1 **Heat Transfer of Ascending Cryomagma on Europa**

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Abstract: Jupiter's moon Europa has a relatively young surface (60-90 Myr on average), which may be due in part to cryovolcanic processes. Current models for both effusive and explosive cryovolcanism on Europa may be expanded and enhanced by linking the potential for cryovolcanism at the surface to subsurface cryomagmatism. The success of cryomagma transport through Europa's crust depends critically on the rate of ascent relative to the rate of solidification. The final transport distance of cryomagma is thus governed by initial melt volume, ascent rate, overall ascent distance, transport mechanism (i.e., diapirism, diking, or ascent in cylindrical conduits), and melt temperature and composition. The last two factors are especially critical in determining the budget of expendable energy before complete solidification. Here we use these factors as constraints to explore conditions under which cryomagma may arrive at Europa's surface to facilitate cryovolcanism. We find that 1-5 km radius warm ice diapirs ascending from the base of a 10 km thick stagnant lid can reach the shallow subsurface in a partially molten state. Cryomagma transport may be further facilitated if diapirs travel along pre-heated ascent paths. Under certain conditions, cryolavas transported from 10 km depths in tabular dikes or pipe-like conduits may reach the surface at temperatures exceeding 250 K. Ascent rates for these geometries may be high enough that isothermal transport is approached. Cryomagmas containing significant amounts of low eutectic impurities can also be delivered to Europa's surface by propagating dikes or pipe-like conduits.

1 Introduction

The surface of Jupiter's moon Europa is littered with many putative cryovolcanic features (Fig. 1). While several of these features are thought to be evidence of effusive

1 flows (Wilson et al., 1997; Figueredo et al., 2002; Fagents 2003), it is possible that low-
2 albedo deposits that surround lenticulae and lie along lineated features may be cryoclastic
3 mantlings deposited during explosive eruptions (Pappalardo et al., 1999; Fagents et al.,
4 2000; Quick et al., 2013). Moreover, lenticulae, chaotic terrain, and ridge and band
5 formation may be surface manifestations of subsurface cryomagmatic processes
6 associated with diapirism (Pappalardo et al., 1998, 1999; Rathbun et al., 1998; Collins et
7 al., 2000; Fagents et al., 2000; Figueredo et al., 2002; Prockter et al., 2002; Sotin et al.,
8 2002; Fagents, 2003; Tobie et al., 2003; Pappalardo and Barr, 2004; Han and Showman,
9 2005; Prockter and Schenk, 2005; Schmidt et al., 2011), diking, and sill emplacement
10 (Greenberg et al., 1998; Johnston and Montési, 2014; Michaut and Manga, 2014; Craft et
11 al., 2016).

12 Therefore, **models for cryovolcanism on Europa** (cf. Crawford and Stevenson,
13 1988; Wilson et al., 1997; Wilson and Head, 1998; Fagents et al., 2000; Fagents, 2003,
14 Quick et al., 2013) **can be** expanded and enhanced by linking observational evidence for
15 cryovolcanism at the surface to **inferred** subsurface cryomagmatic processes. **Our** work
16 attempts to provide this linkage by evaluating the **heat transfer associated with** specific
17 styles of cryomagma transport in Europa's interior. **We aim to place rigorous bounds on**
18 **the tradeoffs between magma volume, ascent rate, and body geometry to achieve**
19 **cryovolcanism at the surface. In exploring these limits, constraints are also placed on the**
20 **physics of possible ascent mechanisms.**

Figure 1. Putative cryovolcanic features on Europa include (a) a smooth circular feature located near the equator; (b) dark deposits along Rhadamanthys Linea; (c) putative flows at the edge of Thrace Macula; (d) Murias Chaos (The Mitten).

1.1 Silicate Magma vs. Cryomagma

A particularly striking contrast exists between cryovolcanism on icy satellites and terrestrial volcanism. Terrestrial volcanism is distinct in three important ways: first, because silicate magma has an adiabat that is much steeper than its liquidus, which itself decreases with decreasing pressure or depth, terrestrial magmas tend to become strongly superheated during ascent (e.g., Marsh, 2007); second, because silicate magmas are multi-component systems, they possess a large crystallization interval between the liquidus and solidus of 200 degrees or more. This interval produces a substantial sub-liquidus corridor or “mush zone”, consisting of a highly mobile mix of melt and crystals, which can extrude onto the surface as lava (Fig. 2); third, silicate magmas are generally

1 less dense than the country rock through which they must ascend. These three features of
2 silicate magmas ensure that the associated volcanism will be frequent and multifaceted.

3 Conversely, for warm ice and/or relatively pure water, which likely serve as
4 magmatic fluids on icy satellites like Europa, there is virtually no solidus-liquidus gap or
5 mush corridor, and the melting point slightly *increases* with decreasing pressure or depth.

6 As a result, ascending cryomagma tends to become supercooled. If the the liquidus is
7 crossed, the cryomagma has no pressure-temperature corridor of partial melt available,
8 and further ascent is immediately arrested by quenching (Fig. 2a). Furthermore, water
9 and other briny solutions that may serve as cryomagmas are negatively buoyant with
10 respect to the surrounding ice they must traverse. For this reason, these fluids may never
11 rise to the surface except for the case of pressure-driven ascent, or when significant
12 amounts of volatiles are included in solution (Crawford and Stevenson, 1988; Fagents et
13 al., 2000; Muñoz-Iglesias et al., 2014; Neveu et al., 2015). It is therefore essential to
14 determine the heat transfer parameters, transport mechanisms, and ascent rates that will
15 enable cryomagmas to: (1) successfully reach Europa's surface and (2) arrive before
16 undergoing thermal death by solidification.

17

Phase Relations for Basalts and Water - Ice

Figure 2a. Phase diagrams for basaltic magmas (left) and pure H₂O cryomagmas (right). As basaltic magma rises to Earth's surface, it tends to become superheated and a mush zone of partial melt exists between the liquidus and solidus. Pure H₂O cryomagmas tend to supercool in rising to the surface, and no mush zone exists. Note that 1 kbar = 100 MPa. (See also Fig. 2b).

Figure 2b. The addition of low-eutectic salts or mineral acids may allow for slight superheating of cryomagmas on Europa, and the formation of a mush zone of partial melt. Note that 1 kbar = 100 MPa.

1.2 Briny Cryomagmas

Data returned from Galileo's Near-Infrared Mapping Spectrometer (NIMS) suggests the presence of hydrated minerals on Europa's surface. These minerals have been identified as hydrated Mg- and Na-sulfate salts and are mainly found near recently disrupted surface areas and potential sites of extrusion of subsurface materials, i.e., along lineae and near chaotic terrain (McCord et al., 1998; 1999). Linear mixture modeling has confirmed that these species are prevalent in low-albedo surface units on Europa (Dalton et al. 2005; Shirley et al., 2010; Dalton et al., 2012). It is possible that these hydrated minerals originated in the subsurface ocean (Kargel et al., 2000; Spaun & Head, 2001), and previous workers have suggested that they may play a critical role in cryovolcanic processes on Europa and other icy satellites (Kargel, 1991; Hogenboom et al., 1995; Spaun and Head, 2001; Muñoz-Iglesias et al., 2013; 2014). Brown and Hand (2013) have suggested that ocean-derived salts on Europa may be chlorides instead of sulfates. These authors argue that salts such as MgCl_2 , KCl , and NaCl may be endogenically produced and quite abundant on Europa. If this is indeed the case, then chloride salts could also be constituents of European cryomagmas.

Sulfuric acid and its hydrates have also been detected on Europa's surface (Carlson et al., 1999; 2002; Dalton et al., 2005 and references therein). While there is sufficient evidence that these species may have been exogenically emplaced (Carlson et al., 1999; Dalton et al., 2007; 2012; 2013), previous work by Kargel et al. (2000) and Pasek and Greenberg (2012) suggests that Europa's ocean could be highly acidic and that sulfuric acid and its hydrates could be endogenic in origin. It has also been suggested that low-albedo spots on Europa's surface could be products of small-scale flooding of

1 sulfuric acid species (Fagents, 2003). Furthermore, it is possible that surface species on
2 Europa may be cycled into the interior by impact gardening (Cooper et al., 2001), crustal
3 cycling (Prockter and Pappalardo, 2000; Prockter et al., 2002; Kattenhorn and Prockter,
4 2014), and ridge formation (Greenberg et al., 2010), among other processes. This could
5 allow exogenically produced compounds like sulfuric acid to participate in geophysical
6 processes on the icy moon. Hence even if exogenically emplaced, it is still possible that
7 sulfuric acid species could participate in cryovolcanic processes on Europa. We therefore
8 consider the possibility that European cryomagmas could also consist of sulfuric acid
9 hydrates, albeit with the aforementioned caveats for exogenic emplacement.

10 The addition of any of these components to an H₂O-dominated cryomagma may
11 allow for a 5-75 K separation of the liquidus and solidus and the formation of a
12 cryomagmatic mush zone (Fig. 2b). As previously discussed in Section 1, and as
13 illustrated in Fig. 2a and b, when considering cryomagmatism on icy satellites, the
14 formation of this mush zone may be of critical importance, as low-eutectic contaminants
15 can aid in separating the liquidus from the solidus of cryomagmas and will ensure that a
16 partial melt zone exists (Fig. 2b).

17 Possible brine-containing aqueous solutions that may erupt on Europa as a result
18 of overpressurization of fluid reservoirs within the ice shell are listed in Table I. This
19 Table includes the properties of these solutions including their wt% of salt or acid,
20 eutectic temperatures, and densities as provided in Kargel (1991), McCarthy et al. (2007),
21 Green and Perry (2007) and Brady (2009). We note here that for certain species, it is only
22 with substantial concentrations of added salts or mineral acids that the eutectic is
23 significantly affected. This is the case for calcium chloride hexahydrate and the hydrated

sulfuric acids, where eutectic temperatures may be as low as 223 K and 198 K, respectively at concentrations of 30-37 wt.%. Conversely, adding smaller amounts of hydrated NaCl, MgCl₂, KCl, MgSO₄, and Na₂SO₄ serves to reduce the eutectic by, at most, about 34 K (see Table 1 and references therein).

Table 1. Possible Cryomagma Compositions

Aqueous Solution (Hydrate + H ₂ O Phases)	Hydrated Mineral Wt. %	Density of Solution, ρ (kg/m ³)	Eutectic Temperature (K)
Water	N/A	1000	273
Mirabilite	4 wt.% Na ₂ SO ₄ -10H ₂ O	1038	272
Meridianiite	17 wt.% MgSO ₄ -11H ₂ O	1177	269.55
Epsomite	20 wt.% MgSO ₄ -7H ₂ O	1226	268
Bloedite	Na ₂ Mg(SO ₄) ₂ -4H ₂ O: 16 wt.% MgSO ₄ ; 3 wt.% Na ₂ SO ₄	1190	268
Potassium Chloride Hydrate	19.9 wt% KCl-nH ₂ O	~1132	262
Hydrohalite	23 wt% NaCl-2H ₂ O	~ 1200	252
Magnesium Chloride Hydrate	21 wt% MgCl ₂ -nH ₂ O	~1200	239.4
Calcium Chloride Hexahydrate	30.2 wt% CaCl ₂ -6H ₂ O	1282	223.2
Sulfuric Acid Hydrate	35.6 wt.% H ₂ SO ₄ -6.5H ₂ O	1283	211
Sulfuric Acid Hydrate	37 wt.% H ₂ SO ₄ -4H ₂ O	1290	198

These data have been extracted from Kargel (1991), McCarthy et al. (2007), Green and Perry (2007), and Brady (2009).

1.3 Modes of Cryomagma Ascent

Akin to magmatic ascent on Earth, the principal modes of magma transport considered here for Europa include diapiric ascent, dike propagation, and ascent in vertical, pipe-like conduits (e.g., Marsh, 1978; Marsh and Kantha, 1978; Marsh, 1982; 1984). These mechanisms have also been suggested for fluid transport on Europa by various authors (e.g., Crawford and Stevenson, 1988; Pappalardo et al., 1998; Rathbun et

1 al., 1998; Wilson and Head, 1997, 1998; Head and Pappalardo, 1999; Collins et al., 2000;
2 Figueredo et al., 2002; Sotin et al., 2002; Fagents, 2003; Pappalardo and Barr, 2004,
3 among others).

4 Diapirism may be initiated when a pocket of ice in a 250 K layer near the base of
5 the ice shell becomes tidally heated (Sotin et al., 2002) to a temperature near 269 K. Once
6 heated, thermal buoyancy would cause the diapir to continue to rise (Rathbun et al.,
7 1998). Diapirs may also be generated by thermal convection in a ductile sublayer of ice
8 (McKinnon, 1999; Sotin et al., 2002; Mitri and Showman, 2005). In this case, buoyant
9 thermals would repeatedly rise from the base of the sublayer until steady Rayleigh-
10 Bernard convection is established (Marsh, 1989).

11 Dikes may be initiated when tidal effects generate stresses in Europa's ice shell
12 that open fluid-filled fractures. These fractures may extend from the underlying ocean
13 and would subsequently propagate upwards through the ice shell as they attempt to close
14 (Crawford and Stevenson, 1988; Matson et al., 2012). Dike propagation may also ensue
15 when crustal stresses acting on discrete fluid reservoirs or, excess pressures caused by
16 partial freezing of these reservoirs, generate fractures that propagate to the surface
17 (Fagents, 2003; Manga and Wang, 2007; Craft et al., 2013; 2016).

18 Pipe-like conduits are likely to be formed when rapidly ascending cryomagmas
19 melt and/or remove dike walls (Wylie et al., 1999; Carrigan, 2000). They may also form
20 when the distal, thinner ends of dikes rapidly seal off (cf. Daniels et al., 2012). This
21 sealing will cause melt flow to become increasingly concentrated into restricted regions
22 where it will become faster, more voluminous, and will be better able to stave off rapid
23 solidification (Petcovic and Dufek, 2005 and references therein; Wylie et al., 1999).

1 To a large degree, the cooling rates of each of these transport mechanisms reflect
2 their surface area to volume ratios (Marsh, 1978; 1982). Diapirs, being roughly spherical,
3 have small surface area to volume ratios and thus cool relatively slowly (Marsh, 1978;
4 1982). Highly elliptical, tabular dikes, on the other hand, having large surface area to
5 volume ratios, cool quickly (Marsh 1978; 1982). Cooling rates for magma flowing in
6 cylindrical conduits is intermediate between the two.

7 **2 Methodology**

8 In sections 2.2-2.4 we present models for cryomagmatic ascent under three unique
9 conditions: the first model investigates heat transfer in cryomagmas that travel from the
10 base of the stagnant lid to the surface as diapirs of warm ice. The second and third models
11 investigate cryomagma ascent in propagating dikes and cylindrical conduits, respectively.
12 Here, we review the processes by which these transport mechanisms may occur, and
13 place bounds on their respective dimensions.

14 Diapirs are likely to warm the adjacent country ice as they rise, causing it to
15 become ductile. In this way they are able to traverse the rigid lid by causing the
16 surrounding ice to “flow” around them. Terrestrial magmatic diapirs are believed to move
17 through Earth’s crust in this way, forming an ascent path by warming and mobilizing the
18 brittle rock around them (Marsh 1978; 1982 and section 4.1.1 for more details). We
19 envision that European diapirs tidally heated to 269 K arrive at the base of the stagnant lid
20 and may subsequently travel to the surface in this way.

21 Fagents (2003) showed that excess pressures caused by freezing of just 1% of
22 fluid in discrete reservoirs 10 km beneath Europa’s surface was enough to drive the
23 eruption of fluids at the surface. Accordingly, it has been assumed that dikes and pipe-

1 like conduits deliver fluids from isolated reservoirs at these depths to the surface.
2 Because the ascent of cryomagmatic solutions in dikes and cylindrical conduits is
3 pressure driven, buoyancy is not a problem. Hence cryolava delivered to the surface via
4 these mechanisms may be liquid water (Fagents, 2003), aqueous solutions containing
5 chloride or sulfate salts or mineral acids (Kargel, 1991; Hogenboom et al., 1995; Spaun
6 and Head, 2001; Muñoz-Iglesias et al., 2013; 2014), or slushy mixtures of water and ice
7 (Fagents, 2003; Miyamoto et al., 2005). The models presented here assume that
8 cryomagma delivered to Europa's surface via tabular dikes and pipe-like conduits is in
9 the form of liquid water initially at 273 K.

10 Regardless of ascent mechanism, the significant factor in the present analysis is
11 not the exact means of sequestering cryomagmatic fluids, but rather that cryomagmatic
12 fluids very likely exist and are capable of being transported via the aforementioned
13 mechanisms. In this respect, the sizes of expected diapirs, dikes, and cylindrical conduits
14 are fundamentally important to estimate, as all results of the overall solidification time,
15 and thus transfer rates, depend critically on these measures.

16 In terms of determining the volume of magma involved in a typical cryovolcanic
17 event, estimates have been made based on the sizes of putative cryovolcanic features that
18 have been observed on the surface (e.g., see Figs. 1 & 3). This is similar to the scheme
19 that is followed for terrestrial magmatism where the volumes of lavas, volcanoes, and
20 plutons place strong constraints on the modes and volumes of magma transport (e.g., see
21 Marsh, 1978; 2007; 2013; Crisp, 1984; Wylie et al., 1999; White et al., 2006; Annen,
22 2011; Barboni et al., 2015). According to Crisp (1984) and White et al. (2006), the ratio
23 of intruded to erupted magma in terrestrial systems is, on average, 5:1. Further, Europa's

1 surface does not exhibit evidence of widespread cryovolcanism (Phillips et al., 2000;
2 Fagents, 2003; Miyamoto et al., 2005; Quick et al., 2013). For these reasons, we envision
3 that only a portion of available cryomagma in Europa's shallow subsurface will be
4 erupted during any cryovolcanic event. Hence, we assume that cryovolcanic features are
5 likely to provide minimum volume estimates.

6 Rathbun et al. (1998) suggested that diapirs responsible for the creation of domal
7 lenticulae on Europa (Fig. 3) may have had radii less than or equal to a few kilometers,
8 while Figueredo et al. (2002) suggest that Murias Chaos (Fig. 1d) may have been formed
9 when a 5-8 km diapir extruded and flowed out onto the surface. Previous workers have
10 asserted that lenticulae generally have diameters on the order of 10 km (Spaun et 1998;
11 1999; 2001; Pappalardo et al., 1998; Rathbun et al., 1998). Therefore, we assume
12 spherical diapirs that may have produced features on Europa's surface have radii between
13 1 and 5 km. These dimensions are similar to those estimated for plutons in terrestrial
14 settings (Fyfe, 1970; Muroi, 1973; Marsh, 1978; 2000; Farina et al., 2010; Bungler and
15 Cruden, 2011; Nabelek et al., 2012; Marsh, 2013).

16 Wilson et al. (1997) considered dike widths between 1 and 30 m on Europa,
17 concluding that if putative flow lobes associated with Thrace Macula were
18 cryovolcanically emplaced, they may have been produced by cryomagmas ascending to
19 the surface in dikes 5-10 m wide. Recent work has shown that fractures capable of
20 bringing cryolavas to the surface of icy moons such as Enceladus may be at least 1m
21 wide (Neveu et al., 2015), while the width of an active fissure on the small moon has
22 been estimated to be 9 m (Goguen et al., 2009). Fagents (2003) considered conduits 1,10,
23 and 100 m wide capable of bringing cryomagmatic effusions to Europa's surface. She

also noted that European fractures, which could transport cryomagmas to the surface, may be 50-80 m wide (Fagents, 2003). We note however, that it may not be mechanically or thermodynamically possible for such large fractures and conduits to be maintained on Europa.

Based on this information, and to keep the models conservative, we take $r_d = 2$ m. Since conduits are likely to be produced when rapidly ascending cryomagmas melt and/or remove dike walls (Carrigan, 2000), it is expected that these cylindrical passageways will be several times wider than the eroded dikes/elastic cracks they were produced from. Therefore, we take $r_c = 6$ m as the radius of the cylindrical conduit in the models presented here. All cryomagmatic volumes considered are assumed to be reasonable based on the volumes of potential eruptive features they may have produced on Europa's surface.

Figure 3. Domal lenticulae on Europa. In both images, illumination is from the right.

2.1 Consistent Model Parameters

The model assumptions discussed in this section have been applied to each form of cryomagmatic ascent investigated.

Based on the work of Quick and Marsh (2015), we assume that the ice shell/ocean interface on Europa is ~ 30 km below the surface, which is in agreement with previous

1 work suggesting an ice shell ≥ 20 km thick (Ojakangas and Stevenson, 1989; Rathbun et
2 al., 1998; Figueredo et al., 2002; Sotin et al., 2002; Schenk, 2002; Hussmann et al.,
3 2002). A 30 km thick ice shell is likely to convect (Pappalardo et al., 1998; McKinnon,
4 1999; Barr and Pappalardo, 2003; Tobie et al., 2003; Barr and Showman, 2009).
5 Cryomagma transport in all three cases is therefore explored in the presence of a
6 convecting sublayer of warm ice beneath a stagnant conducting lid (Fig. 4). Sotin et al.
7 (2002) suggested that for an approximately 35 km ice shell, the stagnant lid may be on
8 the order of ~ 10 km thick, while Tobie et al. (2003) suggested that the stagnant lid may
9 be at least 35% of the total thickness of a 20-25 km ice shell. We have therefore assumed
10 a 30 km thick ice shell with a 10 km thick stagnant lid, and convective sublayer 20 km
11 thick (Fig. 4).

12 Pappalardo et al. (1998) suggested that the temperature at the top of a convecting
13 sublayer in Europa would be approximately 197 K, while the lower boundary
14 temperature would correspond to the ocean temperature of 273 K (Fig. 4). Once again,
15 the surface temperature is taken to be 100 K. Hence the temperature gradient in the
16 stagnant lid would be approximately 9.7 K/km. Therefore, the temperature $T(z)$ at some
17 distance above the base of the stagnant conducting lid is of the form $T(z) = T_o - mz$, where
18 T_o is the initial temperature of the cryomagma, $m = 9.7$ K/km is the temperature gradient,
19 and z is the distance above the base of the stagnant lid, which is situated at $z_o = 10$ km
20 (Fig. 4). For simplicity, our models assume all cryomagmas ascend at a constant rate. We
21 also anticipate that cryomagmatic fluids which reach the surface at 200 K or more are
22 able to incite cryovolcanism (Miyamoto et al., 2005; Goguen et al., 2013). 200 K is
23 therefore taken as the minimum temperature for which molten cryomagmas successfully

1 arrive at Europa's surface. In the case of cryomagmas undergoing pressure-generated
2 transport via dike propagation and conduit ascent, this rather low temperature for aqueous
3 solutions contains the possibility of the cryomagma being a briny solution.

4 It should be noted that in each case, a constant thermal conductivity of 3.3 W/m-K
5 was assumed for the ice shell (Barr and Showman, 2009). The effects of possible
6 compositional disparities in the ice shell were not taken into consideration because our
7 current knowledge of Europa's crustal compositional stratigraphy is too limited to
8 include such effects. For simplicity, the effects of cryomagma solidification at the walls
9 of dikes and cylindrical conduits (e.g., see Wylie et al., 1999; Carrigan, 2000; Daniels et
10 al., 2012) have also not been taken into consideration. In addition, we do not investigate
11 the effects that volatiles may have in delivering cryomagmatic fluids to the surface
12 (Crawford and Stevenson, 1988; Fagents et al., 2000; Muñoz-Iglesias et al., 2014).
13 Notwithstanding, given the limits of our understanding of cryomagmatic processes on icy
14 worlds, and on Europa's subsurface structure, the simplifications made here do not
15 prevent our models from placing useful bounds on heat transfer associated with
16 cryomagmatic ascent. Further should these effects become known, they could easily be
17 incorporated into the models presented here.

Figure 4. The structure of Europa's shell used in the models presented here. The shell consists of an approximately 10 km thick stagnant conducting lid atop a 20 km thick convecting sublayer. Adapted from McKinnon, 1999 with temperature values from Pappalardo et al., 1998.

2.2 Diapiric Ascent

Previous diapir models applied to the formation of domes and flow-like features on Europa assumed that ascent occurred in an isothermal medium with a constant background temperature (Rathbun et al., 1998; Figueredo et al., 2002). Conversely here, the general formulation of Marsh (1978; 1984) is followed for heat transfer in diapirs.

The total heat flux from an ascending magma body, Q_T , relative to the conductive heat flux, Q_C , for the same magma body at rest, is defined as the Nusselt number, Nu :

$$Nu = \frac{Q_T}{Q_C} \quad (1)$$

From (1) it is clear that $Q_T = Nu * Q_C$, and by Fourier's Law:

$$Q_c = \frac{Ak_c}{L} [T - T_m(t)] \quad (2)$$

is the conductive heat flux for a magma body with surface area, A , characteristic length scale, L (e.g., here L equals the radius, r_s , of a spherical diapir), and average temperature, T , located in a medium whose temperature far from the body is $T_m(z) = T_m(t)$. In (2), we are considering the heat flux for a magma body at rest, so that k_c is the thermal

1 conductivity of the surrounding wall ice. Combining (1) and (2), the total heat flux of an
 2 ascending magma body may be expressed as

$$3 \quad Q_T = Nu \frac{Ak_c}{L} [T - T_m(t)] \quad (3)$$

4 The thermal energy, E , of the magma body is

$$5 \quad E = \rho_m c_{p_m} VT \quad (4)$$

6 where ρ_m , and c_{p_m} are the density and specific heat of the magmatic fluid, V is the
 7 volume of the magma body, and T is the same as in (2). By energy conservation

$$8 \quad \frac{dE}{dt} = -Q_T \quad (5)$$

9 Letting $J = Nu \left(\frac{k_c}{\rho_m c_{p_m}} \right) \left(\frac{A}{LV} \right)$, (5) may be rewritten as

$$10 \quad \frac{dT}{dt} + JT = JT_m(t) \quad (6)$$

11 Multiplying (6) by e^{Jt} yields $\frac{dT}{dt} e^{Jt} + JTe^{Jt} = JT_m(t)e^{Jt}$ which may be alternatively

12 expressed as

$$13 \quad \frac{d}{dt} (Te^{Jt}) = JT_m(t)e^{Jt} \quad (7)$$

14 According to the expression for the geotherm, $T_m(t) = T_m(z) = T_o - mz$. However, $z = v_o t$
 15 where v_o is ascent velocity, taken to be constant. The geotherm may therefore be
 16 expressed as $T_m(t) = T_m(z) = T_o - mv_o t$. (7) now becomes

$$17 \quad \frac{d}{dt} (Te^{Jt}) = JT_o e^{Jt} - mv_o e^{Jt} \quad (8)$$

18 which may be rewritten as

$$19 \quad [Te^{Jt}]_o^t = T_o \int_o^t e^{Jt} J dt - \frac{mv_o}{J} \int_o^t J t e^{Jt} dt \quad (9)$$

20 Integration of (9) yields

$$T(t) = T_o - mv_o t + \frac{mv_o}{J} [1 - e^{-Jt}] \quad (10)$$

As previously noted, $v_o t = z$, and the constant ascent velocity constraint means that $v_o = v$. Thus, $1 = v/v_o = (z/t)/(z_o/t_o)$, and $t/t_o = z/z_o$ so that $t = t_o z/z_o$. The temperature distribution in an ascending diapir can therefore be expressed as

$$T(t) = T_o - mz + mz \left[\frac{1 - e^{-Jt_o z/z_o}}{Jt_o z/z_o} \right] \quad (11)$$

where $T_o = 269$ K is the diapir's initial temperature at base of the stagnant lid, $m = 9.7$ K/km, and $z_o = 10$ km. Here $z = 0$ refers to the base of the stagnant lid, and z is any height above the base so that $z = z_o$ at Europa's surface (Fig. 4). Moreover,

$$Jt_o = Nu \left(\frac{k_c}{\rho_m c_{p_m}} \right) \left(\frac{A}{r_s V} \right) \frac{z_o}{v_o} \quad (12)$$

is the non-dimensional parameter controlling the cryomagma cooling rate. In (12), t_o , the total ascent time of the diapir, has been expressed as z_o/v_o . Nu is, again, the Nusselt number, or the ratio of convective to conductive heat transfer in the diapir. For a liquid

sphere, $Nu = 0.46 \sqrt{\frac{v_o r_s}{\kappa_m}} = 0.46 (Pe)^{1/2}$ (Levich, 1962, p.408; Marsh, 1982). $Pe = v_o L/\kappa_m$ is

the Péclet number, which is similar to Nu in that it too may be thought of as the ratio of convective to conductive heat loss from the cryomagma as it ascends through the crust.

With a 100 K surface temperature and a 197 K bottom boundary temperature, the average temperature of ice in the conducting lid is approximately 149 K. Utilizing equation (15) of Barr and Showman (2009) with 10^{14} Pa s as the dynamic viscosity of ice at 273 K returns an average viscosity of 2.4×10^{23} Pa s for the conducting lid; a warm ice diapir at 269 K has a viscosity of 1.5×10^{14} Pa s. According to Marsh (1982) and Weinberg and Podlachikov (1995), the extreme viscosity contrast between the diapir and surrounding country ice allows for the utilization of Nu for a liquid sphere here. Note also that for the

1 warm ice diapir at 269 K, $k_m = 2.3$ W/m-K (equation 5.6 of Hobbs, 1974), $c_{pm} = 1944$
 2 J/kg-K (Fig. 5.7 of Hobbs, 1974), and $\rho_m = 918$ kg/m³, in agreement with the density
 3 value for warm ice provided in Hobbs (1974) and Fagents (2003). Thermal diffusivity of
 4 the cryomagma, κ_m , is therefore taken to be 1.3×10^{-6} m²/sec.

5 The rate of heat lost by diapirs as they move through the lithosphere can be
 6 represented by various values of the dimensionless parameter Jt_o (Marsh and Kantha,
 7 1978; Marsh, 1978;1982), where J is a constant proportional to the rate of heat lost by
 8 convection and conduction as diapirs ascend, and t_o , again, represents total ascent time to
 9 the surface. From (12) it is clear that small values of Jt_o correspond to short total ascent
 10 times, high ascent rates, and small quantities of heat loss en route to the surface.

11 By assigning a range of values for Jt_o , the average temperature of diapirs moving
 12 through the crust can be found without explicitly specifying the size of the diapir under
 13 consideration. Diapir ascent velocity has been solved for as follows. Recall that $Nu =$

14 $0.46 \sqrt{\frac{v_o r_s}{\kappa_m}}$, and from (12), $Jt_o = Nu \left(\frac{k_c}{\rho_m c_{pm}} \right) \left(\frac{A}{r_s V} \right) \frac{z_o}{v_o}$. Combining these expressions
 15 yields

$$16 \quad Jt_o = 0.46 \sqrt{\frac{v_o r_s}{\kappa_m}} \left(\frac{k_c}{\rho_m c_{pm}} \right) \left(\frac{A}{r_s V} \right) \frac{z_o}{v_o} \quad (13)$$

17 where all constants are the same as before. Due to the constant ascent velocity
 18 assumption, $v = v_o$, so that (13) becomes

$$19 \quad Jt_o = 0.46 \sqrt{\frac{v r_s}{\kappa_m}} \left(\frac{k_c}{\rho_m c_{pm}} \right) \left(\frac{3}{r_s^2} \right) \frac{z_o}{v} \quad (14)$$

20 which, after combining like terms, is

$$21 \quad Jt_o = 1.38 \left(\frac{k_c}{\rho_m c_{pm}} \right) \frac{z_o}{\sqrt{v \kappa_m r_s^3}} \quad (15)$$

1 Squaring each side of (15) and solving for v leads to

$$2 \quad v = 1.9 \left(\frac{k_c}{\rho_m c_{p_m}} \right)^2 \frac{z_o^2}{(Jt_o)^2 \kappa_m r_s^3} \quad (16)$$

3 Inserting $k_c = 3.3$ W/m-K, $\rho_m = 918$ kg/m³, $c_{p_m} = 1944$ J/kg-K, $\kappa_m = 1.3 \times 10^{-6}$ m²/s and z_o
 4 = 10 km into (16), returns

$$5 \quad v \cong \frac{500}{r_s^3 (Jt_o)^2} m^4/s \approx 5 \left(\frac{r_s}{0.215m} \right)^{-3} (Jt_o)^{-2} m s^{-1} \quad (17)$$

6 as an approximate expression for diapir ascent velocity. We reiterate here that Jt_o is an
 7 adjustable parameter that is to be set. (17) illustrates that the velocities obtained from our
 8 calculations correspond to a chosen value of Jt_o . Note that from (14) Jt_o is simply equal to

9 $Nu \left(\frac{3k_c}{r_s^2 \rho_m c_{p_m}} \right) \frac{z_o}{v}$. It is clear that $Nu \left(\frac{3k_c}{r_s^2 \rho_m c_{p_m}} \right)$ represents the heat transfer timescale,

10 t_{trans} , while $\frac{z_o}{v}$ is total ascent time, t_{asc} . Thus (14) may be alternatively expressed as

$$11 \quad Jt_o = t_{asc} / t_{trans} \quad (18)$$

12 **2.3 Cryomagmatic Ascent in Tabular Dikes**

13 As with terrestrial magmas, the Prandtl number ($Pr = \nu/\kappa$) of cryomagmas in
 14 propagating dikes is expected to be large enough that a steady, fully developed velocity
 15 profile may be assumed upon entrance into to the dike (Marsh, 1978). Owing to Europa's
 16 geothermal gradient, however, the temperature of the ice shell continuously changes with
 17 distance from the base of the stagnant lid. It is therefore unlikely that the temperature
 18 profile will ever be fully developed. Since the temperature profile in the dike will never
 19 become invariant with respect to position, the entire length of the dike is similar to an
 20 extended thermal entrance region. It is therefore expected that the value of the Nusselt
 21 number for the cryomagma continuously changes with distance from the base of the

1 stagnant lid (Marsh, 1978; Chapter 4 of Mills, 1999). Accordingly, cryomagma
 2 ascending in tabular dikes from subsurface fluid reservoirs can be effectively modeled as
 3 fluid flow in vertical, flat ducts with varying wall temperatures. Following the
 4 formulation in Sellars et al. (1955) and chapters 8 and 9 of Kays (1966) and Kays and
 5 Crawford (1993), respectively, an expression for mean magma temperature in
 6 propagating dikes can be derived (Quick, 2013), namely

$$7 \quad T = T_o + 2mz^+ + \frac{9}{8}m \sum \frac{G_n}{\lambda_n^4} - \frac{9}{8}m \sum \frac{G_n}{\lambda_n^4} \exp\left(-\lambda_n^2 \frac{8}{3}z^+\right) \quad (19)$$

8 Here $T_o = 273$ K, $m = 9.7$ K/km and $z^+ = \frac{z}{r_d Pe}$ is the non-dimensional distance above the
 9 base of the stagnant lid. Heat loss is once again controlled by the Péclet number, now

10 expressed as $Pe = \frac{vr_d}{\kappa}$, where r_d is dike half-width. Details leading to the derivation of

11 (19) may be found in chapter 8 of Kays (1966) and chapter 7 of Rohsenow and Hartnett
 12 (1973) and Quick (2013). In accordance with Fagents (2003), we take $\kappa = 5 \times 10^{-7}$ m²/s
 13 for pure water at 273 K.

14 Large Pe means that fluid ascends at a high rate, resulting in a short ascent time
 15 and a relatively small amount of heat loss. G_n and λ_n represent, respectively, numbers
 16 resulting from eigenfunctions, and eigenvalues, found from the solution of a Sturm-
 17 Liouville-type equation associated with flow in a flat duct. Values for λ_n^2 are listed in
 18 Table II of Sellars et al. (1955), while values for G_n in the flat duct geometry are taken
 19 from Brown (1960). For completeness when $n = 0, 1, 2, 3$, $G_n = 1.707, 1.135, 0.95$ and
 20 0.845 (Brown, 1960) and corresponding $\lambda_n^2 = 2.779, 32.11, 93.45, 186.9$ (Sellars et al.,
 21 1955). Here, twice the total width of the dike has been taken as the hydraulic diameter

(Chapter 9 of Kays and Crawford, 1993). Owing to their large surface area to volume ratios, in comparison to diapirs, dikes must propagate at very high rates in order to arrive at the surface before solidification.

2.4 Cryomagma Ascent in Pipe-Like Conduits

As previously mentioned, cylindrical conduits are usually generated by processes associated with diapirism and diking. Here they are presented as secondary ascent mechanisms formed via the continued passage of high velocity cryomagmas that erode dike walls (Wylie et al., 1999; Carrigan, 2000).

Cryomagma ascending in pipe-like conduits is analogous to fluid undergoing Poiseuille flow in cylindrical pipes where wall temperatures change continuously with distance from the source. Hence mean magma temperature as a function of distance from the top of fluid reservoir is represented by the following equation, which describes the temperature of fluid flowing in a round tube with linearly varying wall temperature

$$T = T_o + mz^+ - 8m \sum \frac{G_n}{\lambda_n^4} + 8m \sum \frac{G_n}{\lambda_n^4} \exp(-\lambda_n^2 z^+) \quad (18)$$

(Sellars, et al., 1955; Chapter 8 of Kays, 1966; Chapter 9 of Kays and Crawford, 1993;

Quick, 2013). Here T_o , m , and z are the same as in section 2.3, and $z^+ = \frac{z}{r_c Pe}$. Overall

heat loss is again controlled by Pe , which may be expressed here as $Pe = \frac{vr_c}{\kappa}$, where r_c is

conduit radius and $\kappa = 5 \times 10^{-7} \text{ m}^2/\text{s}$. As was the case for dikes, large Pe means that

magma ascends quickly through the conduit with a short ascent time and a relatively

small amount of heat loss. Values for G_n and λ_n^2 for a cylindrical geometry are listed

1 in Table 8-3 of Kays (1966). For completeness, when $n = 0, 1, 2, 3$, $G_n = 0.749, 0.544,$
2 0.463 and 0.414 , and $\lambda_n^2 = 7.312, 44.62, 113.8, 215.2$.

3 **3 Results**

4 **Figures 5a-c** show cooling curves for cryomagmas traversing Europa's ice shell.

5 Depending on ascent rate, cryomagmatic diapirs may reach the surface at various
6 temperatures (Fig. **5a and Table 2**). In accordance with (11), diapirs rising through the
7 stagnant lid with $T_o = 269$ K may reach the surface at temperatures as high as ~ 264 K,
8 and as low as 186 K for Jt_o between 0.1 and 7, respectively. However, only diapirs with
9 $Jt_o \leq 3$ will be able to reach the surface at temperatures greater than 200 K (Fig. 5a).

10 From (17), this corresponds to ascent velocities greater than 5.6×10^{-8} m/s for 1 km
11 radius diapirs, and ascent velocities greater than 4.4×10^{-10} for 5 km radius diapirs. Note
12 that for all Jt_o values considered here, diapirs will arrive within 3 km of the surface at
13 approximately 200 K (Fig. 5 and Table 2). It is worth noting that if diapirs incorporate
14 low-eutectic impurities as they travel through the ice shell (e.g., see Head and
15 Pappalardo, 1999; Schmidt et al., 2011), these species will serve as antifreezes that allow
16 them to travel more slowly **without solidifying**. Further, according to (11) and (17), for
17 any given Jt_o , larger diapirs may ascend several orders of magnitude slower than their
18 smaller counterparts, yet still arrive at the surface at equivalent temperatures. **This is**
19 **because low surface area to volume ratios help large diapirs hold on to heat more**
20 **efficiently than smaller ones**. Table 2 contains ascent rates corresponding to 1 and 5 km
21 radius diapirs for each Jt_o value investigated. Diapir temperatures halfway to the surface,
22 and at the surface, are also included for each Jt_o value.

Cooling curves for cryomagmas that originate in fluid reservoirs near the base of the stagnant lid and travel to the surface in propagating dikes or cylindrical conduits are displayed in Figs. 5b & c. According to (19), Pe must be close to 700 for cryomagmas ascending in 4 m wide dikes to reach even half way to the surface at temperatures ≥ 200 K (Fig. 5b and Table 3). For $Pe = 1000$, fluids with temperatures greater than 200 K will arrive within 3 km of the surface (Fig 5b). According to the Péclet number relation, corresponding ascent velocities are 2.5×10^{-4} m/s. Dikes must attain $Pe = 1300$ and propagation rates of at least 3.4×10^{-4} m/s for the fluids within to be delivered to the surface at 200 K (Fig. 5b). Conversely, Pe and v must be very large, on the order of 10^5 , and 2.6×10^{-2} m/s, respectively, for fluids traveling in 4 m wide dikes to obtain isothermal ascent. Such high Pe corresponds to flow in the turbulent regime for pure water and briny aqueous solutions. Table 3 contains propagation and corresponding fluid temperatures 50% of the way to the surface, and at the surface, for each Pe investigated.

Figure 5c shows that fluids ascending in 12 m wide pipe-like conduits must attain $Pe = 200$ to reach the very shallow subsurface at temperatures greater than 200 K. For $Pe = 300$ and $v = 2.5 \times 10^{-5}$ m/s, fluids will be delivered to the surface at temperatures around 220 K (Fig. 5c and Table 4). If $Pe = 10^4$, and an ascent rate of 8.3×10^{-4} m/s can be attained, fluids travelling in cylindrical conduits will ascend nearly isothermally. Note that under the conditions explored here, isothermal ascent in a pipe-like conduit requires Pe and v about an order of magnitude less than for isothermal ascent in a tabular dike. Table 4 contains ascent rates and corresponding fluid temperatures 50% of the way to the surface, and at the surface, for each Pe investigated.

Figure 5a. Cooling curves for diapirs initially at 269 K, travelling from the base of the stagnant lid to the surface. In accordance with (11) and (17), the smallest J_{to} values correspond to the quickest ascent rates and highest temperatures upon reaching the surface. The grey line represents the 200 K temperature mark required in our models for cryovolcanism to be incited at the surface by the successful ascent of cryomagmas to the surface. Also see Table 2.

Figure 5b. Cooling curves for fluids initially at 273 K that originate near the base of the stagnant lid and travel to the surface in tabular dikes 4 m wide. Here, large Pe correspond to a short total ascent time and highest temperatures upon reaching the surface. Fluids reach the surface at 200 K for $Pe = 1300$. The grey line represents the 200 K temperature mark required for successful ascent of cryomagmas to the surface. Also see Table 3.

Figure 5c. Cooling curves for fluids originally at 273 K that travel to the surface in cylindrical conduits 12 m wide. As is the case for dikes, large Pe corresponds to high ascent rates, short total ascent time, and high fluid temperatures upon reaching the surface. Fluids reach 90% of the way to the surface at temperatures greater than 200 K for $Pe = 200$. The grey line represents the 200 K temperature mark required for successful ascent of cryomagmas to the surface. Also see Table 4.

Table 2. Approximate temperatures and ascent rates for diapirs for each Jt_0 value investigated. Diapir temperatures 50% of the way to the surface, and at the surface, are displayed in the second and third columns.

			$r = 1 \text{ km}$	$r = 5 \text{ km}$
Jt_0	$T \text{ (K) at } z = 5 \text{ km}$	$T \text{ (K) at } z = 10 \text{ km}$	$v \text{ (m/s)}$	$v \text{ (m/s)}$
0.1	268	264	5×10^{-5}	4×10^{-7}
0.2	267	260	1.3×10^{-5}	1×10^{-7}
0.3	266	256	5.6×10^{-6}	4.4×10^{-8}
0.4	264	252	3.3×10^{-6}	2.5×10^{-8}
0.5	263	248	2×10^{-6}	1.6×10^{-8}
0.6	262	245	1.4×10^{-6}	1.1×10^{-8}
0.7	261	242	1×10^{-6}	8.2×10^{-9}
0.8	260	239	7.8×10^{-7}	6.3×10^{-9}
0.9	260	236	6.2×10^{-7}	5×10^{-9}
1	259	233	5×10^{-7}	4×10^{-9}
3	246	203	5.6×10^{-8}	4.4×10^{-10}
5	238	191	2×10^{-8}	1.6×10^{-10}
7	234	186	1×10^{-8}	8.2×10^{-11}
9	231	183	6.2×10^{-9}	4.9×10^{-11}
11	229	181	4.1×10^{-9}	3.3×10^{-11}
13	228	179	3×10^{-9}	2.4×10^{-11}
15	227	178	2.2×10^{-9}	1.8×10^{-11}
20	225	177	1.3×10^{-9}	1×10^{-11}

Table 3. Approximate temperatures and ascent speeds for fluids traveling in 4 m wide propagating dikes. Initial temperatures of these fluids were 273 K. Propagation speeds were calculated using the Péclet number relation: $Pe = v r_d / \kappa$, where $\kappa = 5 \times 10^{-7} \text{ m}^2/\text{s}$ (Fagents, 2003).

Pe	$v \text{ (m/s)}$	$T \text{ (K) at } z = 5 \text{ km}$	$T \text{ (K) at } z = 10 \text{ km}$
500	1.3×10^{-4}	178	81
700	1.8×10^{-4}	206	137
1000	2.6×10^{-4}	227	178
1300	3.4×10^{-4}	238	201
2000	5.2×10^{-4}	251	227
3000	7.8×10^{-4}	259	243
5000	1.3×10^{-3}	266	256
7000	1.8×10^{-3}	268	262
2×10^4	5.2×10^{-3}	272	270
1×10^5	2.6×10^{-2}	273	273

Table 4. Approximate temperatures and ascent speeds for fluids traveling in 12 m wide cylindrical conduits. Initial temperatures of these fluids were 273 K. Propagation speeds were calculated using the Péclet number relation: $Pe = v r_c / \kappa$, where $\kappa = 5 \times 10^{-7} \text{ m}^2/\text{s}$ (Fagents, 2003).

Pe	$v \text{ (m/s)}$	$T \text{ (K) at } z = 5 \text{ km}$	$T \text{ (K) at } z = 10 \text{ km}$
100	8.3×10^{-6}	193	112
200	1.7×10^{-5}	234	193
300	2.5×10^{-5}	247	220
500	4.2×10^{-5}	258	242
700	5.8×10^{-5}	263	251
1000	8.3×10^{-5}	266	258
3000	2.5×10^{-4}	271	269
5000	4.2×10^{-4}	272	271
1×10^4	8.3×10^{-4}	273	272

4 Discussion

4.1.1 Diapirs

In our models, the “wall rock” is Europa’s crustal ice. In Sections 2 & 3, large Jt_o values signify that the diapir, in effect, is inefficient in warming and softening the surrounding lithospheric ice; as a result, it ascends very slowly and significant amounts of heat are lost to the surrounding ice. Conversely, small Jt_o values signify that the diapir is able to efficiently warm and soften the brittle ice surrounding it, enabling more rapid ascent to the surface and retention of a significant fraction of its original heat (Marsh, 1982). We now turn to estimated ascent rates of terrestrial diapirs to determine if ascent rates calculated in the previous section are reasonable.

In terrestrial settings, the drag experienced by an ascending diapir and its associated velocity field are relatively unaffected by the properties of the diapirs themselves, but rather by viscosity variations in the wall rock (Marsh, 1982; Weinberg and Podladchikov, 1995). According to Grout (1945), in order for magma to reach the surface successfully, the viscosity of the wall rock, and by extension, the drag that it induces on the diapir, must be reduced along the diapir’s periphery. To put it more

1 simply, the velocity of an ascending diapir is determined by its ability to warm and soften
2 wall rock on its way to the surface.

3 As diapirs ascend, they are surrounded by thermal boundary layers. At the
4 external edge of the boundary layer wall rock viscosity is unchanged. However nearest
5 the diapir, the wall rock is heated and softened, and its viscosity substantially decreases
6 (Marsh, 1982; Weinberg and Podladchikov, 1995). This process is analogous to a hot
7 knife slicing through cold butter. According to Marsh (1978; 1982), the softened layer of
8 wall rock encasing diapirs as they travel is at most, 20-30 % of the diapir's radius.

9 Assuming a softened layer 20% of the diapir's radius, this corresponds to adjacent
10 softened layers between 0.2-1 km thick on each side of diapirs 1-5 km in radius.
11 Assuming ice amenable to flow in Europa's subsurface must be warmed to at least 180 K
12 (Barr and Showman, 2009) and applying equation (4-113) of Turcotte and Schubert
13 (2002), we find that it will take approximately 400 years for a diapir at 269 K to warm
14 0.2 km of wall ice from 100 K to 180 K. Conversely, it will take close to 10,000 years to
15 warm 1 km of adjacent wall ice to 180 K. Here we have assumed that $k_c = 3.3$ W/m-K
16 and $\rho_{ice} = 1000$ kg/m³; from Fig. 5.7 of Hobbs (1974), $c_p = 1277$ J/kg-K at 180 K (Fig.
17 5.7 of Hobbs). This returns $\kappa = 3 \times 10^{-6}$ m²/s for use in Turcotte and Schubert's (2002)
18 equation (4-113). We can compare this to the time for softening wall rock adjacent to
19 dikes on Earth. Petcovic and Dufek (2005) have shown that wall rock melting in a 5 m
20 thick layer adjacent to a 10 m wide dike can be initiated in less than 380 days, which
21 reflects the extreme heat carried by my terrestrial magmas.

22 Based on the timescale for conductive cooling ($t_{cool} = r^2 / \kappa$), diapirs with $r_s = 1$ km
23 will take about 3×10^4 years to completely cool, while those with $r_s = 5$ km will take

1 approximately 8×10^5 years to cool. This means that both 1 km and 5 km radius diapirs
2 traversing a 10 km thick stagnant lid could successfully reach the surface. We hasten to
3 add, however, that with diapirs, in addition to the adjacent country ice that must be
4 softened, roof ice ahead of the advancing diapir must also be melted. We have not taken
5 this factor into account here.

6 Akin to terrestrial diapirs, European diapirs may be more likely to reach the surface
7 if they ascend paths that have been warmed and softened by the repeated passage of
8 previous generations of diapirs (Marsh 1978; 1982). Traversing a thermally insulated
9 path along which country ice and roof ice have already been processed may help diapirs
10 much smaller than those considered here reach the surface in a molten state. However, it
11 may take hundreds of thousands of years for these insulated chimneys to be established.
12 Once diapirism has been initiated it is possible that diapirs will ascend at different rates,
13 owing to the relative “warmth” of the lithospheric path they travel. On Earth, it is
14 common for magmas to travel rather quickly to the shallow subsurface and subsequently
15 wait for long periods of time before finally erupting (Marsh, 1978; Marsh and Kantha,
16 1978). Thus, variable ascent velocities are likely, but this involves a much greater
17 knowledge of both the details of the actual dynamics of ascent and the viscoelastic
18 properties of the ice shell than is presently known.

19 On Earth, diapirs ascending through un-preheated lithosphere are likely to travel
20 on the order of 10^{-8} to 10^{-9} m/s, while diapirs ascending paths that have been insulated by
21 the heat of previously ascending diapirs are likely to travel at rates between 10^{-7} and 10^{-6}
22 m/s (Marsh, 1978; 1982). The viscosity of basaltic magma erupting at temperatures near
23 1700 K is about 10 Pa-sec (Spera, 2000), while the viscosity of Earth’s solid mantle is

1 10^{21} Pa-sec (Chapters 1 & 6 of Turcotte and Schubert, 2002). Hence the difference
2 between magma and mantle rock viscosity on Earth is, at most, about 21 orders of
3 magnitude. Assuming a minimum viscosity of 10^{14} Pa-sec for warm ice on Europa,
4 equation (15) of Barr and Showman (2009) gives a viscosity on the order of 10^{33} Pa-sec
5 for lithospheric ice at 100 K. Hence the maximum viscosity difference between warm ice
6 in diapirs, and the coldest part of Europa's lithosphere, is about 19 orders of magnitude.
7 Due to the similar viscosity differences between Earth's mantle rocks and magma and
8 Europa's brittle lithosphere and warm ice, we expect ascent velocities of diapirs of warm
9 ice moving through Europa's lithosphere to be comparable to ascent velocities of
10 terrestrial magma diapirs moving through Earth's mantle.

11 We therefore assume that the maximum ascent rate attainable for diapirs on
12 Europa is 10^{-7} - 10^{-6} m/s. According to (17), this corresponds to Jt_o equal to 2.2 - 0.7 for
13 diapirs with 1 km radii, and Jt_o equal to 0.2 - 0.06 for diapirs with 5 km radii. From (11)
14 and (17), a 1 km radius diapir with $Jt_o = 2.2$ will reach the surface at 211K, while a 1 km
15 radius diapir with $Jt_o = 0.7$ will arrive at the surface at 242 K. Conversely for $Jt_o = 0.2$,
16 diapirs that are 5 km in radius would reach the surface at 260 K, while their counterparts
17 with $Jt_o = 0.06$ would reach the surface at 266 K. Hence if European diapirs were able to
18 attain these ascent rates, cryomagma could be successfully brought to the surface to
19 facilitate cryovolcanism. Extra heating and insulation provided by the use of pre-heated
20 ascent paths may allow diapirs to travel to the surface at these speeds.

21 As mentioned above, up to 1 km of surrounding country ice must be softened to
22 allow a 5 km radius diapir to pass through. Thus in order to traverse a 10 km thick brittle
23 layer of ice, ~ 10 times as much country ice must be processed. This means that it could

1 take more than 10 successive diapirs to preheat a full 10 km ascent path through the
2 stagnant lid. This assumes that there are no nested diapirs and that each successive diapir
3 reaches a point closer to the surface than its progenitor. For the case of the 1 km radius
4 diapir, it could take as many as 50 successive diapirs to preheat the 10 km path. An
5 extensive treatment of the dynamics of warm ice diapirs ascending through an icy
6 lithosphere must be undertaken to further validate these conclusions. Such a treatment is
7 beyond the scope of this paper. However, preliminary calculations suggest that European
8 diapirs on the order of 5-10 km in radius will have ascent rates between 10^{-9} - 10^{-7} m/s if
9 traveling preheated ascent paths to the surface. This corresponds to Jt_o between 2-0.2 for
10 5 km radius diapirs, and temperatures between 214 -260 K upon reaching the surface.
11 Hence, successful cryomagma transport is possible. A more detailed analysis of the
12 dynamics of diapiric ascent on Europa will be the focus of a future manuscript.

13 **4.1.2 Tabular Dikes and Cylindrical Conduits**

14 Dikes having a width of 4m that propagate at 2.6×10^{-2} m/s could bring
15 cryomagmas to the surface under isothermal conditions. Our results indicate that the
16 same is true for 12 m wide cylindrical conduits that are able to attain ascent rates greater
17 than 8.3×10^{-4} m/s. These velocities are reasonable in comparison to those estimated for
18 terrestrial magmas undergoing ascent in dikes and pipe-like conduits (Marsh, 1978).
19 Since excess pressure applied to fluid pockets in the subsurface is able to overcome the
20 negative buoyancy of fluids within, it is possible that ice slurries or briny cryomagmatic
21 solutions mentioned in Section 1.2 may be brought to the surface via diking or ascent in
22 cylindrical conduits. Reddish-brown areas on Europa could therefore represent places

1 where tabular dikes or pipe-like conduits containing briny solutions have intersected with
2 the surface.

3 It must be noted that in terrestrial systems, pipe-like conduits are often parts of
4 intricate magmatic plumbing systems in well-developed, mature volcanic centers (Wright
5 and Klein, 2014). Diapirism and diking are more easily initiated and are commonly
6 observed in deeply eroded terrains. Hence depending on Europa's subsurface structure,
7 diapirism and diking may be the most likely mechanisms to bring cryomagmas to the
8 surface.

9 **5 Conclusions**

10 The areal extent and compositions of cryolavas erupted onto Europa's surface
11 could reveal a great deal about the moon's subsurface structure and evolution Information
12 that could be gleaned from cryolavas at the surface include the depth, composition, and
13 present state of subsurface cryomagma plumbing systems, as well as the dimensions of
14 diapirs, dikes, and pipe-like conduits that may have produced various surface features.
15 Tracking the cycling of cryomagma from the source to the surface could also shed light
16 on whether or not certain surface species such as H_2SO_4 , MgSO_4 , etc., could be
17 endogenically produced or if they are strictly the results of exogenic processes. Further, if
18 cryomagmas ascending as diapirs originate at the ice-shell ocean interface, or if fluid
19 reservoirs in the ice shell are nothing more than stalled diapirs that contain oceanic
20 constituents, sampling of cryolava flows on the surface could reveal details concerning
21 the composition of the subsurface ocean. Bearing in mind the time it takes for them to
22 completely solidify, stalled diapirs could also represent habitable subsurface niches.

1 These niches would be transient in nature; nevertheless, they would still be of
2 astrobiological interest.

3 On Earth, it is expected that diapirs must travel ascent paths that have been used
4 for millions of years by their predecessors in order to arrive at the surface before
5 solidification takes place. The same may be true on Europa. Hence, uplifts that may have
6 been created as a result of diapirism and areas where cryolavas may have **erupted** (Figs. 1
7 & 3) could represent areas of the surface that lie atop hot spots. Cryomagmatic diapirs
8 ascending at rates between 10^{-9} - 10^{-6} m/s through an approximately 10 km thick ice shell
9 will take 10^2 - 10^4 years to arrive at the surface as cryolavas. Cryomagmatic dikes
10 travelling at speeds of 10^{-5} - 10^{-2} m/s would take 12 days to 32 years to travel from 10 km
11 depths to the surface. A 12-day time constraint corresponds to isothermal ascent for the
12 dike geometry investigated here. On Earth, dikes have been inferred to travel at these
13 rates, and in some cases even faster, when accompanying volcanic events. Hence it is
14 possible that under special circumstances, fluids originating from 10 km depths in
15 Europa's subsurface may travel to the surface in dikes under isothermal conditions.

16 Any putative lava flows or cryovolcanic constructs on Europa that were not
17 present during the Galileo mission but are imaged by future Europa missions (e.g.
18 NASA's Europa Mission, or ESA's JUICE Mission) may have been recently brought to
19 the surface via diking or near surface diapirs that had yet to emerge during the Galileo
20 spacecraft's observations. In order to determine the primary means of cryomagma
21 transport on Europa, future missions should look for evidence of migrating hotspots
22 associated with areas of plate recycling and subduction, as well as for evidence of
23 **eruptive** vents at the surface.

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