable and what types of structures develop otherwise. By quantifying such behaviors, we can gain insights of how the rate and frequency of pulses of magma may shape the underlying plumbing system of a volcano.

2. Methods

2.1. Experimental Setup

We performed a series of analogue experiments, in which we repeatedly injected warm, liquid gelatin into a solid gelatin medium (both made from the same granular gelatin, 250 bloom), housed in a 50 x 50 x 50-cm³ cubic tank. We made five injections per experiment, each 500 ml in volume, providing sufficient volume to propagate up through the gelatin and erupt. Between experiments, we varied only the magnitude of flux and the time interval between injections. Below the tank, we attached a small 10-cm-tall, 10-cm-diameter cylindrical reservoir, which contained a copper heating coil and maintained a constant temperature of the liquid injection (Figure 1a).
We chose gelatin as an analogue for both the solid medium and the injected liquid because it is very versatile. Gelatin is commonly used as an analogue for the Earth’s crust because its Young’s modulus can be controlled by varying the concentration during preparation and because it is transparent, so we can observe what occurs inside the medium (or a dike intruded behind another dike). Additionally, it behaves like the Earth’s upper crust, in that it is a brittle, elastic material on the time scale of a propagating dike and a viscoelastic material on a longer time scale (Di Giuseppe et al., 2009; Kavanagh et al., 2013). As an injected liquid, gelatin quickly solidifies at the experimental temperature, so we can observe flow paths that develop as a dike cools heterogeneously.

We prepared each experiment in a similar fashion (see Table 1 for experimental conditions), in which the gelatin medium was either 3.5 wt% or 4.0 wt% of gelatin and the dike was either 4.0 wt% or 4.57 wt% of gelatin (maintaining a consistent concentration ratio). The high concentration of the dike, relative to the medium, ensured that a solidified dike was stronger than the surrounding medium, preventing subsequent injections from following the previous dike’s path. Since liquid gelatin is not very buoyant in solid gelatin, it remains denser overall, allowing us to observe the flow paths as a dike cools heterogeneously.

### Table 1

| Exp | $C_{L\text{gel}}$ (wt%) | $C_{S\text{gel}}$ (wt%) | $C_{S\text{sug}}$ (wt%) | $E$ (Pa) | $K_c$ (Pa m$^{1/2}$) | $\nu$ (–) | $Q$ (ml/min) | $t_{\text{int}}$ (s) | $\Delta\rho$ (kg/m$^3$) |
|-----|------------------------|------------------------|------------------------|----------|----------------------|----------|--------------|----------------|------------------|
| 1   | 4.0                    | 3.5                    | 6                      | 1,550    | 56                   | 0.5      | 12           | 2              | 80               |
| 2   | 4.57                   | 4.0                    | 6                      | 4,050    | 90                   | 0.5      | 12           | 210            | 80               |
| 3   | 4.0                    | 3.5                    | 5                      | 1,590    | 56                   | 0.5      | 12           | 20,700         | 70               |
| 4   | 4.0                    | 3.5                    | 5                      | 1,920    | 62                   | 0.5      | 1.2          | 210            | 70               |
| 5   | 4.0                    | 3.5                    | 5                      | 4,570    | 96                   | 0.5      | 3.8          | 210            | 70               |
| 6   | 4.0                    | 3.5                    | 5                      | 3,990    | 89                   | 0.5      | 38           | 210            | 70               |
| 7   | 4.0                    | 3.5                    | 5                      | 4,110    | 91                   | 0.5      | 12           | 2,070          | 70               |
| 8   | 4.0                    | 3.5                    | 5                      | 4,580    | 96                   | 0.5      | 38           | 2,070          | 70               |
| 9   | 4.0                    | 3.5                    | 5                      | 6,650    | 115                  | 0.5      | 3.8          | 20,700         | 70               |
| 10  | 4.0                    | 3.5                    | 5                      | 3,850    | 88                   | 0.5      | 38           | 6,600          | 70               |

Note. From left to right are the gelatin concentration of the liquid and solid medium, additive sugar concentration, Young’s modulus, fracture toughness, Poisson’s ratio, flux rate, time interval and density difference.
we added 5.0 wt% to 6.0 wt% of sugar to the medium to increase its density and thus make the dikes positively buoyant. We added food dye to the injected gelatin for pigment, to help distinguish the dikes from the medium and give a sense of where they are thicker (darker looking) or thinner (lighter looking). The concentration of the food dye was low and did not otherwise affect the liquid’s properties.

For the magma analogue, we injected liquid gelatin at a temperature of 31 to 33 °C, marginally hotter than the melting temperature of 28 °C, which allows it to quickly cool as it is injected into the solid medium. We maintained the temperature of our gelatin supply using a thermal bath into which we submerged our supply beaker. The thermal bath also circulates heated water through flexible tubes, which we attach to the basal, cylindrical reservoir’s copper heating coil (Figure 1a). We also threaded these tubes with the gelatin supply tube, to maintain its temperature between the supply beaker and the experimental tank (Figure 2b). When we were ready to start an experiment, we made an initial cut into the gelatin using a retractable blade (Figure 3c), to force the dike to be vertically oriented, and perpendicular to the visualization system. After making this cut, we removed the blade and pumped liquid gelatin into the cut, which then grew into a dike. We controlled the injections flux, duration, and time interval, using a programmable peristaltic pump. The flux range achievable with our current setup ranges from 0.8 to 170 ml/min.

These experiments had the potential to intrude a lot of material into the solid medium, which would build up large strains of the medium. We attempted to mitigate strain accumulation by constructing vertical free...
surfaces, filled with water, on opposing ends of the tank. As a dike is intruded and pushed the medium apart laterally, the free surfaces flexed outward to accommodate the strain. To prepare these surfaces, we embedded vertical plastic sheets, 1 cm in thickness, during preparation of the gelatin medium. After the medium cooled and solidified, we removed the sheets and replaced them with water. However, gelatin became firmly attached to the sheets during solidification, so we cut them out in order to remove them without badly damaging the gelatin. We adhered a metal wire to the plastic sheet prior to solidification and pulled it upward to make a smooth, continuous cut along the interface between the gelatin and the sheet. The cut was never perfectly flat and had a slightly undulating form. Removal of the sheets tended to leave the water-filled boundaries littered with fragments of gelatin, which impeded our view of the dikes. We therefore flushed the boundaries with fresh water until these fragments were sufficiently cleaned out.

We recorded each experiment using a set of cameras positioned to look at the dike’s broad side. Each time an injection erupted and each time an injection terminated, we took several photos. We processed each photo against an initial reference photo to determine the geometry of the dike as a whole. We also processed each photo against the previous injection to see the change between injections, in order to determine which regions of the dike were actively flowing and which had solidified.

2.2. Scaling

For the scaling of these experiments we consider two primary factors, the time interval between pulses of activity and the volumetric flux of magma during a pulse, and to a lesser degree, the temperature of the

Figure 3. The effect of flux, which controls the ability of a dike to erupt. (a) Sufficiently high flux (>10 ml/min) allows the dike to reach the surface before solidifying (Experiment 6). (b) A transitional behavior at semilow flux (3.8 ml/min), in which the dike can reach the surface, but is unstable (Experiment 9). Eruptions can last for tens of minutes but eventually solidify and force intrusion to begin at the base of the medium. (c, d) Images showing two stages of dike propagation (both from Experiment 4), for experiments with low flux (1.2 ml/min). The dike cannot propagate to the surface before solidifying, so new lobes begin to propagate again from the base. The complex dike eventually grows wide and thick, but struggles to reach the surface.
Table 2
Values for Both Our Analogue Experiments and Basaltic Dikes in Nature

| Parameter (units) | Meaning | Experiments | Nature (basalt) |
|-------------------|---------|-------------|-----------------|
| \( \alpha (m^2/s) \) | Thermal diffusivity (crust) | \( 7.4 \times 10^{-8} \) | \( 10^{-6} \) |
| \( t_{inj} (s) \) | Time between injections | \( 10^2-10^4 \) | \( 10^6-10^{11} \) |
| \( H (m) \) | Dike opening thickness | \( 10^{-3}-10^{-2} \) | \( 10^{-1}-10^1 \) |
| \( L (m) \) | Dike vertical length | \( 0.27-0.38 \) | \( 10^{-3} \) |
| \( B (m) \) | Dike horizontal breadth | \( 0.08-0.31 \) | \( 10^{-6} \) |
| \( Q (m^3/s) \) | Magma flow rate | \( 10^{-8}-10^{-6} \) | \( 10^{-1}-10^3 \) |
| \( V (m^3) \) | Dike volume | \( 10^{-4}-10^{-3} \) | \( 10^6-10^8 \) |
| \( T_0 (°C) \) | Dike initial temperature | \( 30-32 \) | \( 1,100-1,200 \) |
| \( T_\infty (°C) \) | Ambient temperature | \( 15-17 \) | \( 300 \) |
| \( T_s (°C) \) | Solidus temperature | \( 28 \) | \( 1,000 \) |
| \( Fo \) | Fourier number \( at/H^2 \) | \( 10^{-1}-10^3 \) | \( 10^{-2}-10^7 \) |
| \( \Phi \) | Flux parameter \( QH/(\alpha LB) \) | \( 10^{-2}-10^1 \) | \( 10^{-3}-10^2 \) |
| \( Q/\alpha L \) | Alternate flux parameter | \( 10^{-1}-10^2 \) | \( 10^{-5}-10^3 \) |
| \( \Theta \) | Temperature parameter | \( 0.76-1.00 \) | \( 0.9-0.95 \) |
| \( S \) | Stefan number \( T_f/(T_0-T_\infty) \) | \( 0.04^a \) | \(-4^1 \) |

\(^a\)This study, see supporting information. \(^b\)Douglas et al. (2016). \(^c\)Molloy et al. (2009). \(^d\)Keating et al. (2008). \(^e\)Wylie et al. (1999). \(^f\)Bunger et al. (2013). \(^g\)Wnuk and Wauthier (2017). \(^h\)Wauthier et al. (2015). \(^i\)Chevallier and Verwoerd (1990). \(^j\)Delaney and Pollard (1982). \(^k\)Natland and Dick (2009). \(^l\)Morse (2011).


For the time interval we scale our experiments simply using a dimensionless time term, the Fourier number, \( Fo \):

\[
Fo = ta/H^2,
\]

where \( t \) is the time interval between injections of magma (Çengel, 2008; Turcotte & Schubert, 2002). This is a measure of the amount of heat lost from a heated slab via conduction versus what is stored, assuming one-dimensional heat transfer; that is, the slab is much taller and broader than it is thick, so heat transfer is perpendicular to and homogenous along its surface. For \( Fo \ll 1 \), much of the initial heat remains in the slab and the system is thermally heterogeneous. For \( Fo \gg 1 \), heat has been conducted away and the system became thermally homogeneous (see Table 2 for a comparison between experiments and nature). Values for \( Fo \) on the order of \( 10^2 \) or \( 10^3 \) are fundamentally no different than \( Fo = \infty \), since in both cases, the dike has completely cooled down and solidified. Sometimes the \( Fo \) is used to define a characteristic time scale for cooling in a system, \( t_d \), at which heat loss is balanced with stored heat, such that \( Fo = 1 \) and therefore \( t_d = H^2/\alpha \) (Fialko & Rubin, 1998). Note that the thermal diffusivity relates to other material properties, the thermal conductivity, \( k \), density, \( \rho \), and specific heat capacity, \( C_p \), such that \( \alpha = k/(\rho C_p) \) (Çengel, 2008).

For the magnitude of flux, we use a dimensionless flux term, \( \Phi \), estimated following Taisne and Tait (2011), which balances heat into the system via influx and advection of heat out of the system via conduction (equation (1)). If \( \Phi \) is large, the system is dominated by heating, while if \( \Phi \) is small, the system is dominated by cooling. This is inherently related to the dimensionless time duration of the propagation, \( Fo_{prop} \), such that

\[
\Phi = QH/\alpha LB = QH^2/\alpha V = H^2/\alpha t_{prop} = 1/Fo_{prop}.
\]

Here, \( t_{prop} \) is the propagation duration and \( V \) is the volume of the dike, assuming \( V \approx LBH \approx Qt_{prop} \).

The relationship between flux and time poses a question for the experiment: Do we hold flux constant and test the time and thus volume of an injection or do we do the reverse and test the flux of an injection, holding volume constant? A good argument can be made for either approach. We would argue that by varying the flux, we can observe the effect of driving pressure on the geometry of the plumbing system that develops in the medium. In such a way, high driving pressures may favor dikes with relatively large dimensions, which grow both vertically long and horizontally broad, while small driving pressures favor dikes that grow primarily vertically.
The parameters $Fo$ and $\Phi$ are highly dependent on the geometry of the intrusion, in that for a broad, thin dike, $Fo$ may initially be large and $\Phi$ small. If the dike channelizes into a conduit-like feature, the parameters could evolve so that $Fo$ decreases and $\Phi$ increases. When we show the scaling for these parameters (Table 2), we show ranges encompassing both dike flow and conduit flow.

We thermally scale the experiments following Delaney and Pollard (1982) and Taisne and Tait (2011), via the dimensionless solidus temperature, $\Theta$:

$$\Theta = (T_s-T_\infty)/(T_0-T_\infty).$$

where $T_s$, $T_\infty$, and $T_0$ are, respectively, the temperature of the solidus, the ambient host rock, and the magmatic source. The injection is above or below the solidus respectively when $\Theta < 1$ or $\Theta > 1$. Delaney and Pollard (1982) make the case that magma is probably injected at temperatures near the liquidus, such that $0.9 < \Theta < 0.95$. For a gelatin melting point of 30 °C (melting is rapid, so we assume this characterizes the solidus and liquidus temperature) and ambient temperature of 15 °C, we therefore choose an injection temperature of ~32 °C. In practice, the temperature in our setup fluctuated by a few degrees as liquid flowed through the system. As described in the previous section, we configured experimental setup to minimize this fluctuation and keep the system thermally stable.

Note that in nature, the thermal evolution of a dike is dependent on another dimensionless parameter, the Stefan number, $S$:

$$S = \gamma/[C_p(T_0-T_\infty)],$$

where $\gamma$ is the latent heat (Fialko & Rubin, 1998). The $S$ is a ratio of the available latent heat to sensible heat in a system. The proportion of latent heat in a basaltic magma can be large and represent the majority of available enthalpy (Morse, 2011). In the context of our experiments, this term is neglected because the latent heat release from our gelatin is relatively small, such that $S = 0.04$ (see the supporting information).

3. Results

3.1. Time-Dependent Behaviors and Flux-Dependent Behaviors

Each experiment began with an initial dike and the systems evolved as a function of the time interval between liquid injections. For experiments with a short time interval (seconds to minutes), the flow in the dike quickly channelized into a narrow conduit (Figure 2a), which tended to be elliptical in cross section, a few centimeters across and approximately half a centimeter thick. We were not able to detect any clear sign of melt back of the gelatin medium due to flow through the conduit, even for the highest flux; this is probably due to the low injection temperature we used. Experiments with a moderate time interval (30–60 min) had enough time for the dike to partially solidify. Such dikes were still soft at the start of the subsequent injection, which resulted in a new dike propagating inside the original dike (Figure 2b). We will refer to this as “nested dike” behavior. For experiments with a very long time interval (3+ hr), the dike had time to completely solidify. Subsequent injections created new dikes, leading to a dike swarm being generated (Figures 2c and 2d).

The flux seemed to have less control on the form of the plumbing system (i.e., conduit vs. dike swarm formation) and more control on the amount of material that remained solidified in the gelatin medium. Fluxes that were sufficiently high (>$10$ ml/min) allowed the dike to propagate to the surface and begin erupting (Figure 3a), which continued steadily until the injection finished. Very low flux experiments (~1 ml/min) could not propagate quickly enough to reach the surface before solidifying. In such cases, intrusion restarted at the base of the experiment and propagated alongside (in the same plane) the previous failed dike lobe (Figure 3c). This created a fingering appearance, with each lobe originating from the same source location. As the complex dike became wide and full of lobes, the liquid began to flow inside the older lobes, widening them, without significantly advancing the dike front (Figure 3d). We were also able to detect a transitional behavior at semilow flow rates (~3 ml/min), which formed a larger-scale version of the fingering behavior. Large dike lobes sometimes propagated all the way to the surface (other times no) and erupted in an unsteady fashion, flowing for tens of minutes before solidifying and forcing propagation to begin again at the base of the tank (Figure 3b). In such cases, it seems as though the viscous forces in the increasingly cold dike eventually overwhelmed the flux and propagation of a new dike became the easiest path for the liquid.
### Table 3

**Dike Dimension Measurements**

| Exp | H (cm) | L (cm) | $A_s$ (cm$^2$) | V (L) | Form          |
|-----|--------|--------|---------------|-------|--------------|
| 1   | 0.57 → 0.78 | 31.8   | 552 → 52     | 0.33 → 0.50 | Conduit      |
| 2   | 0.60 → 0.88 | 32.5   | 685 → 57     | 0.40 → 0.47 | Conduit      |
| 3   | 0.58 → 0.11 | 35.2   | 590 → 326    | 0.34 → 0.98 | Dike swarm   |
| 4   | 0.75 → 2.05 | 39.4   | 692 → 777    | 0.50 → 1.96 | No eruption  |
| 5   | 0.68 → 0.67 | 34.8   | 205 → 341    | 0.18 → 0.95 | Unstable     |
| 6   | 0.48 → 0.65 | 34.1   | 877 → 107    | 0.40 → 0.44 | Conduit      |
| 7   | 0.54 → 0.16 | 33.1   | 536 → 127    | 0.29 → 0.42 | Nested       |
| 8   | 0.13 → 0.10 | 35.5   | 241 → 261    | 0.10 → 0.13 | Nested       |
| 9   | 0.43 → 0.61 | 36.4   | 542 → 481    | 0.24 → 0.42 | Dike swarm   |
| 10  | 0.20 → 0.07 | 21.7   | 475 → 248    | 0.09 → 0.17 | Dike swarm   |

*Note.* From left to right are the median thickness, length, and surface area of the active region of the dike, the total intruded volume, as well as the form of the active intrusion. Each pair of values indicate the initial (left number; from the first injection) and final (right number; last injection) measurements. For dikes that erupted, these measurements are taken after the onset of eruption. For the dike that did not, these numbers represent the end of the first and the final injections. The vertical length does not change between intrusions in an experiment. The form references types of intrusions shown in Figures 2 and 3.

### 3.2. Image Processing

The dimensionless numbers $F_o$ and $\Phi$, which we use to analyze our results, are highly dependent on the geometry of the intrusion. To quantify this, we used image processing techniques to identify the thickness distribution and surface area of each intrusion, for each experiment. These techniques generally rely on comparing an image of interest with some reference image. Estimates for the dike geometry via the following methods are shown in Table 3.

#### 3.2.1. Dike Thickness Distribution

To estimate the dike’s opening thickness, we use the Beer-Lambert law, which relates the light transmission through a diffusing medium to its thickness (Taisne et al., 2011):

$$
H = -\ln(I/I_0)V/\sum[-\ln(I/I_0)A_{px}].
$$

The light transmitted through a dike, $I$, relative to a reference photo taken before the dike is intruded, $I_0$, decreases logarithmically as the thickness increases. For a pixel of area, $A_{px}$, summing the $-\ln(I/I_0)A_{px}$ term yields a value that is proportional to the actual volume. Scaling this value with the known volume, $V$, yields the width distribution of the dike. Via this method, we can estimate the thickness of a single dike or of a dike that channelized into a conduit.

For nested dike behavior, using the Beer-Lambert law can result in misleading thickness estimates, since it identifies all gelatin (both solidified and liquid) in the estimate. To determine the thickness distribution of the liquid region of each injection, we employ the Beer-Lambert law for two separate times, for one photograph taken during the injection and for one taken before the injection. Doing so yields a thickness distribution of all of the intruded gelatin, both liquid and solid, $H_{total}$, and one of only the solidified portion that remained in the medium, $H_{solid}$. The difference between the two reveals the thickness distribution of the liquid region, $H = H_{total} - H_{solid}$. Using these photo pairs, we could determine the geometry change between each subsequent intrusion, after the first (Figure 4).

For multiple layers of intrusions (i.e., dike swarm behavior), the Beer-Lambert law fails entirely, as the previous dikes pollute the image and complicate our estimation of the active dike’s thickness. Employing the Beer-Lambert law for multiple dikes tends to yield nonsensical results, which take the form of sudden thickness differences where dikes overlap. Since we can still estimate the surface area, $A_s$, of such dikes to a high

---

**Figure 4.** To estimate the thickness of a dike intruded inside of a previous dike, we use the Beer-Lambert law. We find the volume of intruded gelatin before and during an injection via the expression $-\ln(I/I_0)$ and the difference between the two yields the thickness distribution of the liquid region. We set a threshold value to automatically distinguish the active dike from the remaining gelatin, after which the surface area can be easily estimated by the number of pixels and the area of a pixel.
degree of confidence and we know the volume injected, we simply estimate the average thickness via \( H = V/A_y \).

### 3.2.2. Surface Area of the Active Region of an Intrusion

As each experiment progresses, the intrusion evolves to become more complex. We are primarily interested in the active region of the intrusion, whether it takes the form of a dike or conduit. For dike swarm experiments and nested dike experiments, the active region is the newest dike (or embedded dike), so we could simply compare a photo from one injection, \( I_y \), to the previous, \( I_{y-1} \), to highlight where the newest dike intruded. Viewing the ratio \( I_y/I_{y-1} \) in Matlab causes the new dike to appear very apparent and its surface area and thickness can be estimated, \( A_y = A_{pxN} \), for a dike identified in \( N \) pixels. In practice, we did this simultaneously with the thickness distributions, by defining a threshold degree of change. The pixels with significant change indicated the active surface area of the dike, while those that were lower than the threshold indicated the inactive regions (Figure 4).

For conduit-forming experiments, comparing a photo from one injection with the previous tends to reveal little change, since the conduit maintains a stable geometry. We manually masked the photos of conduit-forming experiments to highlight the conduit shape and, by processing the masked image in Matlab, identify its surface area and thickness.

### 3.3. Framing the Results Through Dimensionless Numbers

To compare our results with natural systems, we employ the dimensionless time parameter, \( Fo \), and the dimensionless flux parameter, \( \Phi \). The dike thickness distributions determined via image analysis provide us with a range of values, so we take the median thickness. We use the horizontal aspect ratio, \( \Psi \), (median opening thickness to the median horizontal breadth) of the actively flowing region of the intrusion to quantitatively characterize what kind of feature it is:

\[
\Psi = H/B.
\]

Cylindrical conduits therefore have low aspect ratios close to 1, while dikes have aspect ratios that approach 0. Note that the \( \Psi \) shares \( H/B \) terms with \( \Phi \) and, in order to eliminate like terms, we plot a form of the dimensionless flux, such that \( \Phi/\Psi = Q/(\alpha L) \). By plotting \( \Psi \) against \( Fo \) and \( Q/(\alpha L) \) (Figure 5), we can see to what degree the intrusion shape depends on either parameter. To do so, we plot an empirical, best fitting surface of the form of \( \Psi = C_1Fo^{-2}[Q/(\alpha L)]^{C_2} \) and solve for the values of \( C_1, C_2 \), and \( C_3 \) by minimizing the sum of the square of the residuals (in log scale) between measurements and the model surface of \( \Psi \). The equation of the surface (shown in Figure 5a) indicates that \( \Psi \) is dependent on both parameters, in that conduits need both a moderate flux and frequent intrusion to remain active. Conduits tend to develop for \( Fo < 1 \), nested, compound dikes form when \( 1 < Fo < 10 \) and dike swarms occur for \( Fo > 10 \). The \( Q/(\alpha L) \) parameter determines the thermal stability or propagation, in which solidification causes dikes arrest when \( Q/(\alpha L) < 1 \) and eruption can proceed unimpeded when \( Q/(\alpha L) > \sim 10 \). In terms of \( \Phi \), these two end members are respectively quantified by \( \Phi < 10^{-1} \) and \( \Phi > 3 \times 10^{-1} \) (though for simplicity we will later use \( \Phi > 1 \)).

We also explore the effect of these two parameters on the amount of material intruded into the crust. For an injection volume, \( V_{inj} \), of 500 ml, we measured the volume of material that erupted, \( V_{erupt} \) and simply analyze the intruded volume ratio, \( V^* \), such that

\[
V^* = (V_{inj} - V_{erupt})/V_{inj}.
\]

Plotting \( V^* \) against \( Fo \) and \( \Phi \) and applying another best fitting surface as done for Figure 5, we see that the intruded volume strongly depends on the flux (Figure 6). This makes sense, in that a low flux allows more time for the intruded liquid to cool and solidify inside the medium. The weak dependence on \( Fo \) stems from the plumbing evolution of each type of behavior. For example, the value of \( Fo \) did not change much throughout conduit-forming experiments, even though they progressively extruded more and more liquid as the conduit developed into a stable form. For nested dike experiments by contrast, each intrusion was thinner than the previous, causing the corresponding values of \( Fo \) to increase and \( V^* \) to decrease (they were thin, so less volume intruded). Similarly, for dike swarms, the first dikes to intrude could freely deform the surrounding
Figure 5. The aspect ratio, $\Psi$, against dimensionless time, $Fo$, and dimensionless flux divided by the aspect ratio, such that $\Phi/\Psi = Q/(\alpha L)$. The $\Phi$ contains the same $H/B$ terms as the aspect ratio, so we use $Q/(\alpha L)$ to show the independent effect of flux. Colors correspond to experiments that were conduit forming (red), unstable flow (magenta), nested dikes (green), dike swarms (blue), and finally dikes that did not erupt (blue x). We do not use noneruptive dikes to analyze the aspect ratio, since they do not match eruptive data and lay visibly below the trend in (b). (a) The parameters are plotted in 3-D. A best fitting surface is applied to the data to see the relative contribution of $Fo$ and $Q/(\alpha L)$. (b–d) Two-dimensional plots showing how the parameters compare to each other.

Figure 6. How dimensionless time, $Fo$, and dimensionless flux, $\Phi$, affect the ratio of intruded volume, $V^*$. Markings are as in Figure 5. (a) The parameters are plotted in 3-D, and a best fitting surface is shown. The intruded volume is almost entirely a function of $\Phi$. (b–d) Two-dimensional plots showing how the parameters compare to each other. $Fo$ has little influence on $V^*$, so there is no apparent trend in (b). By comparison in (c), low values of $\Phi$ correspond to $V^*$ values approaching 1.
medium, resulting in greater thicknesses, lower values of Fo and greater values of $V^\ast$. Subsequent dikes interacted with the local stresses due to the previous ones and therefore were relatively thin.

Regarding the best fitting surfaces shown in Figures 5a and 6a, we plot these to determine the degree to which Fo and $\Theta$, or alternately $Q/(\alpha L)$, have an effect on $\Psi$ and $V^\ast$; however, for various reasons we hesitate to use them as predictive tools and therefore do not further discuss them in this article. For example, the relationship between Fo, $\Theta$, and $V^\ast$ is not linear. As we noted, the amount of intruded material decreased as each experiment progressed, so in Figure 6b, each category of experiment has a negative trend between Fo and $V^\ast$. However, between categories, there is a positive trend; conduit experiment intruded less material than nested dikes, which intruded less than dike swarms. To forecast the value of $V^\ast$, one would require a complete knowledge of the stress and temperature distribution in the medium, as well as the dynamics of the intrusion. Similarly for $\Psi$, we cannot make any predictions as functions of Fo and $Q/(\alpha L)$, since the aspect ratio in gelatin experiments is different from natural magmatic dikes; values are $10^{-2}$ to $10^{-1}$ for gelatin and $10^{-4}$ to $10^{-3}$ for in nature (Kavanagh et al., 2013).

What we can do is quantitatively distinguish between types of behavior. For example, conduit-forming experiments had Fo $< 1$, so we can broadly say that liquid migrates via a conduit geometry when the time interval is short. Dike swarms by contrast had Fo $> 100$. Similarly, experiments that were able to make stable eruptions had $Q/(\alpha L) > ~10$, whereas those that arrested due to solidification had $Q/(\alpha L) < ~1$.

4. Discussion

4.1. Channelization Due to Solidification or Due to Preexisting Flow Paths

It is important to consider that dikes have complex internal flow, which may affect how heat is distributed throughout the dike. Kavanagh et al. (2018) identified several stages of flow within a dike including: (1) initial injection and radial growth of a penny-shaped crack, (2) growth and internal circulation of the magma, and (3) eruption and ejection of magma from the dike. The assumption that magma flows unidirectionally away from the source may not be true during the propagation stage of a dike and instead may only apply for syneruptive, channelized flow. In this context, it is possible that, during dike propagation, a central upward jet provides heat to the center of a dike, while cooler magma circulates downward along the lateral edges, imposing a thermal distribution that favors channelization.

One way to check for this would be by measuring a time series of a fissure eruption’s temperature distribution. Kavanagh et al. (2018) show that, prior to eruption, dikes can have an internal magma circulation, consisting of an upward central jet and lateral downward flow. As the dike breaches the surface, it contracts and squeezes much of the liquid upward onto the surface. Assuming that the preeruptive circulation imposes a thermal distribution that is warmer toward the center and cooler toward the sides, the distribution would be reflected in the initial stage of a fissure eruption (Figure 7). The fissure would initially emit magma from the upward central jet, followed by the remaining somewhat cooler magma. As the eruption continues and channelizes, the temperature would increase to reflect the magma’s direct path from depth to the surface. To the best of our knowledge, an initial drop in temperature following the onset of a fissure eruption has not yet been identified, however, the progressive increase in temperature as a fissure channelizes has been shown at the Holuhraun lava field associated with the 2014 eruption of Bardarbunga volcano (Aufaristama et al., 2018).

4.2. Plumbing Region Distribution

Perhaps, the most important discussion from these results is that of how a plumbing system initiates and is distributed. A key finding from our results is the effect of injection frequency on the form of migration. When there is a long repose time between magma pulses, migration occurs via dike propagation, since heat can entirely conduct away from any intrusion. In such a case, there is no preferred path (thermally speaking) along which a dike will propagate. Sporadic intrusions of magma would be randomly distributed throughout the region, favoring the small, decentralized eruptions associated to volcanic fields (Figure 8a).

However, a repose time may be long enough that migration occurs via dike propagation but short enough that the region gradually builds up heat. Subsequent dike propagation is favored in the relatively warm (decreasing the value of $\Theta$) and relatively soft regions of the crust and it is likely that a dike can migrate laterally along a horizontal temperature gradient, further focusing magmatism into a central region. The long-
term result of such magma influx frequency would be the development of a centralized volcanic plumbing system, along with an associated large, central edifice (Figure 8b). Since dikes in this scenario would be focused to varying degrees (i.e., dikes do not horizontally migrate all the way to the center of the region), the edifice would likely build up into an oblong structure. If the repose time of magma influx is very short, we imagine the system would be more centralized to a narrow region (Figure 8c) and construct a fairly conical edifice.

Figure 8. A model of how volcanic activity is horizontally distributed, depending on the frequency of magma influx defined by the dimensionless time, Fo. The boundary between (a) and (b) is numerically derived (see below) and the boundary between (b) and (c) is experimentally derived (see section 3). (a) A very low frequency allows all heat from an intrusion to be lost. Dikes propagate through the crust, independently to each other, creating small monogenetic cones. (b) A moderate frequency maintains a hot central region but no molten path. Dikes preferably propagate through the hotter region and build a less centralized plumbing system and a large edifice. (c) Very high frequency maintains a molten pathway along which magma migrates. The resulting plumbing system would be very centralized and would build a large edifice.
With this logic in mind, we attempt to define a critical frequency of influx, which separates these different types of systems. To do so, we use a simple 2-D numerical model of transient heat conduction through the crust, in which heat transfer between elements in the model is described by Fourier’s law:

$$\rho \Delta x^3 C_p \frac{dT}{dt} = k \Delta x^2 \frac{dT}{dx},$$  \hspace{1cm} (9)

for density $\rho$, specific heat capacity $C_p$, thermal conductivity $k$, element volume $\Delta x^3$, and surface area $\Delta x^2$ (Çengel, 2008; Turcotte & Schubert, 2002). This model is essentially a simplified version of the model presented by Daniels et al. (2014) and aims to quantify the time scale over which heat is competely lost from a dike. It assumes that a vertical column of 1-m half-width, representing a dike, becomes hot at time, $t = 0$, with a temperature of 1400 K and conducts heat away with time over a horizontal spatial scale of 1,000 m and depth of 8,000 m. The boundary is then reheated on a recurrence frequency of $10^7$ s, $10^8$ s, etc., and the temperature distribution is allowed to evolve for $10^{12}$ s (~30 kyr). The model imposes a surface temperature of 300 K and a geothermal gradient of 30 °C/km. For more details on the model setup, please see the supporting information.

Note that we do not make any assumption of sills, plutons, or other magmatic bodies (of which there could be various shapes, sizes, numbers, or locations) that would naturally develop in such plumbing systems and instead consider only vertical intrusions. Such intrusions would undoubtedly change the heat distribution in the crust but develop due to other, nonthermal processes such as fracturing and deformation of the crust. At any rate, such bodies develop with time and would not exist when a plumbing system first initiates.

The conduction model indicates that repetitive intrusions can build up heat in a narrow band of the crust. For a recurrence interval on the time scale less than $10^{10}$ s (~300 years), the crust stays warm near the intrusion, whereas for $t_{\text{int}} \geq 10^{12}$ s, all heat is lost and the temperature profile returns to the geothermal gradient.

Figure 9. Two-dimensional temperature contours for the crust near an intrusive region, from the numerical model. These model the heat distribution after a fixed time frame of $10^{12}$ s (~30 kyr), which is similar to the repose time of volcanic fields. The model allows an intrusion to reheat the left side of the region on a recurrence time interval, $t_{\text{int}}$, shown in each subfigure and the system evolves via conduction. The surface is kept at a constant temperature of 300 K, and we impose a geothermal gradient of 30 °C/km. For $t_{\text{int}} \leq 10^{10}$ s, the region stays warm near the intrusion, whereas for $t_{\text{int}} \geq 10^{12}$ s, all heat is lost and the temperature profile returns to the geothermal gradient.

Our model informs us of the intrusion recurrence frequency that is necessary for the focusing of a plumbing system, assuming that heating of the crust indeed promotes dike propagation into a narrow region (i.e., Figure 8b). Such focusing can occur when the intrusion recurrence interval is less than 1 kyr, whereas a less frequent recurrence allows intrusions to be spatially dispersed. This agrees with Valentine and Connor (2015), who indicate that volcanic fields typically have eruption recurrence intervals 10–100 kyr but may be as short as 1 kyr. Plumbing systems with shorter recurrence intervals are perhaps characterized by
centralization of intrusive events into a focused plumbing system, which in turn constructs a large, central-
ized edifice. As such edifices are constructed, their loading on the crust further focuses intrusions toward
the edifice (Roman & Jaupart, 2014). However, it should be reiterated that the model is simplistic, so it does
not account for complicating factors, like a hydrothermal system, which would greatly accelerate heat trans-
port at shallow depth, distributing the heat from a dike over a relatively wide area in a relatively short time
(Cherkaoui et al., 2003; Chevallier & Verwoerd, 1990; Pashkevich & Taskin, 2009).

Note that temperature profiles estimated by our model are fairly straightforward and show agreement with
other studies. For example, Delaney (1988) developed a numerical model for 1-D conduction from a dike
with a temperature-dependent thermal properties and found that it rapidly cools from magmatic tempera-
ture to ambient temperature on a time scale of ~10^7 s (~4 months); in our model a single dike cools off on
the same time scale. In terms of 2-D modeling, our model generates temperature gradients similar to those
modeled by Rudman and Epp (1983), on the same time scales and spatial scales, though with a different
shape (they model a relatively wide dike, which narrows to a point). Our model results also resemble the
heat accumulation modeled by Daniels et al. (2014), who investigated the heating due to repeated dike intru-
sions during continental rifting. They found that a recurrence interval of 3 × 10^10 s can significantly heat the
nearest ~500 m to the dike on a time frame of 20 kyr, which nearly matches the conditions and results for the
model in Figure 9a.

4.3. Stable Conduit Development at Depth

Our experiments provide insight into the conditions for which a stable, molten conduit may develop in the
crust (Fo < 1 and Φ > 1). Considering that the dimensionless numbers control the plumbing geometry, we
can rewrite equations (1) and (2) to solve for the necessary conditions to maintain a molten channel. We
label these as the critical flux, \( Q_{\text{crit}} \), and critical repose time, \( t_{\text{crit}} \). Assuming steady-state heat transfer in a
conduit geometry with radius, \( r \), the heat into the conduit due to magma flux balances with loss due to conduction:

\[
Q_{\text{crit}} \rho C_p T_0 = 2\pi L k (T_0 - T_\infty) / \ln(r/R),
\]

for a magma and ambient temperatures \( T_0 \) and \( T_\infty \) and conduction occurring over a radial distance into the
crust, \( r \) (Çengel, 2008). The \( Q_{\text{crit}} \) is dependent on temperature and the temperature gradient. For values of 2
< \( r/R < 1,000 \), the there is a constricted range of values of 0.7 < \( \ln(r/R) < 7 \). Since it varies over a small range,
we assume it to be constant, so

\[
Q_{\text{crit}} = 2\pi \alpha L (1 - T_\infty / T_0).
\]

For magma much hotter than the ambient temperature (so that \( T_\infty / T_0 \approx 0 \)), which ascends through a 1- to
10-km-long conduit, a relatively low mean flux of only \( 10^{-2} - 10^{-1} \) m^3/s is required to maintain a liquid conduit.

Perhaps unsurprisingly, the time parameter has stricter conditions for a conduit to remain molten. The time
for solidification implies a transient heat conduction problem, such that

\[
V p C_p (\partial T / \partial t) = k A_s (\partial T / \partial r),
\]

assuming the flow has stopped, in a conduit of surface area, \( A_s \), and volume, \( V \). Therefore,

\[
\partial T / \partial t = (2\alpha / R) (\partial T / \partial r),
\]

in which \( \partial T / \partial t \) and \( \partial T / \partial r \) decrease as time goes by and the conduit cools. For solidification to occur, the
magma only needs to cool to the solidus temperature, and we integrate the cooling rate over time, \( t_{\text{crit}} \), so
that

\[
(\partial T / \partial t)_{\text{crit}} = (T_0 - T_s) / t_{\text{crit}}
\]

for a solidus temperature, \( T_s \). This approximates the temperature gradient to a time-averaged gradient. If we integrate again with respect to \( r \), over a range of \( R \) to \( r \), then

\[
(T_0 - T_s) / t_{\text{crit}} = 2\alpha (T_0 - T_\infty) / [R (R - R)].
\]

If \( r \) is proportional to \( R \), such that \( R^* = r / R \), then
In the above equation, characteristic time scale represented by \( R^2/2\alpha \) has a strong dependence on the conduit radius, in that large conduits have a large ratio of volume to surface area. For a 1-m-radius conduit, the time scale is several days, whereas for a 10-m-radius conduit, it is around 19 months. However, the other terms in the equation modify the time estimate to a significant degree. For an initial temperature defined by \( \Theta \) (Table 2), the temperature terms in this equation together have a value of 0.28 to 0.38, suggesting quicker cooling. The temperature gradient term, \( R^* - 1 \), meanwhile suggests longer cooling times. It is difficult to judge which radial distance approximates the temperature gradient, but if conduction occurs over tens or hundreds of meters, this term could be large (Figure 10).

In terms of sensible heat, reasonable estimates for conduit cooling would be weeks to decades, depending on the magma temperature, crust temperature and conduit geometry. Latent heat however can provide a large amount of thermal energy and represent the majority of available enthalpy (Morse, 2011). In this context, as magma crystallizes, it likely enters an approximately steady state of heat loss, in which conduction balances with the energy released during phase changes. If the available amount of latent heat is indeed several times larger than the sensible heat, then solidification would take a similar factor longer to conclude. In the context of our experiments, it is expected that the latent heat does not play a large role, since most of the available heat is sensible heat (see the supporting information).

The conditions for solidification in a conduit geometry have also been addressed in previous studies. Solidification occurs by the freezing of magma along the annulus of conduit, which proceeds when the heat loss due to conduction exceeds heat supply due to advection, for laminar flow conditions with a constant pressure drop; when heat loss balances with supply, the conduit maintains a steady-state radius (Mulligan & Jones, 1976; Zerkle & Sunderland, 1968). Holmes-Cerfon and Whitehead (2011) show that the critical flux necessary for steady-state flow depends on the pressure drop, the Stefan number (equation (5)), the dimensionless flux \( \Phi \) (equation (1), modified for a conduit geometry, in which \( H/B = 1 \)) and the dimensionless temperature difference, \( \Theta_d \), similar to \( \Theta \) described above, in which \( \Theta_d = (T_s - T_\infty)/(T_0 - T_s) \). The \( \Theta_d \) corresponds to the degree of solidification, in which \( \Theta_d \gg 1 \) indicates that solidification should occur fairly rapidly. For magma driven by a constant flux (which is the driving condition for our experiments, in which pressure is variable), they find that any flux can maintain magma flow, but may cause large driving pressures. However, for magma flow driven by a pressure drop (either constant or somewhat variable), they find that stable flow in a conduit requires a minimum critical flux of \( \Phi \sim 6,000 \), significantly higher that that observed in our experiments (\( \Phi \sim 1 \)). This could be partially explained by the time-dependent solidification of gelatin in our experiments, which is discussed later in section 4.5 and allows gelatin to temporarily (~10 min) remain liquid as it cools below the solidus. However, we expect this yields a similar effect as
latent heat release, which also delays the solidification process. Although there is not a significant latent heat release from our gelatin, the time-dependent solidification effectively takes its place to a degree.

A natural setting that indicates the necessary critical flux for a conduit is a lava lake, which represents the uppermost surface of a molten conduit. Lava lakes are thought to persist due to magma convection, in which bidirectional flow supplies heat from depth throughout the conduit and maintains a molten pathway (Harris et al., 1999; Oppenheimer et al., 2004; Sweeney et al., 2008). Depending on their level of activity, the magma flux can vary in magnitude, but tends to be greater than 0.01 m³/s. For example, Harris et al. (1999) estimated that Mt. Erebus lava lake in Antarctica had a magma flux of ~0.01 m³/s during measurements taken in both 1986 and 1989; similarly Erta’ Ale lava lake in Ethiopia had fluxes varying from 5 m³/s in 1973 to 0.01 m³/s in 1986. They reported that other lava lakes tended to have higher fluxes. Although such systems are characterized by bidirectional flow, rather than the unidirectional flow that we analyze in this study, we can assume this 0.01-m³/s base value provides an order-of-magnitude estimate of the necessary critical flux for stable flow. Assuming that such conduits are between 1 and 10 km long and that α = 10⁻⁵ m²/s, then the minimum value of Φ is 1–10, similar to what we estimate for our experiments (solving equation (1), in which H/B = 1).

If we consider deeper magma migration, in the form of magma supply from deep sources into shallow crustal reservoirs, these estimates hint that the form of magma migration ultimately depends on how magma enters the crust from the mantle. A slow, steady trickle of fresh magma may be sufficient to maintain a molten conduit, thereby continuously feeding shallower bodies. Conversely, if magma accumulates at the bottom of the crust in infrequent, discrete pulses, a previously existing conduit would not have the necessary heat to remain liquid, so the magma would have to migrate in the form of a new dike. In this sense, magma supply into shallow reservoirs may depend on how effectively and how steadily partial melt can form and percolate into the crust.

A good example of the effect of supply frequency and cooling on deep magma migration may be Piton de la Fournaise. Prior to the 1998 eruption, magma migration from depth produces a series of volcano tectonic earthquakes, whose hypocenters tracked the migration front (Battaglia et al., 2005). The lateral distribution of these earthquakes indicated that the migration followed a fairly narrow path below sea level and then a sheet-like path above sea level. This sudden change in geometry was considered to be due to the magma flowing through a preexisting conduit (perhaps partially molten) at depth and then rapidly propagating a shallow dike. The deep conduit structure was reactivated during the 2015 seismic episode, which also led to an eruption (Lengliné et al., 2016). The small magnitudes of the deep seismic events are evidence that magma flowed through the existing, partially molten structure and therefore implies that it was persistently active on the 1998–2015 time frame. We argue that there may be a thermal discontinuity located at sea level, in which the underlying crust is well insulated and maintains a open, persistently active conduit. Above this discontinuity, the crust may be generally cooler, so that intrusions are thoroughly solidified and require migration in the form of a dike.

4.4. Shield Volcano Rift Zones

In addition to identifying the nature of supply pathways, we can apply our findings to rift zones (namely those that occur on the flanks of shield volcanoes; for example, Kilauea’s east rift zone, Piton de la Fournaise rift zones) to gain some sense of the internal structure. Such rift zones are known to be frequently active pathways along which magma repeatedly travels.

In the case of Kilauea, magma frequently drains from the summit region along the rift zone and this background activity is occasionally punctuated by more intense fissure eruptions. The frequent, relatively low flux of the background phase likely favors the form of a subhorizontal conduit, which minimizes the heat loss due to conduction. The higher-intensity fissure eruptions occur due to some combination of increased magma flux (internal pressure) and subsidence of the volcano’s flank (external extension), which cause the conduit to fail and propagate a dike.

Over the lifetime of the rift zone, it is plausible that the geometry would oscillate between cylindrical and planar, depending on the level of activity, slowly accumulating layers into a dike complex (Figure 11). Indeed, geophysical measurements (seismic reflections, gravity, and magnetic data) indicate that a dense intrusive body spans the length of Kilauea’s east rift zone to the submarine Puna Ridge and contains a narrower, hot (~1000 °C) region along its axis (Furumoto, 1978; Leslie et al., 2004). Subsidence of the volcano’s
flank generates extensional forces that cause dikes to orient perpendicularly to the direction of subsidence, ensuring the intrusive dike complex is subvertical, with dimensions much taller and longer than wide.

4.5. Experimental Limitations

Our experimental setup has the inherent limitation that gelatin solidifies as a function of temperature, time and shear strain. We consistently measured eruptive temperatures below the solidus of gelatin, indicating that solidification either requires time or is inhibited due to the liquid's motion. In either case, the resulting dike's form could be somewhat different from a magmatic dike (perhaps thinner, with less material intruded). A material that solidifies independently of time and shear strain may have grown a thicker chilled margin and restricted the internal flow. Additionally, the time dependence of gelatin solidification implies that its viscosity may also have some time dependence. This potentially complicates comparison with other studies of flow channelization in a fissure, in which it has been shown that channelization strongly depends on the viscosity ratio between warmer and cooler regions of a dike.

That said, we believe our results are fundamentally sound. Taisne and Tait (2011) made analogue experiments of solidifying dikes using paraffin wax, which is strongly temperature-dependent. The general form of their dikes were similar to ours, albeit with greater thickness and distinct chilled margins. It is likely that the inner fluid core of the dike undergoes flow patterns similar to those observed in this study and those described by Kavanagh et al. (2018). Furthermore, we were able to replicate behaviors observed in nature, such as the heterogeneous solidification and the subsequent channelization of a dike.

Quantitatively speaking, the characteristic time for solidification, $H^2/\alpha$, for our experimental dikes is on the time scale of ~10 min, which matches the time frame for channelization that we observed in our experiments, which seems to imply the time dependence did not play a large role in the solidification of a thin dike. For experiments that did not result in conduit formation, either the time interval between injections or the flux duration (or both) tended to be much larger (>30 min) than this 10-min time frame. All things considered, it seems as though the gelatin's time dependence had no effect on the form of liquid migration. Rather, its effects were confined to the aspect ratio of the dikes, which may have been somewhat thinner than for a material with a time-independent phase transition.
5. Conclusions

The experiments we present in this paper illustrate the effects of solidification on magma migration. For very frequent intrusions ($Fo < 10^0$), magma maintains a hot, open pathway from depth to the surface. This path initiates as a dike and, upon eruption, quickly channelizes into a conduit. For very infrequent intrusions ($Fo > 10^5$), the dike can completely solidify, so a subsequent intrusion forms a new dike and the system eventually accumulates a dike swarm. We also show a transitional behavior ($Fo \sim 10^3$), in which a dike intrusion solidifies but remains soft, so subsequent intrusion propagate new dikes inside the original dike.

We find that the effect of flux primarily controls the stability of flow through the crust. Intrusions that propagate with low fluxes ($\Phi < 10^{-1}$) struggle to reach the surface and form thick dikes that are composed of many small, individual lobes that are compounded together. Dikes with sufficiently high flux ($\Phi > 1$) propagate to the surface and can steadily flow until the supply stops. Between these two regimes, the flux has a transitional regime, in which a dike may (or may not) propagate to the surface and any conduit that forms is temporary, and eventually closes due to solidification.

We expand these results using a 2-D numerical model of thermal conduction. For injections more frequent than $10^{11}$ to $10^{13}$ s (about 3 to 30 kyr; $Fo = 10^{-4}$ to $10^{-5}$), the crust accumulates heat near the intrusive region, which we argue softens the crust and encourages subsequent dikes propagate nearby. For less frequent intrusions, heat is lost and the crust returns to its background temperature. Considering the experimental and numerical evidence together, the frequency of migration dictates whether magma flows through a conduit or a series of spatially focused or spatially dispersed dikes.

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