Cryomagma ascent on Europa

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Abstract

Europa’s surface exhibits morphological features associated with a low craters density that demonstrate a recent internal activity. In particular, the morphology of the smooth plains covering parts of the surface, and their relationship to the surrounding terrains suggest that they result from viscous liquid extrusions. Furthermore, recent literature explains the emplacement of liquid-related features, such as double ridges, lenticulae and chaos by the presence of liquid reservoirs beneath the surface. We model the ascent of liquid water through a dike or a pipe-like conduit, from a sub-surface reservoir to Europa’s surface and derive the eruption time-scale and the total volume extruded at the end of the eruption, depending on the chamber volume and depth. We also estimate the freezing time of the sub-surface reservoir necessary to trigger an eruption. Considering available data for density and eutectic temperature of salt impurities recently proposed for Europa, we discuss their effect on the cryomagma freezing time and ascent. For plausible volumes and depths varying between $0,1 \text{ km}^3 \leq V \leq 10 \text{ km}^3$ and $100 \text{ m} \leq H \leq 10 \text{ km}$, the total extruded cryolava volume ranges from $10^5$ to $10^8 \text{ m}^3$ and the time scale of the eruptions varies from few minutes to few tens of hours. The freezing time-scale of the cryomagma pocket varies with the cryomagma composition: it varies between $10^2$ to $10^3$ years for a pure water cryomagma and from $10^2$ to $10^4$ years for a briny cryomagma.

Keywords: cryovolcanism, icy satellite, chamber, freezing, salt inclusion

1. Introduction

Europa’s surface is relatively young, with an estimated age of approximately 70 My (Zahnle et al., 2003), thus implying active resurfacing processes. Data from the Galileo spacecraft acquired between 1995 and 2001 show a great diversity of geological features on Europa, demonstrating an internal activity of the moon (Fagents, 2003; Greenberg and Geissler, 2002; Kattenhorn and Prockter, 2014). Among those structures, smooth plains and lobate layers cover parts of the surface (see Fig. 1). Their morphologies and relationship to the surrounding terrains, suggest that they result from a viscous liquid extrusion (Miyamoto et al., 2005). Furthermore, the presence of an internal global liquid water ocean in contact with the silicate mantle beneath Europa’s icy crust (Quick and Marsh, 2015) is supported by the decoupling of the surface and the rocky interior (Schenk et al., 2008). This makes Europa a putative candidate to support the development of life forms (Greenberg and Geissler, 2002). The spectroscopic analysis show that Europa’s surface is entirely covered with water ice, mixed with impurities such as salts and sulfates (Dalton, 2007; Ligier et al., 2016).

Two missions: JUICE (ESA) and Europa Clipper (NASA) aiming to study Europa’s surface and subsurface
environment are in preparation. In this context, understanding where and for how long liquid water is produced at the surface should help to predict where the detection of biosignatures is the most likely.

Recent literature explains the emplacement of common geological features at Europa’s surface such as double ridges (Craft et al., 2016; Dambard et al., 2013; Dameron and Burr, 2018), chaos (Schmidt et al., 2011; Greenberg, 1999) and lenticulae (Manga and Michaut, 2017) by the presence of near-surface liquid water reservoirs. Indeed, the morphological studies of these features are in good agreement with models taking into account hot water lenses at shallow depths. On the other hand, Manga and Wang (2007) showed that it seems unlikely that liquid water rises directly from the internal ocean to the surface through a large crack because of the extremely high pressure required for this mechanism to occur. For a 50km thick ice layer, freezing of few kilometers of water in the ocean would induce an overpressure in the ocean from 1 to 10 kPa, which is enough to propagate a dike over the ice crust thickness (Manga and Wang, 2007; Neveu et al., 2015). However, the overpressure necessary to bring ocean water to the surface is higher, around the MPa, and thus cannot be attained considering the ocean freezing mechanism (Manga and Wang, 2007). Due to the density contrast between ice and water those subsurface reservoirs might be unstable and water should percolate within the ice shell to the internal ocean (Kalousová et al., 2014).

In this study, we focus on the effusive water flows generated by the cryovolcanic activity. We follow the eruption mechanism proposed by Fagents (2003): cryolavas are stored in cryomagmatic chambers within Europa’s ice shell that cool and freeze over time generating an overpressure inside the reservoir eventually leading to the surrounding ice rupture. Her study hence demonstrates the feasibility of this mechanism, and for which pressure and depth range liquid water is able to reach the surface. However, it did not take into account water transport dynamics. Recently, Quick and Marsh (2016) calculate a minimum cryomagma velocity of order of few meters per second in order to reach the surface without freezing for a 4 m wide dike. Also, Neveu et al. (2013) recently studied the impurities which could modify the water properties and the cryovolcanic activity on the icy moons.

The goal of the present study is to build a simple first order predictive model, based on basic mechanical and thermophysical considerations, in order to estimate the order of magnitude of the eruption timescale and cryolavas volume erupted at the surface during one effusive cryovolcanic event. This allows us to link the cryovolcanic features at Europa’s surface and the chemical and physical conditions in Europa’s crust.

We consider a cryomagmatic chamber as a spherical cavity within the ice shell, filled with pure/briny liquid water at isostatic pressure. The cryomagma freezes over the time and generates an overpressure in the chamber which fractures its wall when the tensile stress exceeded the tensile strength of the ice, i.e. when the pressure reaches the critical value. The cryomagma then rises at the surface through a crack. Our work allows us to calculate the velocity of this flow and to predict the evolution of the pressure in the cryomagma chamber throughout the eruption, as well as its duration and the volume of cryolavas emitted at the surface. We also estimate the time required to freeze enough cryomagma to generate an eruption. We then discuss the stability of these reservoirs for different liquid composition.
Figure 1: a) Low albedo zones along a double ridge [Quick and Marsh, 2016]; b) Lobate structure covering an older terrain (Image ID: 15E0071); c) Circular smooth plain [Fagents, 2003]; d) Smooth plain flanking a double ridge [Fagents, 2003].
2. Model

We consider here the freezing and eruption of a cryomagma pocket located in Europa’s ice shell. The generation of this kind of reservoir is beyond the scope of this paper, and was studied previously by several authors (Peddinti and McNamara, 2015; Kalousová et al., 2014, 2016; McKinnon, 1999; Mitri and Showman, 2008; Tobie et al., 2003).

In this section, we first introduce our model and the assumptions made. Then, we detail the cryomagma freezing and ascent mechanisms. Finally, we calculate the cryomagma freezing time-scale in order to compare it with (1) the eruption duration and (2) percolation time available from existing literature (e.g. Kalousová et al., 2014).

2.1. Assumptions

Various processes might explain the formation of cryomagmatic reservoirs in the ice shell. For example, Kalousová et al. (2016) showed that the heat generated by tidally activated faults might be sufficient to generate partially molten water lenses in the ice crust. Mitri and Showman (2008) also showed that local-partial melting of the ice is also possible thanks to the tidal heating in the ice shell, which is enhanced in hot convective plumes because of the temperature dependence of the tidal dissipation rate.

Europa’s global internal ocean is most probably salty (Kargel, 1991; Quick and Marsh, 2016; Fagents, 2000; Head and Pappalardo, 1999; Hogenboom et al., 1995; Dalton, 2007), but its precise composition remains unknown. It is speculated that these salts are in majority MgSO$_4$ and Na$_2$SO$_4$ sulfates and in minor quantity sulfates of K, Mn and Ni, and chlorides such as MgCl$_2$ and CaSO$_4$ (Kargel, 1991; Quick and Marsh, 2016; Fagents, 2000; Head and Pappalardo, 1999; Hogenboom et al., 1995; Dalton, 2007). These salts were present in carbonaceous chondrites, the primitive material of icy moons, and have been released in the ocean because of the interactions between the rocky interior and the water layer (Kargel, 1991). Here, we consider first pure liquid water as a reference scenario. Then we quantify the effect of salts in section 4.

We model the chamber as a spherical cavity at isostatic pressure $P_0$. Since there is no geophysical measurement on Europa’s surface, it seems reasonable to start with this simple condition within the framework of this first order model.

Previous studies showed that Europa’s ice shell could have been weakened by tidal effects (Greenberg and Geissler, 2002; Lee et al., 2005; Harada and Kurita, 2006; Quillen et al., 2016) or due to the global cooling of the moon (Nimmo, 2004; Manga and Wang, 2007). Nevertheless, as discussed in the introduction, it seems difficult to drive an eruption for dike more of few tens kilometers in length (Lee et al., 2005; Neveu et al., 2015). Here we focus on the water reservoirs leading to an eruption at the surface and then only consider reservoirs at less than 10 km depth.

We use the same approach as Fagents (2003) summarized in Fig. 2: the cryomagma contained in a chamber cools over the time, and eventually freezes. Due to confinement, and during freezing, the cryomagma exerts an overpressure $\Delta P$ inside the chamber. This overpressure generates a tensile stress on the cavity walls. When this stress overcome the tensile strength of the ice, the walls crack and the fluid can then flow to the surface. The eruption is thus generated and maintained by a pressure gradient between the inside of the
cryomagmatic chamber and the satellite’s surface. The eruption stops when the overpressure is balanced with the weight of the cryomagma column inside the crack.

All notations used in this section are summarized in Table 1.

2.2. Overpressure in a cooling cryomagmatic chamber

As an initial stage, we consider that the total volume of the chamber $V$ is filled with pure liquid water at isostatic pressure $P_0 = \rho_s g H$, where $\rho_s$ is the water ice density, $g$ is the gravity on Europa, and $H$ is the chamber depth under the surface. The chamber cools with time, and we want to estimate the overpressure $\Delta P$ generated when a volume fraction of liquid $n = 1 - V_i/V$ freezes, with $V_i$ being the initial volume occupied by the liquid remaining in a liquid state after freezing (see Fig. 2b). After freezing, the remaining liquid occupies a volume $V_f$ with $V_f < V_i$, corresponding to the pressure increase $\Delta P$ associated to the liquid compression (see Fig. 2b). This overpressure depends on the liquid water compressibility $\chi$:

$$\chi = -\frac{1}{V} \frac{\partial V}{\partial P}$$  \hspace{1cm} (1)

where $V$ is the liquid volume and $P$ the liquid pressure. Equation (1) gives in our case:

$$\Delta P = -\frac{1}{\chi} \ln \left( \frac{V_f}{V_i} \right)$$  \hspace{1cm} (2)

Keeping in mind that the mass is conserved during freezing, $V_i$ and $V_f$ are defined as:

$$V_i(n) = (1 - n) V$$  \hspace{1cm} (3)

$$V_f(n) = \left( 1 - n \frac{\rho_l}{\rho_s} \right) V$$  \hspace{1cm} (4)
where \( \rho_w \) is the pure liquid water density and \( \rho_s \) is the ice density.

Combining Equations (2), (3) and (4), we obtain the necessary fraction of cryomagma \( n \) that has to freeze in order to induce the overpressure \( \Delta P \):  

\[
    n = \frac{\exp (\chi \Delta P) - 1}{\rho_w \exp (\chi \Delta P) - 1}
\]  

(5)

Note that the density \( \rho_{lf} \) of the liquid contained in the chamber after freezing is given by:  

\[
    \rho_{lf} = \rho_{li} \frac{V_i(n)}{V_f(n)}
\]  

(6)

where \( \rho_{li} \) is the liquid density before being compressed.

One has to note that the pressure never reaches values of order of GPa, so that high pressure ice phase can be ignored (see the water phase diagram in Sanz et al. [2004] readapted from Petrenko and Whitworth [2002]). Also here, we consider a constant compressibility for pure water. A value of \( \chi = 5 \times 10^{-11} \text{ Pa}^{-1} \) is in agreement with the work of Fine and Millero [1973] for pressure of order of a few MPa.

2.3. Tensile failure of a cooling cryomagmatic chamber

In this model, we assume that ice will break in a tensile mode [McLeod and Tait, 1999]. The maximum pressure that could be achieved in the chamber is commended by the ice tensile strength. Litwin et al. [2012] measured the water ice tensile strength depending on its temperature, grain size and composition. Their measurements were made under Titan conditions, with polycrystalline ice. On Europa, we only know the ice structure of the first millimeter of the surface [Hansen, 2004], so we use the ice tensile strength value \( \sigma_c \) measured by Litwin et al. [2012] for pure water ice and mixed grain size around the millimeter.

To apply these measurements to Europa, we infer a temperature gradient in the ice shell in agreement with the work of Quick and Marsh [2015]: at a depth of 10 km in the ice shell, the temperature is nearly 200 K, which gives \( \sigma_{c,200K} = 1.7 \text{ MPa} \) [Litwin et al. 2012], and at the surface, the temperature is approximately 100 K, with \( \sigma_{c,100K} = 1 \text{ MPa} \) [Litwin et al., 2012]. We make the assumption of a conductive lid extending from the surface to a depth of 10 km with a linear temperature variation in the ice shell from 0 to 10 km. As in Litwin et al., 2012, we consider a linear dependence of the tensile strength on the temperature:

\[
    \sigma_c = \sigma_{c,100K} + \frac{(\sigma_{c,200K} - \sigma_{c,100K})}{10} H
\]  

(7)

Far away from the chamber, the lithostatic pressure induces an isotropic pressure field \( \sigma^0 = \rho_s g H \) where \( \rho_s \) is the ice density. The cavity is filled with cryomagma which generates an overpressure \( \Delta P \) as it freezes. Then, the total pressure in the chamber is given by:

\[
    P_{tot} = \rho_s g H + \Delta P
\]  

(8)

where \( g \) is the gravity on Europa.

The overpressure \( \Delta P \) generates a tensile strength \( \sigma_{\theta \theta} \) on the cavity walls, which is given by [McLeod and Tait, 1999; Sammis and Julian, 1987]:

\[
    \sigma_{\theta \theta} = \frac{2}{3} \sigma_c (1 - \nu)
\]
Figure 3: Fraction $n_c$ of the liquid that needs to freeze to fracture the chamber walls.

\[
\sigma_{\theta\theta} = \sigma^0 \left[ 1 + \frac{1}{2} \left( 1 - \frac{P_{\text{tot}}}{\sigma^0} \right) \right]
\]  

(9)

where $\sigma^0 = \rho_s g H$ is the lithostatic pressure field around the cavity. If $\sigma_{\theta\theta}$ exceeds a critical value $\sigma_c$, the cavity walls break. Then, hydraulic fracturing occurs if:

\[
\sigma_{\theta\theta} = \sigma_c
\]

(10)

Combining Equations (8), (9) and (10), we deduce that the chamber breaks if the overpressure reaches a critical value $\Delta P_c$:

\[
\Delta P_c = 2 (\sigma_c + \sigma^0)
\]

(11)

or:

\[
\Delta P_c = 2 (\sigma_c + \rho_s g H)
\]

(12)

Equation (12) shows that the critical overpressure $\Delta P_c$ depends on the chamber depth.

Using (5), the critical fraction of liquid $n_c$ that has to freeze to generate the critical overpressure $\Delta P_c$ is given by:

\[
n_c = \frac{\exp (\chi \Delta P_c) - 1}{\rho_c \exp (\chi \Delta P_c) - 1}
\]

(13)

Here, $n_c$ only depends on the chamber depth (see Fig. 3).

2.4. Cryomagma flow

It was suggested that cracking of Europa’s ice shell might be favored by its weakening due to tidal effects (Greenberg and Geissler 2002; Lee et al. 2005; Harada and Kurita 2006; Quillen et al. 2016), but also due to its global cooling (Manga and Wang 2007). The formal mechanical crack propagation process is out of
the scope of this paper and we consider that once the tensile strength is exceeded at the wall of the chamber, the crack propagates instantaneously to the surface.

In our model, a cryovolcanic eruption begins when the pressure in a cryomagmatic chamber reaches the critical value $\Delta P_c$ (see section 2.3) and ends when the chamber is back at isostatic pressure.

The nature of the flow is given by the Reynolds number $Re$:

$$Re = \frac{\rho Ul}{\mu}$$

where $U$ is the mean velocity of the flow in the crack, $l$ the crack width, $\rho_l$ the liquid density and $\mu$ the pure liquid water dynamic viscosity.

The limit between turbulent and laminar flow is $Re \approx 10^3$ (Bejan, 1948; Bird et al., 1960), i.e. the flow is in turbulent regime for velocities greater than approximately $10^{-4}$ to $10^{-3}$ m.s$^{-1}$ depending on the conduit geometry. Moreover, Quick and Marsh (2016) recently showed that the minimum velocity of the cryomagma fluid required to reach Europa’s surface before freezing is of order of $2.5 \times 10^{-2}$ m.s$^{-1}$ for a 4 m wide tabular dike, and of order of $8 \times 10^{-4}$ m.s$^{-1}$ for a 12 m in radius cylindrical conduit. For this reason, we make the hypothesis that the up-welling cryomagmatic flow is turbulent, and this hypothesis will be verified afterward.

When a cryomagmatic chamber opens, a dike is created (McLeod and Tait, 1999). However, a dike can evolve to become a pipe-like conduit, as it is observed on Earth (Quick and Marsh, 2016). In the case of Europa, we have no information about the conduit geometry, so we consider two different crack geometries: a dike with an elongated rectangular cross-section or a pipe-like conduit with a circular cross-section. In the following, we consider the more general hydraulic diameter $D_h$, that is define by Bejan (1948) as:

$$D_h = \frac{4A}{p}$$

where $A$ is the cross-section area of the dike or conduit and $p$ its perimeter.

At Europa’s surface, the pressure is nearly zero Pa (Hall et al., 1995), and the pressure in the chamber is $P_{tot} = P_0 + \Delta P$. The flow is maintained by the pressure difference between the two ends of the conduit. The mean flow velocity results from a force balance in the dike (Bejan, 1948; Bird et al., 1960). The total friction applied on the dike walls is $\tau_w p H$ where $\tau_w$ is the shearing stress on the walls and $p$ and $H$ are respectively the dike perimeter and length. The vertical momentum balance for a fully developed and incompressible flow gives:

$$A(P_{tot} - \rho_l g H) = \tau_w p H$$

where $A$ is the cross-section area.

$\tau_w$ is classically expressed as a function of the Fanning friction factor $f$ in turbulent flow (Bird et al., 1960):

$$\tau_w = \frac{1}{2} f \rho_l U^2$$

Combining equations (15), (16) and (17), we obtain the expression for the mean ascent velocity.
Knowing the velocity of the flow, we can deduce the cryomagma flow rate emitted at the surface. Integrating Equation (18), we can also determine the total emitted volume at the surface during one cryovolcanic event.

We develop a Runge-Kutta model to solve for the the time evolution. The method used is summarized in the flowchart of Fig 4.

### 2.5. Cryomagma freezing time-scale

As shown by Kalousová et al. (2014), a liquid water lens stored in the ice shell should efficiently percolate in direction of the subsurface ocean. In fact, the density contrast between liquid water and water ice defies the reservoir stability in the ice shell. They concluded that a liquid water reservoir could remain in the ice shell over approximately 1 to 100 kyrs, depending on the ice permeability and on the heating process. For an eruption to occur, the time to freeze do not exceed the time scale for the liquid water to percolate through the ice shell to the ocean.

In order to get an order of magnitude of the freezing time of the chamber and to evaluate the time necessary to reach the critical overpressure, we solve the Stefan problem in 1D using cartesian coordinates (see Fig. 5). This estimation should be valid because the volume of the chamber necessary to freeze to initiate an eruption is a thin ice shell layer covering the chamber walls (less than the half of the chamber radius for the biggest chambers). The freezing time scale is then given by:

\[
\tau_c = \left( \frac{S(t)}{2\lambda \sqrt{\kappa_s}} \right)^2
\]
(see Appendix A for details). Equation (19) gives the solidification time scale necessary for liquid in the chamber to freeze over a $S(t)$ thickness of the chamber. The necessary thickness required to freeze in order to fracture the chamber is obtained replacing $S(t)$ by the critical value of Equation (20):

$$S_c = n_{c}^{1/3}R_{chamber}$$  

(20)

### 3. Results on pure water

#### 3.1. Time evolution during an eruption

In this section, we consider the following parameters as an example: $V = 10 \text{ km}^3$, $H = 2 \text{ km}$. The time evolution of the pressure inside the cryomagma chamber and the mean velocity of the flow in the dike are given in Fig. 6. We consider here that the cryomagma travels across a dike with a rectangular cross-section with an area $A = 100 \text{ m}^2$.

The Fanning factor $f$ depends on the geometry and rugosity of the conduit. Since we have no information on the conduit rugosity, we take a mean value of $f = 0.01$ [Bejan, 1948; Bird et al., 1960] which is an acceptable approximation because the order of magnitude of this factor does not vary for high Reynolds flow, which is our case.

On the graph given in Fig. 6(a), we show the cryomagma mean velocity is maximum when the chamber opens, at the beginning of the eruption. At this starting point, the velocity reaches $\approx 20 \text{ m.s}^{-1}$, which gives
| Symbol | Definition | Value | Unit |
|--------|------------|-------|------|
| $A$    | dike/pipelike conduit cross section | 2.10$^3$ | J.kg$^{-1}$.K$^{-1}$ |
| $c_p$  | pure water ice heat capacity | 900 | kg$^{-m^{-3}}$ |
| $D_h$  | dike/pipelike conduit hydraulic diameter | 0.01 | m |
| $f$    | Fanning friction factor in the conduit | 1.315 | m.s$^{-2}$ |
| $g$    | gravity on Europa | 2.10$^3$ | J.kg$^{-1}$.K$^{-1}$ |
| $H$    | chamber depth under the surface | 2.10$^3$ | J.kg$^{-1}$.K$^{-1}$ |
| $k_s$  | water ice thermal conductivity | ≃2 | W.m$^{-1}$.K$^{-1}$ |
| $h$    | crack width | 1 | m |
| $L_s$  | latent heat of solidification of pure water | 3.10$^5$ | J.K$^{-1}$.kg$^{-1}$ |
| $n$    | fraction of cryomagma that freezes | - | - |
| $n_c$  | critical fraction of cryomagma that freezes | - | - |
| $p$    | dike/pipelike conduit perimeter | 1 | m |
| $P_{tot}$ | total pressure in the chamber | Pa | Pa |
| $P_0$  | lithostatic pressure | Pa | Pa |
| $\Delta P$ | overpressure generates by freezing | Pa | Pa |
| $\Delta P_c$ | critical overpressure | Pa | Pa |
| $R_{chamber}$ | chamber radius | m | m |
| $S(t)$ | location of the solidification front in the chamber | m | m |
| $S_c$  | location of the critical solidification front | m | m |
| $T_f$  | pure water fusion temperature | 273 | K |
| $T_0$  | chamber walls temperature | K | K |
| $T_s(z,t)$ | temperature gradient in the frozen part of the chamber | K | K |
| $U$    | cryomagma mean velocity in the crack during the eruption | m.s$^{-1}$ | m.s$^{-1}$ |
| $V$    | total volume of the chamber | m$^3$ | m$^3$ |
| $V_f$  | virtual volume of the liquid if not compressed | m$^3$ | m$^3$ |
| $V_{fj}$ | actual volume of liquid after freezing | m$^3$ | m$^3$ |
| $V_{emitted}$ | volume of cryomagma emitted at the surface | m$^3$ | m$^3$ |
| $\mu$  | liquid water dynamic viscosity | 10$^{-3}$ | Pa.s |
| $\rho_s$ | ice density | 900 | kg.m$^{-3}$ |
| $\rho_l$ | cryomagma density | 900 | kg.m$^{-3}$ |
| $\rho_{li}$ | virtual density of the liquid if not compressed | 900 | kg.m$^{-3}$ |
| $\rho_{lf}$ | actual density of liquid after freezing | 900 | kg.m$^{-3}$ |
| $\kappa_s$ | water ice thermal diffusivity | - | m$^2$.s$^{-1}$ |
| $\lambda$ | constant related to heat transfer | - | - |
| $\sigma_0$ | lithostatic pressure around the chamber | 1.7$ \times$ 10$^{6}$ at 200K, 1$\times$ 10$^{6}$ at 100K | Pa |
| $\sigma_{th}$ | tensile strength on the cavity walls | Pa | Pa |
| $\sigma_c$ | pure water ice tensile strength | 4.9$ \times$ 10$^{-10}$ | Pa$^{-1}$ |
| $\tau_w$ | shear stress on the dike/pipelike conduit walls | Pa | Pa |
| $\tau_c$ | solidification time-scale of the chamber | s | s |

Table 1: Table of all variables used in this article.
a Reynolds number $Re \simeq 10^8$. Only at the very end of the eruption, the flow is in the laminar regime, with velocity $< 10^{-3} \text{ m.s}^{-1}$. The hypothesis of turbulent flow is thus validated.

The graph given in Fig. 6 b) shows the pressure decrease in course of the eruption, when cryomagma is emitted at the surface. The eruption ends when the pressure inside the chamber equals the pressure due to the water column in the dike. In this particular case, the eruption lasts only a tens of hours.

These computations have also been made for cylindrical conduits after Quick and Marsh (2016). Thanks to the formulation we used in section 2.4, the cryomagma velocity on the dike only depends on the $A/p$ ratio where $A$ is the cross-section area and $p$ is the perimeter. Thus, the cryomagma flow velocity is higher for a circular conduit than for a tabular dike. Fig. 7 shows the flow velocity at the beginning of the eruption as a function of the $A/p$ ratio for the range of parameter used in our study. We took the maximum value $A/p = 50$ m, which corresponds to a 100 m in radius cylindrical conduit.

As the $A/p$ ratio has a strong influence on the flow velocity, it modifies highly the eruption duration. Fig. 8 shows the flow velocity for the parameters selected before ($V = 10 \text{ km}^3$, $H = 2 \text{ km}$, $A = 100 \text{ m}^2$).

### 3.2. Effect of the chamber depth $H$ and total volume $V$

In order to link these results to the structures we observe on Europa’s surface, we conduct a parametric study varying the chamber depth $H$ and the total chamber volume $V$. We represent the total duration of the eruptions $\tau_{\text{eruption}}$ and the total cryolavas emitted volume $V_{\text{emitted}}$ at the surface on Fig. 9 as a function of $H$ and $V$. We vary the chamber volumes between 0.1 to 10 km$^3$. In fact, these volumes are in agreement with the dimensions of putative cryovolcanic features identified by Fagents (2003).

For the simulations given in Fig. 9, we assume that the cryomagma rises through a dike with a 100 m$^2$ cross section.

During a cryovolcanic event, the total volume emitted at the surface ranges from 0.1 to 10 km$^3$, which represents 0.1 to 1% of the chamber. These volumes are relatively small, and would create only small features...
Figure 7: Flow velocity at the beginning of the eruption depending on the $A/p$ ratio. $A/p$ is maximum for a cylindrical conduit.

Figure 8: Eruption duration depending on the $A/p$ ratio for the following parameters: $V = 10$ km$^3$, $H = 2$ km, $A = 100$ m$^2$. 
These results also show that the eruption duration varies from few minutes to few tens of hours for the biggest chambers. These small time scales are in agreement with the hypothesis we made previously: the cryomagma rises isothermally through the dike or pipe-like conduit. In fact, Quick and Marsh (2016) predicted the cryomagma rising would be isothermal if this one travels faster than $10^{-2}$ m.s$^{-1}$, which is indeed the case here.

3.3. Chamber freezing time-scale

The Stefan problem solved in section 2.5 gives the thickness $S_c$ of cryomagma necessary to freeze in order to trigger an eruption. $S_c$ is given in Fig. 10 a) depending on the chamber depth and volume. The solidification of a $S_c$ thick layer takes a time $\tau_c$, which is plotted in Fig. 10 b). $S_c$ and $\tau_c$ vary as a function of the temperature difference between the inside and the outside of the chamber (respectively $T_f$ and $T_0$). The difference $T_f - T_0$ is the highest for near-surface reservoirs, where the ice temperature tends towards value of order of 100 K.

Previously, Kalousová et al. (2014) predicted that a partial molten water pocket should be efficiently transported downward in Europa’s ice crust because of the density contrast between liquid water and water ice. They concluded that a water reservoir can remain during $10^3$ to $10^6$ years, depending mostly on the ice porosity. The results we obtain here are compared with the percolation time computed by Kalousová et al. (2014) in section 5.
4. Effect of anti-freeze impurities

4.1. Impurities compositions and properties

Kargel (1991) predicted the Europa’s composition, based on the carbonaceous chondrites composition and chemical evolution. Two main impurities are expected to be present in the Europa’s aqueous crust: the magnesium sulfate $\text{MgSO}_4$, which represents 75% of the carbonaceous chondrites mass (Hogenboom et al., 1995), and the sodium sulfate $\text{Na}_2\text{SO}_4$, the second most abundant chondritic component. Thus, Europa’s crust is probably composed of the following mixture (Kargel, 1991) (percents in weight percents): 81% $\text{H}_2\text{O}$ + 16% $\text{MgSO}_4$ + 3% $\text{Na}_2\text{SO}_4$. The main physical/chemical properties (solid/liquid densities, melting/eutectic temperature) of these pure salts hydrates and of a probable mixture similar to Europa’s crust one are given in Table 2. Unfortunately, to our knowledge other properties (such as compressibility, heat capacity, thermal conductivity, latent heat, tensile strength) of briny solution have not been measured. Therefore, we use pure water values.

Thanks to the internal and tidal heating, the differentiation of Europa has been completed from the chondritic material, leading the upper hydrated crust to a mixture of 81% $\text{H}_2\text{O}$ + 16% $\text{MgSO}_4$ + 3% $\text{Na}_2\text{SO}_4$ (Kargel, 1991), used here as a reference and noted in bold in Table 2.

Other minors components are expected to be present in Europa’s crust (Kargel, 1991; Hogenboom et al., 1995; Neveu et al., 2015), which is confirmed by the spectroscopic studies (Dalton, 2007). Nevertheless, these
hydrates are expected to represent less than 1% in weight in Europa’s crust composition, and then should not have a strong influence on the water and ice densities. Their physical properties are summed-up in Table 2.

4.2. Impact of impurities

The critical fraction of cryomagma \( n_c \) that freezes to begin an eruption depends on the density contrast between the liquid in the chamber and the surrounding ice \( \rho_l - \rho_i \) (see eq. (13)). The ratio \( n_c \) is a determinant factor for the freezing time \( \tau_c \) (see section 2.3) and the emitted cryolava volume at the surface during an eruption (see sections 2.2 and 2.4). In order to take into account the inclusion of salts and impurities in this section, we take appropriated values for ice and liquid densities. We used \( \rho_l = 1180 \text{ kg/m}^3 \) and \( \rho_s = 1130 \text{ kg/m}^3 \) corresponding to the most likely water-salts mixture after Table 2.

Fig. 11 gives \( n_c \) depending on the chamber depth, as made in section 2.3, and in taking into account the salts inclusion. Much crystallization is required to reach a similar critical pressure in the chamber. In this situation, the critical fraction of cryomagma \( n_c \) increases by a factor of two. In fact, the density contrast decreases by a factor of 2, which makes a water and salts mixture less efficient than pure liquid water to generate an overpressure in the chamber. Nevertheless, the fraction \( n_c \) reaches only 25% of the chamber volume for the deepest chambers, which still allows a hydraulic fracturing of the chambers.

The modification of \( n_c \) due to the salts inclusion induces an increase of the cryomagma freezing time-scale. Fig. 12 shows (a) the thickness \( S_c \) and (b) the corresponding freezing time \( \tau_c \) obtained with \( \rho_l = 1180 \text{ kg/m}^3 \) and \( \rho_s = 1130 \text{ kg/m}^3 \). \( S_c \) is lightly increased with the addition of salts in the cryomagma by ~20 %, and \( \tau_c \) is increased by a factor of 2, but those changes are not enough to modify the eruption behavior.

We also calculate the difference between the eruption durations and total emitted volumes obtained with pure water cryomagma (see Fig. 6) and these obtained with the anti-freeze mixture. Fig. 13 a) represents the variation of the eruption duration \( \delta t = \frac{t_{\text{mixture}} - t_{\text{pure water}}}{t_{\text{pure water}}} \) in percent induced by the inclusion of salts.
| Solution | Liquid density (kg/m$^3$) | Solid density (kg/m$^3$) | Melting temperature (K) | Eutectic temperature (K) | Reference |
|----------|-----------------------------|---------------------------|-------------------------|--------------------------|-----------|
| **Major hydrates** | | | | |
| water (ice I) | 1000 | 917 | 273 | - | McCarthy et al., 2007 |
| MgSO$_4$-7H$_2$O (19.6% MgSO$_4$) | 1226 | 1670 | 321.6 | 268 | Quick and Marsh, 2016 and McCarthy et al., 2007 |
| MgSO$_4$-11H$_2$O (17% MgSO$_4$) * | 1180 | 1510 | 275 | 269 | Hogenboom et al., 1995 |
| Na$_2$SO$_4$-10H$_2$O (4% NA$_2$SO$_4$) | 1038 | 1460 | 305.6 | 272 | McCarthy et al., 2007 after Kargel, 1991, Quick and Marsh, 2016 |
| **Minor hydrates** | | | | |
| KCl-nH$_2$O (19.9% KCl) | 1132 | | 262 | | Quick and Marsh, 2016 |
| NaCl-2H$_2$O (23% NaCl) | 1200 | 1610 | 273.3 | 252.4 | Quick and Marsh, 2016 and McCarthy et al., 2007 |
| MgCl$_2$- nH$_2$O (21% MgCl$_2$) | 1200 | | 239.4 | | Quick and Marsh, 2016 |
| CaCl$_2$-6H$_2$O (30% CaCl$_2$) | 1282 | | 223.2 | | Quick and Marsh, 2016 |
| H$_2$SO$_4$-6.5H$_2$O (35.7% H$_2$SO$_4$) | 1283 | 1540 | 219.4 | 211.3 | Quick and Marsh, 2016 and McCarthy et al., 2007 |
| H$_2$SO$_4$-4H$_2$O (37% H$_2$SO$_4$) | 1290 | | 198 | | Quick and Marsh, 2016 |
| **Mixtures** | | | | |
| 47% H$_2$O + 53% MgSO$_4$-12H$_2$O | 1180 | 1126 | | | Kargel, 1991 |
| 81% H$_2$O + 16% MgSO$_4$ + 3% Na$_2$SO$_4$ | 1180 to 1190 | 1133 | | 268 | Kargel, 1991 and Quick and Marsh, 2016 |

* In Peterson and Wang (2006) we found the phase diagram of MgSO$_4$ hydrates from Kargel (1991) but the MgSO$_4$-12H$_2$O phase is modified in MgSO$_4$-11H$_2$O after Peterson and Wang (2006)’s work.

Table 2: Most likely impurities in Europa’s ocean and ices, and their properties. All the hydrates mixtures are given at their equilibrium phase in Europa’s ice crust conditions ($P \approx 1$ bar, $100 \lesssim T \lesssim 273$ K). Salts percents are given in weight percents. The most likely situation in bold is tested in section 4.2.
Figure 12: (a) Thickness $S_c$ of the critical frozen cryomagma fraction and (b) time $\tau_c$ required to freeze a cryomagma layer of thickness $S_c$ depending on the chamber depth and volume. Here, the cryomagma is composed of a 81% H$_2$O + 16% MgSO$_4$ + 3% Na$_2$SO$_4$. 
13 b) represents the variation of the total cryomagma volume emitted at the surface $\frac{\Delta V}{V}=\frac{V_{\text{mixture}}-V_{\text{pure water}}}{V_{\text{pure water}}}$ in percent induced by the inclusion of salts.

Results show that the impurities will increase the duration of the eruption from 0 to approximately 12%. The effect seems to be independent of the chamber volume but has a strong dependence on the chamber depth. It reaches a maximum around $1.5 \times 10^3$ m below the surface. The total emitted cryomagma volume has the same behavior: impurities will increase the volume up to 9% at $2.5 \times 10^3$ m depth.

5. Discussion

We propose a model of effusive eruption of liquid cryomagma from water reservoirs in Europa’s ice shell to the surface. We explore the results of this model for chamber volumes from $10^8$ to $10^{10}$ m$^3$, and chamber depths from few meters to 10 km under the surface. We find that a pure water liquid cryomagma pocket should partially freeze in $10^2$ to few $10^3$ years. When enough cryomagma is frozen (0.1 to 10% of the liquid volume), the overpressure in the chamber can break the chamber walls. The cryomagma then rises at the surface thought a dike or a pipe-like conduit at high velocity, around 10 m.s$^{-1}$. The eruption ends after few seconds for the smallest chambers and few tens of hours for the greatest and deepest ones.

Our results show that the anti-freeze impurities composing Europa’s crust, i.e. MgSO$_4$ and Na$_2$SO$_4$ hydrates, has only a minor influence on the time scale required to freeze the chamber in order to begin the eruption. In fact, this time scale is a factor of two higher with the salts inclusion. A cryomagma pocket composed of a mixture of H$_2$O, MgSO$_4$ and Na$_2$SO$_4$ should freeze in $10^2$ to few $10^4$ years. On the other
hand, once the eruption begins, the evolution of the pressure in the chamber and the flow velocity is nearly identical with or without the impurities.

Previously, Kalousová et al. (2014) showed that a partial melted water lens in Europa’s shell should be efficiently transported downward due to the percolation through the porous ice. They computed that a water lens should join the internal ocean in $10^3$ to $10^5$ years. Fig. 14 shows the percolation time $\tau_p$, taken as $10^3$ years: the shortest computed by Kalousová et al. (2014), compared to our freezing time scale $\tau_c$. This ratio $\tau_p/\tau_c$ is plotted in Fig. 14 for the pure liquid water in a) and for brine in b). The chambers for which $\tau_p/\tau_c > 1$ can generate an eruption before being transported to the ocean. This configuration is represented in blue in Fig. 14. The opposite case, where $\tau_p/\tau_c < 1$, is given in red in Fig. 14. This result show that chamber at shallow depth and small volumes are favoured for the eruption. In addition, the antifreeze blending is restricting the zone of possible cryovolcanic event. If we consider percolation time $\tau_p > 10^4$ years, also possible in the range estimated by Kalousová et al. (2014), the eruption will always occur, since our $\tau_c < 10^4$.

In their recent study, Neveu et al. (2015) shown that an overpressure of $\Delta P = 1$ MPa in a chamber is high enough to propagate a crack over 60 km in a pure water ice shell, which is in good agreement with our results that could produce up to $\Delta P = 10$ MPa (see section 3.1).
Our simulations do not take into account the case where the liquid mixture has a lower density than
the solid corresponding phase, i.e.: \( \rho_i > \rho_w \). In fact, this configuration does not allow the pressure increase in
the chamber while freezing. Nevertheless in this case the liquid cryomagma is buoyant in the ice shell, which
could make easier a diapiric activity.

The principal limitation of our work is the lack of information on some parameters for Europa. For
instance, it is expected that the porosity has an influence on the results, because of its implication on the
ice mechanics. In fact, the ice porosity and grain size affects the ice tensile strength of the ice crust, which
could lower the critical overpressure \( \Delta P_c \) if the ice shell was previously weakened. Future missions should
bring constraint on these physical parameters.

Another source of limitation is about thermomechanical properties of brines. We did not found infor-
mations on the compressibility, heat capacity, thermal conductivity, latent heat, tensile strength of the
salts/hydrates impurities. Even they are not expected to dramatically change from the pure water case, it
would be very interesting to perform laboratory measurements to evaluate thermomechanical properties of
brines.

6. Appendix A:

In this model, we consider that at time \( t = 0 \), the chamber is totally filled with liquid water, at a uniform
temperature that is the fusion temperature \( T_f \). The ice surrounding the chamber is at a temperature \( T_0 \), that
remains constant during the thermal transfer (i.e. the thermal transfer in the ice surrounding the chamber
is neglected). For \( t > 0 \), the liquid in the cavity progressively freezes: the solidification front progresses in
direction of the center of the chamber. At time \( t \), the solidification front is located at the position \( S(t) \) on
the radial coordinate, with \( S(t = 0) = 0 \) and \( S(t \to \infty) = R \) where \( R \) is the chamber radius.

In order to simplify this problem, the orientation of the radial coordinate is set in the same direction
that the solidification front trajectory. Thus, the coordinate \( z = 0 \) refers to the chamber walls, whereas the
coordinate \( z = R \) refers to the center of the chamber. This set up is summed up in Fig. 5.

All along this section, the physical properties referring to the solid part of the chamber (i.e. for \( z < S(t) \))
are specified with an index \( s \), whereas the properties referring to the liquid part (i.e. for \( z > S(t) \)) are
specified with a \( l \) index.

The initial and limit conditions are summarized as follows :

\[
\begin{align*}
t = 0 : S(0) &= 0 \\
t > 0 : T_s(r = 0) &= T_0 \\
(T_s)_{r=S} &= T_f \\
T_l &= T_f
\end{align*}
\]  

The heat transfer in the solid part is ruled by the following equation:

\[
\frac{\partial T_s}{\partial t} = \kappa_s \Delta T_s
\]  

21
where $\kappa_s = \frac{k_s}{\rho_s c_p}$ is the thermal diffusivity in the solid part of the chamber, with $k_s$ the thermal conductivity of the solid part in W.m$^{-1}.K^{-1}$, $\rho_s$ is the pure water ice density and $c_p$ is the pure water ice heat capacity. The thermal transfer only depends on the $z$ coordinate, so we have:

$$\frac{\partial T_s}{\partial t} = \kappa_s \frac{\partial^2 T_s}{\partial z^2}$$

(23)

The Neumann’s solution for the heat transfer takes the form:

$$T_s(z, t) = \text{Berf} \left( \frac{z}{2\sqrt{\kappa_s t}} \right) + C$$

(24)

Thanks to the boundary conditions given in (21), we obtain:

$$C = 0$$

$$B = \frac{T_f - T_0}{\text{erf} \left( \frac{S}{2\sqrt{\kappa_s t}} \right)}$$

Finally, the temperature profile in the solid part of the chamber takes the form:

$$T_s = \frac{T_f - T_0}{\text{erf} \left( \frac{S}{2\sqrt{\kappa_s t}} \right)} \text{erf} \left( \frac{z}{2\sqrt{\kappa_s t}} \right) + T_0$$

(25)

Moreover, the heat flux conservation at the solidification front $S(t)$ gives us:

$$-k_s \left( \frac{\partial T_s}{\partial z} \right)_{z=S} = \rho_s L_s \frac{dS}{dt}$$

(26)

where $L_s$ is the latent heat of solidification of pure liquid water. For the following calculations, we set:

$$\lambda = \frac{S}{2\sqrt{\kappa_s t}}$$

(27)

Thus, the eq. (25) in eq. (26) gives:

$$\lambda \exp \left( \lambda^2 \right) \text{erf} \left( \lambda \right) = \frac{c_p(T_f - T_0)}{L_s \sqrt{\pi}}$$

(28)

Once $\lambda$ is numerically computed, we can deduce a solidification time scale $\tau_c$ using the expression of $\lambda$ given in eq. (27):

$$\tau_c = \left( \frac{S(t)}{2\lambda \sqrt{\kappa_s}} \right)^2$$

(29)

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