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**Key Points:**

- For volcanism to provide life-sustaining reactants to Europa's ocean, asthenospheric-formed melt must traverse the cold thick lithosphere
- Dikes can propagate across the entire lithosphere if either the fracture toughness or the flux into the dike is large
- Volcanism plausibly provides reactants to Europa's ocean every several thousand years, but seafloor volcanism is far from assured

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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 $m^2\ s^{-1}$, 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; Sotin et al., 2002; 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, Běhounková 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 Běhounková 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 Běhounková 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/extrusive 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 (Rivalta et 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; Rivalta & Dahm, 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-pressurized 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

$$L_c = \left[\frac{k_c}{\Delta\rho g\sqrt{\pi}}\right]^{\frac{2}{3}} \tag{1}$$

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)

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

where

$$e = \frac{1}{2}\frac{(1-\upsilon)}{G}\frac{k_c}{\sqrt{\pi}}\sqrt{L_c} \tag{3}$$

and $\upsilon$ 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, $\Delta\rho g$, and is fed at depth by a constant flux, $Q_d$, 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 ($w_\infty$) is given by Roper and Lister (2007) as

$$w_\infty = \left(\frac{3\mu Q_d}{2\Delta\rho g}\right)^{\frac{1}{3}} \tag{4}$$

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

$$v = \left(\frac{Q_d^2 \Delta\rho g}{12\mu}\right)^{\frac{1}{3}} \tag{5}$$

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.

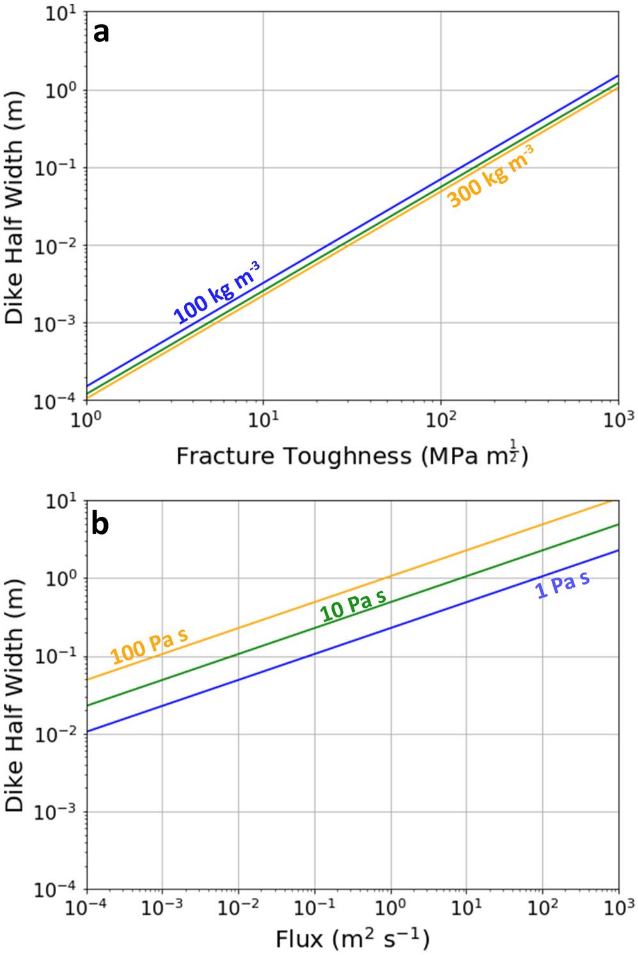


**Figure 1.** The relationship between Europan dike half-width and (a) fracture toughness (the Weertman model from Equations 1–3) and (b) dike flux (the viscous model from Equation 4). In (a) blue, green, and orange curves are for $\Delta\rho$ of 100 kg m$^{-3}$, 200 kg m$^{-3}$, and 300 kg m$^{-3}$, respectively. In (b) blue, green, and orange curves are for $\mu$ of 1, 10, and 100 Pa s, respectively.

## 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 $\Delta\rho = 100$ kg m$^{-3}$, dike half-width varies from 0.15 mm to 1.5 m for $k_c$ 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 $k_c$, 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 $k_c$ 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 ∼1 m for fluxes of ∼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))

$$t_c = \frac{W^2}{4\kappa\lambda^2} \tag{6}$$

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

$$\frac{L\sqrt{\pi}}{c\Delta T} = \frac{e^{-\lambda^2}}{\lambda(1+\mathrm{erf}(\lambda))} \tag{7}$$

where $L$ is the latent heat of fusion, $c$ is the heat capacity, and $\Delta T = T_m - T_s$ where $T_m$ and $T_s$ 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 ∼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

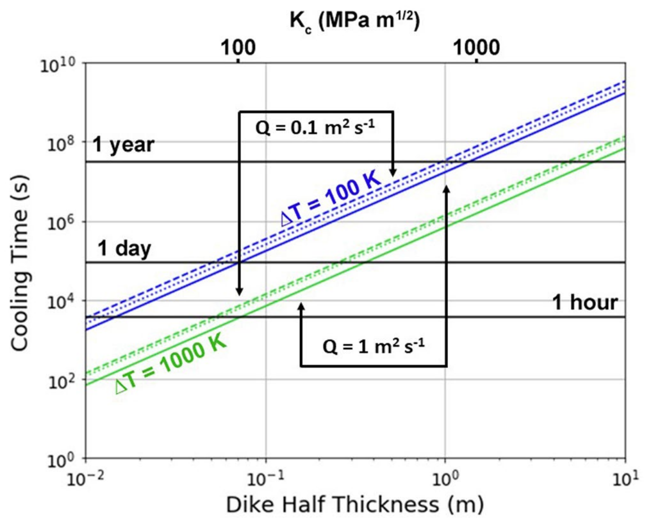


**Figure 2.** Cooling time as a function of dike thickness (Equations 6 and 7). The corresponding fracture toughness is indicated on the upper axis (Weertman model) and the corresponding magma fluxes (viscous model) of 0.1 and 1 $m^2 s^{-1}$ are indicated by black brackets (maximums correspond to $\Delta\rho = 100$ kg $m^{-3}$ and $\mu = 100$ Pa s; minimums correspond to $\Delta\rho = 300$ kg $m^{-3}$ and $\mu = 1$ Pa s). Green and blue curves are for different ΔT, as indicated. Solid lines use $\kappa = 10^{-6}$ $m^2 s^{-1}$, $c = 1{,}200$ kJ $kg^{-1} K^{-1}$, and $L = 320$ kJ $kg^{-1}$. Dotted lines use $\kappa = 0.5 \times 10^{-6}$ $m^2 s^{-1}$, $c = 1{,}200$ kJ $kg^{-1} K^{-1}$, and $L = 400$ kJ $kg^{-1}$. Dashed lines use $\kappa = 0.5 \times 10^{-6}$ $m^2 s^{-1}$, $c = 1{,}000$ kJ $kg^{-1} K^{-1}$, and $L = 400$ kJ $kg^{-1}$.

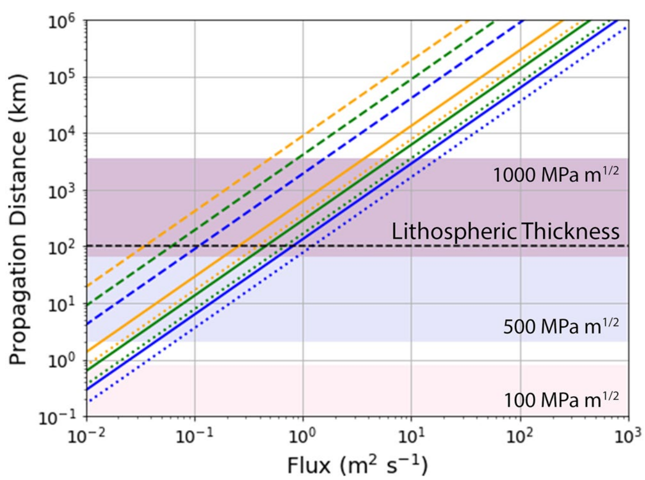


**Figure 3.** Propagation distance for dikes on Europa. Lines show propagation distances for the viscous model and corresponds to the fluxes indicated on the horizontal axis (orange, green, and blue lines correspond to viscosities, μ, of 100, 10, and 1 Pa s, respectively; dashed, solid, and dotted lines correspond to temperature differences, ΔT, of 100 K, 600 K, and 1000 K, respectively). The pink and purple shaded regions correspond to propagation distance derived for the Weertman model, with $k_c$ as indicated and assumed propagation velocities of 0.1, 0.02, and 0.01 m $s^{-1}$ for $k_c$ of 1,000, 500, and 100 MPa $m^{1/2}$, respectively. The upper bound of each assumes $\Delta\rho = 100$ kg $m^{-3}$ and $\Delta T = 100$ K (both maximizing the distance), whereas the lower bound assumes $\Delta\rho = 300$ kg $m^{-3}$ and $\Delta T = 1000$ 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 $k_c \geq 500$ MPa $m^{1/2}$. The more moderate case of $k_c = 100$ MPa $m^{1/2}$ results in dikes widths of <10 cm and solidification times of hours to days. These dikes would require $v \geq 1$ 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 $k_c$ is small (1–10 MPa $m^{1/2}$), dike propagation occurs in the viscous regime. We find that dikes fed by $Q_d \geq 1$ $m^2 s^{-1}$ can penetrate a 100 km thick lithosphere for many combinations of viscosity, density contrast, and temperature contrast. Dikes fed by $Q_d$ of 0.1 $m^2 s^{-1}$ can penetrate the lithosphere as long as $\Delta T \leq 100$ K. The smaller $\Delta T$ results in longer cooling times; however, this condition may be difficult to meet as $\Delta T$ 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 ($Q_d \sim 0.02$ $m^2 s^{-1}$), 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, $Q_d$, is a critical unknown parameter. Elder and Bland (2018) find that for an asthenospheric tidal heating rate of $10^{13}$ 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 $Q_v = u\,A$, where $u$ 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 $4 \times 10^4$ $m^3 s^{-1}$. This flux is quite large, similar to Io's current inferred global surface effusion rate of $2 \times 10^4$ $m^3 s^{-1}$ (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: $2.5 \times 10^3$ $m^3 s^{-1}$ assuming continuous eruption. Běhounková et al. (2021) also calculated local melt volumes of $\sim 10^7$ $km^3$ in 250 Myrs for regions 270–380 km in radius, implying time averaged $Q_v$ of 4 $m^3 s^{-1}$ 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 $m^3 s^{-1}$. It is plausible that these highly localized fluxes would feed a single dike-fed volcanic eruption. For comparison, Kīlauea's Puʻuʻōʻō sustained an effusion rate of 2.4–4.5 $m^3 s^{-1}$ for many years (Kauahikaua et al., 1996; Sutton et al., 2001) and the 1984 Mauna Loa eruption saw magma production rates of 400 $m^3 s^{-1}$ 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$^2$ 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$^2$ s$^{-1}$ times $10^5$ m of along-strike 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).

## Data Availability Statement

No data, code, or software was used for this study.

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## References

Atkinson, B. K. (1984). Subcritical crack growth in geological material. *Journal of Geophysical Research*, *89*, 4077–4114. https://doi.org/10.1029/jb089ib06p04077

Atkinson, B. K., & Meredith, P. G. (1987). Experimental fracture mechanics data for rocks and minerals. In B. K. Atkinson (Ed.), *Fracture Mechanics of rock* (pp. 477–525). Academic Press. https://doi.org/10.1016/b978-0-12-066266-1.50016-8

Balme, M. R., Rocchi, V., Jones, C., Sammonds, P. R., Meredith, P. G., & Boon, S. (2004). Fracture toughness measurements on igneous rocks using a high-pressure, high-temperature rock fracture mechanics cell. *Journal of Volcanology and Geothermal Research*, *132*, 159–172. https://doi.org/10.1016/S0377-0273(03)00343-3

Běhounková, M., Tobie, G., Choblet, G., Kervazo, M., Melwani Daswani, M., Dumoulin, C., & Vance, S. D. (2021). Tidally induced magmatic pulses on the oceanic floor of Jupiter's moon Europa. *Geophysical Research Letters*, *48*, e90077. https://doi.org/10.1029/2020GL090077

Bierhaus, E. B., Zahnle, K., & Chapman, C. R. (2009). Europa's crater distributions and surface ages. In R. T. Pappalardo, W. B. McKinnon, & K. Khurana (Eds.), *Europa* (pp. 161–180). The University of Arizona Press.

Bland, M. T., & McKinnon, W. B. (2019). *Folding, faulting, and subduction on Europa and Enceladus*. EPSC-DPS Joint Meeting. EPSC-DPS2019-1211.

Blaney, D. L., Johnson, T. V., Matson, D. L., & Veeder, G. J. (1995). Volcanic eruptions on Io: Heat flow, resurfacing, and lava composition. *Icarus*, *113*, 220–225. https://doi.org/10.1006/icar.1995.1020

Bruce, P. M., & Huppert, H. E. (1989). Thermal control of basaltic fissure eruptions. *Nature*, *342*, 665–667. https://doi.org/10.1038/342665a0

Bunger, A. P., & Cruden, A. R. (2011). Modeling the growth of laccoliths and large mafic sills: Role of magma body forces. *Journal of Geophysical Research*, *116*, B02203. https://doi.org/10.1029/2010jb007648

Carrigan, C. R., Schubert, G., & Eichelberger, J. C. (1992). Thermal and dynamical regimes of single- and two-phase magmatic flow in dikes. *Journal of Geophysical Research*, *97*, 17377–17392. https://doi.org/10.1029/92jb01244

Christner, B. C., Royston-Bishop, G., Foreman, C. M., Arnold, B. R., Tranter, M., Welch, K. A., et al. (2006). Limnological conditions in subglacial Lake Vostok, Antarctica. *Limnology & Oceanography*, *51*, 2485–2501. https://doi.org/10.4319/lo.2006.51.6.2485

Chyba, C. F., & Hand, K. P. (2001). Life without photosynthesis. *Science*, *292*, 2026–2027. https://doi.org/10.1126/science.1060081

Crisp, J. (1984). Rates of magma emplacement and volcanic output. *Journal of Volcanology and Geothermal Research*, *20*, 177–211. https://doi.org/10.1016/0377-0273(84)90039-8

Dahm, T. (2000a). Numerical simulations of the propagation path and the arrest of fluid-filled fractures in the Earth. *Geophysical Journal International*, *141*, 623–638. https://doi.org/10.1046/j.1365-246x.2000.00102.x

Dahm, T. (2000b). On the shape and velocity of fluid-filled fractures in the Earth. *Geophysical Journal International*, *142*, 181–192. https://doi.org/10.1046/j.1365-246x.2000.00148.x

DeGraff, J. M., & Aydin, A. (1993). Effect of thermal regime on growth increment and spacing of contraction joints in basaltic lava. *Journal of Geophysical Research*, *98*, 6411–6430. https://doi.org/10.1029/92JB01709

Delaney, P., & Pollard, D. (1981). Deformation of host rocks and flow of magma during growth of minette dikes and breccia-bearing intrusions near Ship Rock. *U. S. Geological Survey Professional Paper*, (Vol. *1202*). https://doi.org/10.3133/pp1202

Dombard, A. J., & Sessa, A. M. (2019). Gravity measurements are key in addressing the habitability of a subsurface ocean in Jupiter's moon Europa. *Icarus*, *325*, 31–38. https://doi.org/10.1016/j.icarus.2019.02.025

Elder, C. M. (2015). *The effects of melt on impact craters on icy satellites and the dynamics of Io's interior (Dissertation)*. University of Arizona. Retrieved from http://hdl.handle.net/10150/556825

Elder, C. M., & Bland, M. T. (2018). *The Plausibility of Silicate Volcanism on Europa's Seafloor* (Vol. P41B-07). AGU Fall meeting.

Emerman, S. H., Turcotte, D. L., & Spence, D. A. (1986). Transport of magma and hydrothermal solutions by laminar and turbulent fluid fracture. *Physics of the Earth and Planetary Interiors*, *41*, 249–259. https://doi.org/10.1016/0031-9201(86)90004-x

Fialko, Y., & Rubin, A. (1999). Thermal and mechanical aspects of magma emplacements in giant dike swarms. *Journal of Geophysical Research*, *104*, 23033–23049. https://doi.org/10.1029/1999jb900213

Gaidos, E. J., Nealson, K. H., & Kirschvink, J. L. (1999). Life in ice-covered oceans. *Science*, *284*, 1631–1633. https://doi.org/10.1126/science.284.5420.1631

Geissler, P. E., O'Brien, D. P., & Greenberg, R. (2001). Silicate volcanism on Europa. *Lunar and Planetary Science Conference*, *32*, 2068. https://doi.org/10.1029/2021GL096939

Greeley, R., Chyba, C. F., Head, J. W., McCord, T. B., McKinnon, W. B., Pappalardo, R. T., & Figueredo, P. H. (2004). Geology of Europa. In F. Bagenal, T. Dowling, & W. McKinnon (Eds.), *Jupiter: The planet, satellites and Magnetosphere* (pp. 329–362). Cambridge University Press.

Greenberg, R. (2010). Transport rates of radiolytic substances into Europa's ocean: Implications for the potential origin and maintenance of life. *Astrobiology*, *10*, 275–283. https://doi.org/10.1089/ast.2009.0386

Greenberg, R., Geissler, P., Tufts, B. R., & Hoppa, G. V. (2000). Habitability of Europa's crust: The role of tidal-tectonic processes. *Journal of Geophysical Research*, *105*, 17551–17562. https://doi.org/10.1029/1999je001147

Hand, K. P., Chyba, C. F., Priscu, J. C., Carlson, R. W., & Nealson, K. H. (2009). Astrobiology and the potential for life on Europa. In R. T. Pappalardo, W. B. McKinnon, & K. Khurana (Eds.), *Europa* (pp. 589–630). The University of Arizona Press.

Havlin, C., Parmentier, E. M., & Hirth, G. (2013). Dike propagation driven by melt accumulation at the lithosphere-asthenosphere boundary. *Earth and Planetary Science Letters*, *376*, 20–28. https://doi.org/10.1016/j.epsl.2013.06.010

Howell, S. M. (2021). The likely thickness of Europa's icy shell. *Planetary Science Journal*, *2*, 129. https://doi.org/10.3847/PSJ/abfe10

Howell, S. M., & Pappalardo, R. T. (2019). Can Earth-like plate tectonics occur in ocean world ice shells? *Icarus*, *322*, 69–79. https://doi.org/10.1016/j.icarus.2019.01.011

Hussmann, H., & Breuer, D. (2007). *The thermo-chemical evolution of Europa: Constraints on silicate volcanism at Europa's ocean floor* (pp. P53B–P1239). AGU Fall Meeting.

Hussmann, H., & Spohn, T. (2004). Thermal-orbital evolution of Io and Europa. *Icarus*, *171*, 391–410. https://doi.org/10.1016/j.icarus.2004.05.020

Jin, Z. H., & Johnson, S. (2008). Magma-driven multiple dike propagation and fracture toughness of crustal rocks. *Journal of Geophysical Research*, *113*, B03206. https://doi.org/10.1029/2006JB004761

Johnson, B. C., Sheppard, R. Y., Pascuzzo, A. C., Fisher, E. A., & Wiggins, S. E. (2017). Porosity and salt content determine if subduction can occur in Europa's ice shell. *Journal of Geophysical Research*, *122*, 2765–2778. https://doi.org/10.1002/2017JE005370

Kattenhorn, S. A., & Prockter, L. M. (2014). Evidence for subduction in the ice shell of Europa. *Nature Geoscience*, *7*, 762–767. https://doi.org/10.1038/NGEO2245

Kauahikaua, J., Mangan, M., Heliker, C., & Mattox, T. (1996). A quantitative look at the demise of a basaltic vent: The death of Kupaianaha, Kilauea Volcano, Hawai'i. *Bulletin of Volcanology*, *57*, 641–648. https://doi.org/10.1007/s004450050117

Kavanagh, J. L., Menand, T., & Sparks, R. S. J. (2006). An experimental investigation of sill formation and propagation in layered elastic media. *Earth and Planetary Science Letters*, *245*, 799–813. https://doi.org/10.1144/0016-76492010-137

Kuhn, D., & Dahm, T. (2008). Numerical modelling of dyke interaction and its influence on ocean crust formation. *Tectonophysics*, *447*, 53–65. https://doi.org/10.1016/j.tecto.2006.09.018

Lipman, P., & Banks, N. (1987). Aa flow dynamics, 1984 Mauna Loa. In R. W. Decker, T. L. Wright, & P. H. Stauffer (Eds.), *Volcanism in Hawaii*, USGS professional paper 1350 (pp. 1527–1567). https://doi.org/10.3133/pp1350

Lister, J. R. (1990a). Buoyancy-driven fluid fracture: Similarity solutions for the horizontal and vertical propagation of fluid-filled cracks. *Journal of Fluid Mechanics*, *217*, 213–239. https://doi.org/10.1017/s0022112090000696

Lister, J. R. (1990b). Buoyancy-driven fluid fracture: The effects of material toughness and of low-viscosity precursors. *Journal of Fluid Mechanics*, *210*, 263–280. https://doi.org/10.1017/s0022112090001288

Lister, J. R. (1991). Steady solutions for feeder dykes in a density-stratified lithosphere. *Earth and Planetary Science Letters*, *107*, 233–242. https://doi.org/10.1016/0012-821x(91)90073-q

Lister, J. R., & Kerr, R. C. (1991). Fluid-mechanical models of crack propagation and their application to magma transport in dykes. *Journal of Geophysical Research*, *96*, 10049–10077. https://doi.org/10.1029/91jb00600

Lofgren, G. E., Bence, A. E., Duke, M. B., Dungan, M. A., Green, J. C., Haggerty, S. E., et al. (1981). Petrology and chemistry of terrestrial, lunar, and meteoritic basalts. In *Basaltic volcanism on the terrestrial planets, basaltic volcanism study project*. Pergamon Press, Inc.

Maccaferri, F., Bonafede, M., & Rivalta, E. (2010). A numerical model of dyke propagation in layered elastic media. *Geophysical Journal International*, *180*, 1107–1123. https://doi.org/10.1111/j.1365-246x.2009.04495.x

McCollom, T. M. (1999). Methanogenesis as a potential source of chemical energy for primary biomass production by autotrophic organisms in hydrothermal systems on Europa. *Journal of Geophysical Research*, *104*, 30729–30742. https://doi.org/10.1029/1999je001126

Menand, T., Daniels, K. A., & Benghiat, P. (2010). Dyke propagation and sill formation in a compressive tectonic environment. *Journal of Geophysical Research*, *115*, B08201. https://doi.org/10.1029/2009JB006791

Meriaux, C., & Jaupart, C. (1998). Dike propagation through an elastic plate. *Journal of Geophysical Research*, *103*, 18295–18314. https://doi.org/10.1029/98jb00905

Mitri, G., & Showman, A. P. (2005). Convective-conductive transitions and sensitivity of a convecting ice shell to perturbations in heat flux and tidal-heating rate: Implications for Europa. *Icarus*, *177*, 447–460. https://doi.org/10.1016/j.icarus.2005.03.019

Moore, W. B. (2001). The thermal state of Io. *Icarus*, *154*, 548–550. https://doi.org/10.1006/icar.2001.6739

Moore, W. B. (2003). Tidal heating and convection in Io. *Journal of Geophysical Research*, *108*, 5096. https://doi.org/10.1029/2002JE001943

Moore, W. B., & Hussmann, H. (2009). Thermal evolution of Europa's silicate interior. In R. T. Pappalardo, W. B. McKinnon, & K. Khurana (Eds.), *Europa* (pp. 369–380). The University of Arizona Press.

Nakashima, Y. (1993). Static stability and propagation of a fluid-filled edge crack in rock: Implication for fluid transport in magmatism and metamorphism. *Journal of Physics of the Earth*, *41*, 189–202. https://doi.org/10.4294/jpe1952.41.189

Nunn, J. A. (1996). Buoyancy-driven propagation of isolated fluid-filled fractures: Implications for fluid transport in Gulf of Mexico geopressured sediments. *Journal of Geophysical Research*, *101*, 2963–2970. https://doi.org/10.1029/95jb03210

Olson, J. E. (2003). Sublinear scaling of fracture aperture versus length: An exception or the rule? *Journal of Geophysical Research*, *108*, 2413. https://doi.org/10.1029/2001JB000419

Olson, J. E., & Schultz, R. A. (2011). Comment on "A note on the scaling relations for opening mode fractures in rock" by C H. Scholz. *Journal of Structural Geology*, *33*, 1523–1524. https://doi.org/10.1016/j.jsg.2011.07.004
Pappalardo, R. T., Head, J. W., Greeley, R., Sullivan, R. J., Pilcher, C., Schubert, G., et al. (1998). Geological evidence for solid-state convection in Europa's ice shell. *Nature*, *391*, 365–368. https://doi.org/10.1038/34862
Parfitt, E. A., Wilson, L., & Head, J. W. (1993). Basaltic magma reservoirs: Factors controlling their rupture characteristics and evolution. *Journal of Volcanology and Geothermal Research*, *55*, 1–14. https://doi.org/10.1016/0377-0273(93)90086-7
Pollard, D. D. (1976). On the form and stability of open hydraulic fractures in the Earth's crust. *Geophysical Research Letters*, *3*, 513–516. https://doi.org/10.1029/gl003i009p00513
Pollard, D. D., & Holzhausen, G. (1979). On the mechanical interaction between a fluid-filled fracture and the Earth's surface. *Tectonophysics*, *53*, 27–57. https://doi.org/10.1016/0040-1951(79)90353-6
Pollard, D. D., & Muller, O. (1976). The effect of gradients in regional stress and magma pressure on the form of sheet intrusions in cross section. *Journal of Geophysical Research*, *81*, 975–984. https://doi.org/10.1029/jb081i005p00975
Rivalta, E., & Dahm, T. (2006). Acceleration of buoyancy-driven fractures and magmatic dikes beneath the free surface. *Geophysical Journal International*, *166*, 1424–1439. https://doi.org/10.1111/j.1365-246X.2006.02962.x
Rivalta, E., Taisne, B., Bunger, A. P., & Katz, R. F. (2015). A review of mechanical models of dike propagation: Schools of thought, results and future directions. *Tectonophysics*, *638*, 1–42. https://doi.org/10.1016/j.tecto.2014.10.003
Roper, S. M., & Lister, J. R. (2007). Buoyancy-driven crack propagation: The limit of large fracture toughness. *Journal of Fluid Mechanics*, *580*, 359–380. https://doi.org/10.1017/S0022112007005472
Rubin, A. M. (1993). Tensile fracture of rock at high confining pressure: Implications for dike propagation. *Journal of Geophysical Research*, *98*, 15919–15935. https://doi.org/10.1029/93jb01391
Scholz, C. H. (2010). A note on the scaling relations for opening mode fractures in rock. *Journal of Structural Geology*, *32*, 1485–1487. https://doi.org/10.1016/j.jsg.2010.09.007
Secor, J. R., & Pollard, D. D. (1975). On the stability of open hydraulic fractures in the Earth's crust. *Geophysical Research Letters*, *2*, 510–513. https://doi.org/10.1029/gl002i011p00510
Singer, K. N., McKinnon, W. B., & Schenk, P. M. (2021). Pits, uplifts and small chaos features on Europa: Morphologic and morphometric evidence for intrusive upwelling and lower limits to ice shell thickness. *Icarus*, *364*, 114465. https://doi.org/10.1016/j.icarus.2021
Sotin, C., Head, J. W., & Tobie, G. (2002). Europa: Tidal heating of upwelling thermal plumes and the origin of lenticulae and chaos melting. *Geophysical Research Letters*, *29*, 1233–1241. https://doi.org/10.1029/2001GL013844
Spence, D. A., & Sharp, P. W. (1985). Self-similar solution for elastohydrodynamic cavity flow. *Proceedings of the Royal Society of London*, *A400*, 289–313.
Spence, D. A., Sharp, P. W., & Turcotte, D. L. (1987). Buoyancy-driven crack propagation: A mechanism for magma migration. *Journal of Fluid Mechanics*, *174*, 135–153. https://doi.org/10.1017/s0022112087000077
Spence, D. A., & Turcotte, D. L. (1985). Magma-driven propagation of cracks. *Journal of Geophysical Research*, *90*, 575–580. https://doi.org/10.1029/jb090ib01p00575
Spence, D. A., & Turcotte, D. L. (1990). Buoyancy-driven magma fracture: A mechanism for ascent through the lithosphere and the emplacement of diamonds. *Journal of Geophysical Research*, *95*, 5133–5139. https://doi.org/10.1029/jb095ib04p05133
Stevenson, D. J. (1982). Migration of fluid-filled cracks: Application to terrestrial and icy bodies. *Lunar and Planetary Science Conference*, 768–769.
Sutton, A. J., Elias, T., Gerlach, T. M., & Stokes, J. B. (2001). Implications for eruptive processes as indicated by sulfur dioxide emissions from Kīlauea volcano, Hawai'i 1979-1997. *Journal of Volcanology and Geothermal Research*, *108*, 283–302. https://doi.org/10.1016/s0377-0273(00)00291-2
Taisne, B., & Jaupart, C. (2009). Dike propagation through layered rocks. *Journal of Geophysical Research*, *114*, B09203. https://doi.org/10.1029/2008JB006228
Thordarson, T., & Self, S. (1993). The Laki (Skaftár Fires) and Grímsvötn eruptions in 1783-1785. *Bulletin of Volcanology*, *55*, 233–263. https://doi.org/10.1007/bf00624353
Tobie, G., Choblet, G., & Sotin, C. (2003). Tidally heated convection: Constraints on Europa's ice shell thickness. *Journal of Geophysical Research*, *108*, 5124. https://doi.org/10.1029/2003je002099
Traversa, P., Pinel, V., & Grasso, J. R. (2010). A constant influx model for dike propagation: Implications for magma reservoir dynamics. *Journal of Geophysical Research*, *115*, B01201. https://doi.org/10.1029/2009JB006559
Turcotte, D. L., & Schubert, D. (2002). *Geodynamics* (2nd ed.). Cambridge University Press.
Turtle, E. P., & Pierazzo, E. (2001). Thickness of a Europan ice shell from impact crater simulations. *Science*, *294*, 1326–1328, 1326. https://doi.org/10.1126/science.1062492
Vance, S., Harnmeijer, J., Kimura, J., Hussmann, H., deMartin, B., & Brown, M. (2007). Hydrothermal systems in small ocean planets. *Astrobiology*, *7*, 987–1005. https://doi.org/10.1089/ast.2007.0075
Vance, S. D., Hand, K. P., & Pappalardo, R. T. (2016). Geophysical controls of chemical disequilibrium in Europa. *Geophysical Research Letters*, *43*, 487–4879. https://doi.org/10.1002/2016GL068547
Weertman, J. (1971a). Theory of water-filled crevasses in glaciers applied to vertical magma transport beneath oceanic ridges. *Journal of Geophysical Research*, *76*, 1171–1183. https://doi.org/10.1029/jb076i005p01171
Weertman, J. (1971b). Velocity at which liquid-filled cracks move in the Earth's crust or in glaciers. *Journal of Geophysical Research*, *76*, 8544–8553. https://doi.org/10.1029/jb076i035p08544
Weertman, J. (1980). The stopping of a rising, liquid-filled crack in the Earth's crust by a freely slipping horizontal joint. *Journal of Geophysical Research*, *85*, 967–976. https://doi.org/10.1029/JB085iB02p00967
White, S. M., Crisp, J. A., & Spera, F. J. (2006). Long-term volumetric eruption rates and magma budgets. *Geochemistry, Geophysics, Geosystems*, *7*, Q03010. https://doi.org/10.1029/2005GC001002
Wilson, L., & Head, J. W. (1981). Ascent and eruption of basaltic magma on the Earth and Moon. *Journal of Geophysical Research*, *86*, 2971–3001. https://doi.org/10.1029/jb086ib04p02971
Wilson, L., & Head, J. W. (1992). Magma reservoirs and neutral buoyancy zones on Venus: Implications for the formation and evolution of volcanic landforms. *Lunar and Planetary Science Conference*, *23*, 1533.
Wilson, L., & Head, J. W. (2002). Tharsis-radial graben systems as the surface manifestation of plume-related dike intrusion complexes: Models and implications. *Journal of Geophysical Research*, *107*, 5057. https://doi.org/10.1029/2001JE001593
Zolotov, M. Y., & Shock, E. L. (2004). A model for low-temperature biogeochemistry of sulfur, carbon, and iron on Europa. *Journal of Geophysical Research*, *109*, E06003. https://doi.org/10.1029/2003JE002194