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Silicate Volcanism on Europa's Seafloor and Implications for Habitability

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Abstract Habitable ocean environments on Europa require an influx of reactants to maintain chemical disequilibrium. One possible source of reactants is seafloor volcanism. Modeling has shown that dissipation of tidal energy in Europa's asthenosphere can generate melt, but melt formation cannot be equated with volcanism. Melt must also be transported through Europa's cold lithosphere to erupt at the seafloor. Here, we use two models of dike propagation to show that dikes can only traverse the lithosphere if either the fracture toughness of the lithosphere or the flux into the dike is large (>500 MPa m$^{1/2}$ or ~1 m$^3$ s$^{-1}$, respectively). We conclude that cyclic volcanic episodes might provide reactants to Europa's ocean if magma accumulates at the base of the lithosphere for several thousand years. However, if dikes form too frequently, or are too numerous, the magma flux into each will be insufficient, and volcanism cannot support a habitable ocean environment.

Plain Language Summary Beneath the icy crust of Jupiter's moon Europa lies a deep, liquid-water ocean that might provide an environment suitable for life. In the absence of sunlight, organisms must extract energy from chemical reactions within the ocean. Over time, organisms would use up all the chemical ingredients needed for these reactions, leading to extinction unless new ingredients were added to the environment. Because the ocean is closed-off from Europa's surface by the thick ice crust, seafloor volcanism has been proposed as a source of chemical ingredients. Europa's deep rock interior can get hot enough to form small amounts of magma that will rise upward; however, the magma must also travel across the outer cold portion of the rock interior, which requires the formation and growth of magmatic dikes. Here we determine that dikes long enough to supply magma to Europa's seafloor only form if the rock is either very strong, or the amount of magma entering the dike is very large. In those cases, volcanism could occur on Europa's seafloor every few thousand years, supplying the necessary life-sustaining chemical ingredients. NASA's upcoming Europa Clipper mission will provide new data that helps clarify whether the conditions needed for seafloor volcanism are met.

1. The Relevance of Volcanism to Europa's Habitability

Other than the Earth, Jupiter's moon Europa is one of the most likely bodies in our Solar System to harbor an environment that is currently habitable for life as we understand it. There, liquid water is in direct contact with the silicate layer below, providing both the solvent and elemental building blocks necessary for life. Biologically useful sources of energy may be the limiting condition (Chyba & Hand, 2001; Gaidos et al., 1999). Life as we know it must extract energy from its environment to perform cell maintenance, metabolic activity, and reproduction. In the absence of sunlight, life in Europa's ocean must extract energy from chemical disequilibrium within the environment (i.e., from the available Gibbs free energy (see Hand et al., 2009 for a review)). Both oxidation and reduction reactions are possible (e.g., Hand et al., 2009; Zolotov & Shock, 2004), and the identification of organisms utilizing such biochemical pathways in challenging terrestrial environments such as Lake Vostok (e.g., Christner et al., 2006) lends credence to the plausibility that such reactions could sustain life on Europa's seafloor. Less certain, however, is whether geochemical cycles can maintain chemical disequilibrium in Europa's ocean. That is, without at least a periodic influx of reactants, Europa's ocean could reach chemical equilibrium, leading to what has been termed “thermodynamics-driven extinction” (Gaidos et al., 1999).

Two potential pathways exist for providing reactants to Europa's ocean. In the first, reactants are introduced to the ocean from above by surface-ocean exchange. In a thin (~1 km) ice shell, radiolytically produced oxidants on the surface (e.g., $O_2$, $H_2O_2$) can easily be transferred to the ocean on short timescales (Greenberg, 2010; Greenberg et al., 2000). However, in the more-likely case of a thick (~10s of km) ice shell (e.g., Howell, 2021;
Singer et al., 2021; Tobie et al., 2003; Turtle & Pierazzo, 2001) exchange between the surface and ocean is far from assured. Hand et al. (2009) assumed that the delivery timescale for surface material to the ocean is similar to Europa’s surface age of ~100 Myrs (Bierhaus et al., 2009). However, the mechanism of resurfacing is uncertain and may not involve substantial surface-ocean exchange. For example, chaos formation may involve disruption and resurfacing by warm ice diapirs (Pappalardo et al., 1998) rather than actual interchange with the ocean below. Widespread subsumption of the icy lithosphere may inject surface material directly into the ocean (Kattenhorn & Prockter, 2014), but the subsumption process is mechanically difficult and poorly understood (Bland & McKinnon, 2019; Howell & Pappalardo, 2019; Johnson et al., 2017).

The second pathway is for reactants to be introduced into the ocean from below by volcanism and hydrothermal activity (McCollom, 1999) or by serpentinization (S. D. Vance et al., 2016); however, until recently, the evolution of Europa’s silicate interior has often been neglected entirely or highly simplified (e.g., Mitri & Showman, 2005; Tobie et al., 2003). Tidal dissipation studies (Greeley et al., 2004) and simple attempts to scale Io’s heat flux to Europa (Geissler et al., 2001) suggest that partial melting and volcanism are possibilities, although such scalings have considerable uncertainty. Moore and Hussmann (2009) provide a much more robust investigation of the thermal state of Europa’s silicate interior by applying a model originally derived for Io (Moore, 2001, 2003) that includes the potential for the formation of partial melt and its effect on convection and tidal dissipation. This model identified two stable equilibrium states: a cold state in which no partial melting occurs, and a hot state with temperatures above the solidus. The latter state implies that active volcanism could indeed occur. Models of the coupled orbital-thermal evolution of Io and Europa find that Europa can, in fact, persist in the “hot state” (Hussmann & Breuer, 2007; Hussmann & Spohn, 2004; Moore & Hussmann, 2009).

More recently, Běhoušová et al. (2021) have argued, based on three-dimensional numerical modeling, that melting of Europa’s silicate interior can occur throughout most of Europa’s history, with melt volumes similar to those of large igneous provinces on Earth. However, the model assumes instantaneous melt extraction and thus neglects how the presence of melt would decrease the viscosity of the asthenosphere, increase convective vigor and the associated cooling rate, and thus decrease the rate of melt formation. Elder and Bland (2018) applied the one-dimensional melt formation and migration model of Elder (2015), which explicitly includes the coupling between melt fraction and viscosity, to Europa’s silicate interior and found that melting does occur, but only when substantial tidal energy is dissipated in the silicates (e.g., the hot state of Moore and Hussmann (2009)).

In both the Běhoušová et al. (2021) and Elder and Bland (2018) models, melt is assumed to be extracted from the system by volcanic eruption. Melt formation in the asthenosphere (i.e., the warm, ductile region where tidal dissipation can occur) cannot, however, be equated with silicate volcanism, or even hydrothermal activity, on Europa’s seafloor. For seafloor volcanism to occur magma must be transported via dikes across the entire thickness of the cold silicate lithosphere, a distance >100 km (e.g., the conductive lid is 300–400 km thick in the models of Běhoušová et al. (2021)). Terrestrial studies have shown that a substantial fraction of the Earth’s magma inventory is intruded in the lithosphere rather than extruded on the surface as lava. Typical intrusive/exusive ratios are estimated to be between 5:1 and 10:1 (e.g., Crisp, 1984), although the range of inferred ratios is large (White et al., 2006). Furthermore, the process of dike formation is complex in terrestrial settings, and models of dike propagation have never been applied to magma transport through Europa’s lithosphere. Melt escaping the asthenosphere is a necessary but insufficient condition for sustaining a habitable seafloor environment on Europa by magmatism and/or hydrothermal activity. The driving question for this work, then, is to determine the conditions under which the melt generated within Europa’s asthenosphere can fully penetrate the lithosphere and erupt on the seafloor.

### 2. Models of Dike Formation

Given the existence of melt in the asthenosphere, the plausibility of seafloor volcanism occurring depends, to first order, on whether dikes can transport magma through the lithosphere faster than the dike freezing timescale. The freezing timescale depends most sensitively on the width of the dike. Thus, by determining likely dike widths, propagation velocities, and cooling timescales, we can begin to assess whether seafloor volcanism is possible on Europa.

The mechanism by which a fluid-filled crack can propagate through an elastic medium has been described by at least two different high-heritage endmember conceptual models (i.e., a Weertman style model and a fluid dynamic
model) with different implications for dike geometry and propagation (Rivalet al., 2015). Both endmembers are simple in that they assume a single, isolated dike in an infinite half space with constant material properties. In reality, multiple dikes may form and interact (e.g., Kuhn & Dahm, 2008), and the existence of the free surface will modify fracture dynamics (e.g., Kuhn & Dahm, 2008; Meriaux & Jaupart, 1998; Pollard & Holzhausen, 1979; Rivalet al., 2006). Magma may not be positively buoyant throughout the lithosphere (Lister, 1991; Wilson & Head, 1992, 2002), and it may accumulate to form magma chambers that subsequently become over-pressure and erupt more evolved magmas (e.g., Parfitt et al., 1993; Wilson & Head, 2002). Magma may also move laterally due to reduced buoyancy or in response to stress gradients (Dahm, 2000a; Menand et al., 2010) or layering (Kavanagh et al., 2006). The elastic properties of the lithosphere may change with depth or the medium may be layered (e.g., Maccaferri et al., 2010; Taisne & Jaupart, 2009). Positive feedback between dike width and magma flow can lengthen or shorten dike lifetime (e.g., Bruce & Huppert, 1989; Carrigan et al., 1992; Fialko & Rubin, 1999).

In short, all of the complications of magma transport common to terrestrial volcanology are likely to occur on Europa if magma is generated in the asthenosphere. Of particular importance in eruption dynamics on terrestrial planets is the exsolution of the gas phase as the dike approaches the surface (e.g., Wilson & Head, 1981). Given the relatively high pressure at Europa's seafloor (~100 MPa), we neglect the exsolution of volatiles; however, we note that extremely low fluid pressure at the tip of a propagating dike resulting from the large pressure gradient required to drive fluid into the narrow gap could result in exsolution even deep in the lithosphere (e.g., Lister & Kerr, 1991). Given our current lack of constraints on the composition and structure of Europa's silicate interior, and any magma plumbing system within it, we choose to focus on simple systems to address first-order questions.

The Weertman endmember model for dike propagation posits a static, fluid-filled crack initially in contact with the bottom of an elastic plate (the elastic lithosphere) under tensile stress and calculates the displacement of the crack tip due to the buoyant fluid (Pollard, 1976; Pollard & Muller, 1976; Secor & Pollard, 1975; Weertman, 1971a, 1971b). In this conception, crack growth is governed by a balance between buoyancy forces and the fracture toughness of the medium. When the fluid is positively buoyant, the displacement of the crack tip is positive (i.e., upward) and the crack will grow until it reaches a critical length, \( L_c \), given by, for example, Dahm (2000a) as

\[
L_c = \left[ \frac{k_c}{\Delta \rho g \sqrt{\pi}} \right]^{\frac{3}{2}}
\]

where \( k_c \) is the fracture toughness (with units of MPa m\(^{1/2}\)), \( g \) is the gravitational acceleration (1.3 m s\(^{-2}\)), and \( \Delta \rho \) is the density contrast between the fluid and the elastic plate. Once the critical length is reached, crack growth stops, and the entire crack begins to propagate upward by simultaneously opening at the crack tip and closing at the crack bottom. The result is an isolated, upward-propagating coin-shaped lens of fluid with length \( L_c \) and a narrow, inverted teardrop shape with half-width, \( w \), as a function of depth, \( z \), governed by (Dahm, 2000b; Weertman, 1980)

\[
w(z) = e \left( 1 + \frac{z}{L_c} \right) \sqrt{1 - \left( \frac{z}{L_c} \right)^2}
\]

where

\[
e = \frac{1}{2} \left( 1 - \nu \right) \frac{k_c}{G} \sqrt{L_c}
\]

and \( \nu \) is the Poisson's ratio (0.25) and \( G \) the shear modulus (30 GPa), both chosen for consistency with basalt. Increasing \( G \) by a factor of 2 (a value more consistent with dunite) would decrease the dike half-width by an equal factor. The resulting fluid lens is long in the vertical and one horizontal axis but very narrow in the other.

Whereas calculating the shape of the dike in this conceptional model is relatively straightforward, deriving the upward propagation velocity, \( v \), is less so. Dahm (2000a) used a finite difference model that accounts for both the pressure drop associated with fluid flow within the fracture (Nakashima, 1993; Nunn, 1996) and trailing fluids in the crack tail that prevent crack closure and retard upward propagation (e.g., Stevenson, 1982) to predict propagation velocities of 0.1–1 m s\(^{-1}\) (depending on the dike width) for terrestrial conditions and assuming a
large fracture toughness of 1,000 MPa m$^{1/2}$. The result is consistent with inferences of the rising velocities of mantle xenoliths (Dahm, 2000a). The propagation velocity is directly proportional to both the pressure gradient associated with viscous resistance and the square of the dike width, which are both larger for larger $k_c$ (Rivalta et al., 2015). Thus, smaller fracture toughness somewhat unintuitively results in smaller propagation velocities (a factor of 2 decrease in $k_c$ results in a factor of 5 decrease in $v$ (Dahm, 2000a)). The velocity also depends on the gravity of the body such that $v \sim g$. For identical material properties (to first order, silicate bodies across the Solar System have broadly similar peridotitic mantle compositions that generate basaltic melt (Lofgren et al., 1981)), propagation velocities are thus roughly an order of magnitude lower on Europa than on Earth. We therefore take 0.1 m s$^{-1}$ as the maximum propagation velocity for a Weertman style dike on Europa, assuming $k_c$ of 1,000 MPa m$^{1/2}$ (i.e., an order of magnitude lower than the maximum terrestrial velocity derived by Dahm (2000a)) and we use the relationship between $k_c$ and $v$ reported by Dahm (2000a) to conservatively estimate that for $k_c$ of 100 MPa m$^{1/2}$ the propagation velocity is $\leq 0.01$ m s$^{-1}$.

Of critical importance to this model is the fracture toughness of the medium, which influences both the geometry of the dike (Equations 1–3) and its propagation velocity. Unfortunately, $k_c$ is poorly known. Laboratory measurements typically find that $k_c$ is of order 1 MPa m$^{1/2}$ (e.g., Atkinson, 1984; Atkinson & Meredith, 1987; Balme et al., 2004) but inferences of the effective $k_c$ at outcrop scale suggest that it is at least an order of magnitude larger (e.g., Delaney & Pollard, 1981), and in some cases may exceed 1,000 MPa m$^{1/2}$ (Scholz, 2010). Complicating matters, $k_c$ is also influenced by rock tensile strength, confining pressure, temperature (e.g., DeGraff & Aydin, 1993; Rubin, 1993), and possibly the fracture length itself (Scholz, 2010; although see Olson, 2003; Olson & Schultz, 2011). Numerical approaches derive fracture toughness values of 100–1,000 MPa m$^{1/2}$ for terrestrial crustal rocks (Bunger & Cruden, 2011; Jin & Johnson, 2008; Rivalta & Dahm, 2006), consistent with inference from outcrop scales, but larger than laboratory values. Because of these uncertainties, we consistently treat the fracture toughness as a free parameter throughout this work.

The Weertman conception of dike propagation has faced criticism because the solutions are conceptually based on a stationary crack and generally neglect the fluid dynamics of the magma (Lister & Kerr, 1991). For small $k_c$, the strength of the country rock and associated elastic deformation can be neglected far from the crack tip (Emerman et al., 1986; Spence & Sharp, 1985; Spence & Turcotte, 1985) and dike growth is instead governed by a balance between buoyancy and the viscous resistance of the fluid (Lister, 1990a, 1990b; Lister & Kerr, 1991; Roper & Lister, 2007; Spence et al., 1987; Spence & Turcotte, 1990). This second endmember, fluid dynamic model of dike propagation postulates a planar, two-dimensional crack filled with fluid of viscosity $\mu$, which propagates through an infinite elastic medium (again the elastic lithosphere) due to buoyancy, $g\Delta \rho$, and is fed at depth by a constant flux, $Q_p$, per unit length (along strike) of dike (i.e., units of m$^3$ s$^{-1}$ m$^{-1}$, or m$^2$ s$^{-1}$). The resulting fracture shape includes a rounded head and long neck of approximately uniform thickness (Roper & Lister, 2007). Far from the dike tip, the half-width of the dike ($w_o$) is given by Roper and Lister (2007) as

$$w_o = \left(\frac{3\mu Q_p}{2\Delta \rho g}\right)^{1/3}$$

(4)

and the propagation rate ($v$) is given by

$$v = \left(\frac{Q_p^2 \Delta \rho g}{12\mu}\right)^{1/3}$$

(5)

The two models of dike emplacement (Weertman and fluid dynamics) can be reconciled by considering them as two endmembers (Rivalta et al., 2015; Roper & Lister, 2007), with the Weertman style analysis applicable to cases in which the fracture toughness is large (~1,000 MPa m$^{1/2}$), and the fluid dynamic analysis applicable when the elastic parameters of the lithosphere can be neglected (e.g., $k_c \leq 100$ MPa m$^{1/2}$). Below, we combine these two endmember models with a cooling model to consider the full range of conditions under which dike propagation through Europa’s lithosphere can occur. In both models, we implicitly assume that the magma is positively buoyant throughout the lithosphere, as would be expected for a basaltic crust (which we assume for simplicity to be equivalent to the cold lithosphere). Given the lack of constraints, we believe this assumption is appropriate to our first-order analysis.
3. Can Dikes Penetrate Europa’s Lithosphere?

Figure 1 shows the relationship between dike half-widths on Europa and both fracture toughness (Weertman model, panel a) and dike flux (viscous model, panel b). For the Weertman model and $\Delta \rho = 100 \text{ kg m}^{-2}$, dike half-width varies from 0.15 mm to 1.5 m for $k_f$ of 1 MPa m$^{-1/2}$ to 1,000 MPa m$^{-1/2}$, respectively. These rising lenses of magma are extremely long with critical lengths of 266 m to 26.6 km over the same range of $k_f$, giving them an aspect ratio of nearly 10$^4$:1 (length to full width). Increasing the density contrast between the magma and the lithosphere decreases both the critical dike length and maximum dike width because the larger buoyancy forces permit the dike to begin propagating at smaller sizes. Dikes on Europa are ~4x longer and twice as wide as terrestrial dikes, simply due to the nearly order of magnitude difference in gravity.

The extremely narrow dikes predicted by the Weertman model at small $k_f$ are expected to freeze rapidly (see below); however, in the limit of small fracture toughness (<100 MPa m$^{-1/2}$), dike formation is more likely governed by the fluid dynamics of the magma (Equation 4) rather than the mechanical properties of the lithosphere. Dike half-widths in the fluid dynamic regime are $\sim$1 m for fluxes of $\sim$1 m$^2$ s$^{-1}$ (Figure 1b). These fluxes are consistent with large terrestrial eruptions such as Iceland’s Laki fissure eruption (3 m$^2$ s$^{-1}$; Thordarson & Self, 1993) and result in propagation rates of 1–10 m/s. Smaller fluxes, such as those inferred for Piton de la Fournaise (0.02 m$^2$ s$^{-1}$; Traversa et al., 2010), result in dike widths of just 4–20 cm and propagation velocities of 0.1 m s$^{-1}$. In the fluid dynamic regime, the dependence of dike width on $g^{-1/3}$ results in dikes that are roughly twice the thickness of terrestrial dikes for the same density and viscosity.

Whether the dike can propagate through the lithosphere depends upon the propagation velocity and the solidification time. We use a simple symmetrical model of conductive dike cooling as befits our first-order approach (e.g., Turcotte and Schubert (2002))

$$t_c = \frac{W^2}{4k\lambda^2}$$

(6)

where $t_c$ is the solidification time, $W$ is the half thickness, $k$ is the thermal diffusivity, and $\lambda$ is found from the transcendental equation

$$\frac{L\sqrt{\pi}}{c\Delta T} = \frac{e^{-\lambda^2}}{\lambda(1 + \text{erf}(\lambda))}$$

(7)

where $L$ is the latent heat of fusion, $c$ is the heat capacity, and $\Delta T = T_m - T_i$ where $T_m$ and $T_i$ are the temperature of the melt and wall rock, respectively. This simple cooling model is for a stationary dike and neglects the complexities of wall melt back, viscous dissipation, and conduit blocking that can extend or reduce the dike lifetime (Bruce & Huppert, 1989; Carrigan et al., 1992). Additionally, flow in the dike may cease well before complete solidification occurs. Dike solidification time is shown as a function of dike half-width in Figure 2. For $\Delta T$ of 100 K, the narrow dikes that form when fracture toughness is small (1–10 MPa m$^{-1/2}$) and fluxes are very low (say $10^{-3}$ m$^2$ s$^{-1}$) cool in less than an hour. At the other extreme, if the fracture toughness is very large (~1,000 MPa m$^{-1/2}$) or small but the flux into the dike is large (1 m$^2$ s$^{-1}$) the dike half thickness is $\sim$1 m, and cooling times are on the order of 1 year. Increasing $\Delta T$ to 1000 K results in shorter cooling times due to the rapid conduction of heat away from the dike, which rapidly narrows.

In order for a dike to either erupt at Europa’s seafloor or intrude into the shallow lithosphere where ocean fluids circulate, it must propagate vertically by at least 100 km, which we take as the minimum lithospheric thickness based on the work of Běhounková et al. (2021). Combining the dike width, propagation velocity, and the

![Figure 1. The relationship between Europian dike half-width and (a) fracture toughness (the Weertman model from Equations 1–3) and (b) dike flux (the viscous model from Equation 4). In (a) blue, green, and orange curves are for $\Delta \rho$ of 100 kg m$^{-3}$, 200 kg m$^{-3}$, and 300 kg m$^{-3}$, respectively. In (b) blue, green, and orange curves are for $\mu$ of 1, 10, and 100 Pa s, respectively.](image-url)
Figure 2. Cooling time as a function of dike thickness (Equations 6 and 7). The corresponding fracture toughness is indicated on the upper axis (Weertman model) and the corresponding magma fluxes (viscous model) of 0.1 and 1 m² s⁻¹ are indicated by black brackets (maximums correspond to Δρ = 100 kg m⁻³ and μ = 100 Pa s; minimums correspond to Δρ = 300 kg m⁻³ and μ = 1 Pa s). Green and blue curves are for different ΔT, as indicated. Solid lines use κ = 10⁻⁶ m² s⁻¹, c = 1,200 kJ kg⁻¹ K⁻¹, and L = 320 kJ kg⁻¹. Dotted lines use κ = 0.5 × 10⁻⁶ m² s⁻¹, c = 1,200 kJ kg⁻¹ K⁻¹, and L = 400 kJ kg⁻¹. Dashed lines use κ = 0.5 × 10⁻⁶ m² s⁻¹, c = 1,000 kJ kg⁻¹ K⁻¹, and L = 400 kJ kg⁻¹.

Figure 3. Propagation distance for dikes on Europa. Lines show propagation distances for the viscous model and corresponds to the fluxes indicated on the horizontal axis (orange, green, and blue lines correspond to viscosities, μ, of 100, 1, and 1 Pa s, respectively; dashed, solid, and dotted lines correspond to temperature differences, ΔT, of 100 K, 600 K, and 1,000 K, respectively). The pink and purple shaded regions correspond to propagation distance derived for the Weertman model, with kₜ as indicated and assumed propagation velocities of 0.1, 0.02, and 0.01 m s⁻¹ for kₜ of 1,000, 500, and 100 MPa m⁻¹/², respectively. The upper bound of each assumes Δρ = 100 kg m⁻³ and ΔT = 100 K (both maximizing the distance), whereas the lower bound assumes Δρ = 300 kg m⁻³ and ΔT = 1000 K.

cooling timescale (Equations 6 and 7 and Figure 2) we calculate the propagation distance as a function of the input flux (Figure 3). For the Weertman model, cooling times (dike widths) and propagation velocities are sufficient to propagate through the lithosphere only when kₜ ≥ 500 MPa m⁻¹/². The more moderate case of kₜ = 100 MPa m⁻¹/² results in dikes widths of <10 cm and solidification times of hours to days. These dikes would require v ≥ 1 m s⁻¹ to penetrate the entire lithosphere: much larger than our estimated velocity of 0.01 m s⁻¹. More likely, dikes in this size range would propagate ~1 km and intrude the deep lithosphere. Such intrusions are still much too deep to enable hydrothermal activity, as ocean fluids are unlikely to circulate to depths of more than 25 km (S. Vance et al., 2007, S. D. Vance et al., 2016). When kₜ is small (1–10 MPa m⁻¹/²), dike propagation occurs in the viscous regime. We find that dikes fed by Qₓ ≥ 1 m² s⁻¹ can penetrate a 100 km thick lithosphere for many combinations of viscosity, density contrast, and temperature contrast. Dikes fed by Qₓ of 0.1 m² s⁻¹ can penetrate the lithosphere as long as ΔT ≤ 100 K. The smaller ΔT results in longer cooling times; however, this condition may be difficult to meet as ΔT is expected to increase as the dike rises into the colder part of the lithosphere. If dike fluxes are similar to those at Piton de la Fournaise (Qₓ ~ 0.02 m² s⁻¹), dikes will only propagate at most a few kilometers before freezing. Higher magma viscosity also increases propagation distance. The wider dikes and longer cooling times associated with the higher viscosity magma more than compensate for the lower propagation velocity.

The flux into the dike, Qₓ, is a critical unknown parameter. Elder and Bland (2018) find that for an asthenospheric tidal heating rate of 10¹³ W (consistent with “hot state” equilibrium values of Moore and Hussmann (2009)) melt forms and propagates upward by Darcy-like flow at a globally averaged rate of 0.05 m/yr. The total volumetric flux to the base of the lithosphere is simply given by Qₓ = α A, where α is the upward migration velocity and A is surface area. Assuming an asthenospheric radius of 1,360 km (200 km deeper than Europa’s surface), the global volumetric flux to the base of the lithosphere is 4 × 10¹⁰ m³ s⁻¹. This flux is quite large, similar to Io’s current inferred global surface effusion rate of 2 × 10¹⁰ m³ s⁻¹ (Blaney et al., 1995; although the flux to the base of Io’s lithosphere is likely an order of magnitude higher assuming terrestrial-like intrusion/extrusion ratios), and likely only relevant during past periods of high eccentricity (e.g., Hussmann & Spohn, 2004). In contrast, Běhounková et al. (2021) predict an average global flux more than an order of magnitude smaller: 2.5 × 10⁹ m³ s⁻¹ assuming continuous eruption. Běhounková et al. (2021) also calculated local melt volumes of ~10⁷ km³ in 250 Myrs for regions 270–380 km in radius, implying time averaged Qₓ of 4 m³ s⁻¹ from these regions of focused heating (tidal dissipation and magma generation are not uniformly distributed globally). Applying the flow rate derived by Elder and Bland (2018) to the same-sized region results in a flux of 360–610 m³ s⁻¹. It is plausible that these highly localized fluxes would feed a single dike-fed volcanic eruption. For comparison, Kilauea’s Pu‘u’ō‘ō sustained an effusion rate of 2.4–4.5 m³ s⁻¹ for many years (Kauahikaua et al., 1996; Sutton et al., 2001) and the 1984 Mauna Loa eruption saw magma production rates of 400 m³ s⁻¹ or more (Lipman & Banks, 1987).

Despite the similarity between the estimates of local magma flux to the base of Europa’s silicate lithosphere and terrestrial eruption rates, it is unlikely that the magma flux into a dike is in steady state with asthenospheric melt production. In such a scenario, seafloor volcanism on Europa is in any case,
implausible. Given the upward magma migration rate of 0.05 m yr$^{-1}$ of Elder and Bland (2018), and that a given dike requires $Q_d \sim 1$ m$^3$ s$^{-1}$ to reach the seafloor, a dike with an along-strike length of 100 km (similar to the vertical scale) would require a total flux of $10^5$ m$^3$ s$^{-1}$ (i.e., 1 m$^3$ s$^{-1}$ times 10$^5$ m of along-stripe length), which exceeds the estimated global magma flux to the base of the lithosphere described above. The implication is that for seafloor volcanism to occur, magma must accumulate to significant volumes at the base of the impermeable lithosphere, and then erupt rapidly (see, e.g., Havlin et al. (2013)). For example, if the eruption itself lasts one year, then to achieve the necessary total flux of $10^5$ m$^3$ s$^{-1}$ (for the 100 km long dike) the total magma volume required is $3 \times 10^{12}$ m$^3$, which requires just 2000 years to accumulate in a disk-shaped region 100 km in radius assuming continual magma production and the migration velocity of Elder and Bland (2018). If the eruption time is extended, the accumulation time must also be increased to offset the required additional melt volume and maintain a flux of $10^5$ m$^3$ s$^{-1}$. Thus, although seafloor volcanism on Europa cannot be continual even during periods of high tidal heating, the minimum time required to accumulate sufficient melt is short relative to typical geologic timescales. Seafloor volcanism might therefore provide new reactants to Europa's ocean with recharge cycles of several thousand years. Of utmost importance, however, is the coupling between magma accumulating at the base of the lithosphere and the initiation of diking (e.g., Havlin et al., 2013), which controls both the frequency at which dikes form and the total number of dikes. If dikes form too often, or if dikes are too numerous, the flux into each will be insufficient for dikes to propagate to the seafloor and will instead intrude deep in the lithosphere. Unless the fracture toughness of Europa's lithosphere is large (permitting Weertman-style dikes), seafloor eruptions on Europa must either be relatively voluminous ($>3$,000 km$^3$ yr$^{-1}$) or not occur at all. Such coupled models will be informed by new measurements of Europa's gravity field made by NASA's Europa Clipper mission that may constrain the structure of and heat flow from the silicate interior (Dombard & Sessa, 2019).
