Constraints on effusive cryovolcanic eruptions on Europa using topography from Galileo images

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Abstract

Images of Europa’s surface taken by the Galileo Solid State Imager (SSI) show smooth features measuring a few kilometers, potentially resulting from eruptions of low-viscosity material. We estimate the volumes of four of these smooth features by producing digital elevation models (DEM) from Galileo/SSI images. We use the shape-from-shading method with special care to estimate the uncertainties on the produced DEMs and estimate the features’ volumes to be between \((5.7 \pm 0.9) \times 10^7\) m\(^3\) and \((2.7 \pm 0.4) \times 10^8\) m\(^3\). We discuss the implications for putative sub-surface liquid reservoir dimensions in case of eruptions induced from freezing reservoirs. We improved upon our previous cryovolcanic eruption model by considering a cycle of cryomagma freezing and effusion and by estimating the vaporized cryolava fraction once it spreads onto Europa’s surface. Our results show that the cryomagma reservoirs would have to be quite large to generate these smooth features (1 to 100 km\(^3\) if the flow features result from a single eruption, and 0.4 to 60 km\(^3\) for a full eruption cycle). The two future missions JUICE (ESA) and Europa Clipper (NASA) should reach Europa in the late 2020s. They shall give more information on those putative cryovolcanic regions that are interesting targets to understand the surface/interior exchanges.

1. Introduction

The Jovian moon Europa is highly likely to hide a global liquid water ocean under its ice crust (Khurana et al., 1998; Pappalardo et al., 1999). This ocean is expected to be in contact with the silicate floor, which could allow the chemical exchanges needed to create a rich habitable environment (Kargel, 1991; Kargel et al., 2000). The habitability of ocean worlds highly depends on the chemical conditions and equilibrium in their water layer (Vance et al., 2018), but for now, it remains impossible to directly sample Europa’s ocean. Also,
water coming from the ocean seems unlikely to erupt at its surface because of the very high pressure required for it to ascend through the whole ice crust (Gaidos and Nimmo 2000; Manga and Wang 2007; Rudolph and Manga 2009). However, the strong tides induced in Europa’s ice crust by Jupiter could be at the origin of local melting of the ice (Tobie et al. 2005; Mitri and Showman 2008), and such melted reservoirs seem to be good candidates to concentrate life forms if they were able to remain dormant in the ice between melting episodes (Gaidos and Nimmo 2000). Identifying the geological features emplaced during eruptions of liquid reservoirs could then give the location of the terrains the most likely to present biosignatures, which can be useful to the two upcoming missions JUICE (ESA) and Europa Clipper (NASA). In Lesage et al. (2020), we demonstrated the feasibility of erupting liquid water from sub-surface freezing reservoirs. Here, we propose to test an improved version of our previous model against real data: we first select geological features that could result from the eruption of liquid water/brines at the surface, we then generate the Digital Elevation Model (DEM) of the chosen features and measure their volumes. We then use these results to constrain the volume and depth of the source reservoirs.

The highest resolution images of Europa were acquired during the Galileo mission with the Solid State Imager (SSI) (Belton et al. 1992). These images have shown a geologically active surface, characterized by a wide variety of features (Greeley et al. 1998, 2000). The low crater density found on Europa demonstrates the vigorous resurfacing processes taking place on this moon, making its surface one of the youngest in the solar system, with an age under 90 Myrs (Zahnle et al. 2003). Plate tectonic-like processes are thought to be the cause of the frequent recycling of the icy surface (Sullivan et al. 1998; Kattenhorn and Prockter 2014). Cryovolcanic activity may also participate to cover an older surface with fresh material.

In addition, a wide range of local-scale geological features is observed at Europa’s surface, such as chaos, lenticulae, domes, pits, ridges, and faults (Greeley et al. 1998, 2000). Several formation mechanisms have been proposed to explain the emplacement of these features and invoke, in most of the cases, a diapiric ascent of warm ice (Head et al. 1999; Sotin et al. 2002; Fagents 2003; Schenk 2004; Quick and Marsh 2016), or a direct link with the ocean (Greenberg et al. 1999; Greenberg and Geissler 2002). More recent studies show that the formation of some of these features could be related to the presence of sub-surface liquid reservoirs, as is the case of lenticulae (Michaut and Manga 2014; Manga and Michaut 2017), chaos (Schmidt et al. 2011) or double ridges (Johnston and Montési 2014). Some numerical models also showed the possibility of generating locally melted zones in Europa’s ice crust, more precisely due to the combination of convection and tidal heating in the ice crust (Sotin et al. 2002; Mitri and Showman 2008; Han and Showman 2010).
In their exhaustive classification of the features seen at Europa’s surface, Greeley et al. (2000) introduced the “smooth plains” units, which are defined as smooth surfaces, with no visible texture, that embay or overprint preexisting terrains. Greeley et al. (2000) proposed two models of formation for the smooth plains, which are (1) the cryovolcanic emplacement of low-viscosity material (such as liquid water-based mixture) and (2) the melting of the surface due to a local heat source. Nevertheless, ice melting seems to be possible only for depth greater than 5 km in the ice shell (Sotin et al., 2002; Vilella et al., 2020), which is not consistent with a surface melting scenario. Also, because of geological considerations when looking at the geometry of the contours limiting the smooth plains and their intersections with the surrounding terrains, these features are, to date, widely interpreted as the results of liquid flows onto the surface (Pappalardo et al., 1999; Fagents, 2003; Miyamoto et al., 2005). Miyamoto et al. (2005) modeled the flow of a liquid under Europa’s surface conditions and have shown that the effusion of a low viscosity material such as water or brines may create flow-like features before freezing that are consistent with the morphologies of some thin smooth plains. In this study, we focus on small-scale smooth plains a few kilometers wide that we call “smooth features” hereafter.

In Lesage et al. (2020), we modeled the eruption of liquid cryolava from a freezing sub-surface reservoir. In the case of a single eruption, we showed that the erupted volume of cryolava mainly depends on the reservoir volume and depth. Assuming that smooth features may be formed by the effusion of liquid cryolava to the surface, measuring the volume of the cryolava flows will provide constraints on the volume and depth of the cryomagma reservoirs using the framework and results of Lesage et al. (2020). To estimate the volume of putative flow features, we generate DEMs of Europa’s surface using the AMES Stereo Pipeline (ASP) (Beyer et al., 2018). This open-source tool has been used previously by several authors to generate DEMs of Europa’s surface: for instance, Schenk (2004) has generated DEMs of chaotic terrains; Dameron (2015) proposed a statistical study of the double ridges morphology; a large database that includes DEMs of putative cryolava domes is currently being constructed (Nunez et al., 2019). Here we develop our own cryovolcanic features database based on morphologic criteria (see sec. 2.1), such as a flat topography, a smooth aspect, and an elevated relief. We selected four images with smooth features compatible with liquid flows, and we calculate the volume of these features. We also take special care of the estimation of the uncertainties on the DEMs (see supplementary materials for details).
2. Methods

2.1. Selection of cryovolcanic features: criteria

To obtain the volume of the putative products of cryovolcanic eruptions, we generate DEMs of features that could have possibly been emplaced during the effusion of liquid water at the surface. These features, called hereafter “smooth features”, have common characteristics with the smooth plains, previously defined by [Greeley et al. (2000)]. To select the most relevant features on the Galileo images and avoid confusion with other features, such as chaos, we choose three criteria to define the putative cryovolcanic features:

1. **A flat topography.** Because we are interested in features that may have been emplaced during the effusion of liquid onto the surface, and as modeled by [Miyamoto et al. (2005)], the resulting features are expected to be very smooth, with a relatively flat topography.

2. **No visible blocks of older surface.** Some features at Europa’s surface present a smooth matrix, but also contain blocks presenting a texture similar to the pre-existing, ridged plains. These features, named chaos, are identified as the result of the local melting and disruption of the surface, followed by its freezing ([Greenberg and Geissler, 2002], [Figueredo, 2002], [Schmidt et al., 2011]). Here we study smooth features assumed to have been emplaced by effusion of liquid cryolava onto the surface, which is a completely different process. Nevertheless, we do not rule out the possibility of forming new blocks of ice, either during the flowing of water at the surface because of the very low temperature at Europa’s surface, or due to the transport of blocks coming from the reservoir or fracture. Hence, we do not exclude features containing relatively small blocks (covering a surface fraction less than $\sim 10\%$), as long as they do not present ridged surfaces.

3. **An elevated relief.** If a liquid is extruded at Europa’s surface, some material is added on top of the older terrain, so the resulting feature is expected to be higher than the surrounding terrain, or at least higher than the surrounding valleys. This criterion is mostly verified after the DEM generation, as it is difficult to be estimated from the images only.

We looked at all SSI images and found 4 regions fulfilling these criteria (see Fig. 1). Actually, a fifth zone was identified (image 8613r) but unfortunately, we never managed to produce a comprehensive DEM.

2.2. DEM generation

We use the NASA AMES Stereo Pipeline (ASP) [Beyer et al., 2018] to produce DEMs from the Galileo SSI images. A few steps are required to obtain a ready-to-use DEM, that are summarized in the “ISIS” and
Figure 1: Images selected to generate DEMs. (a) Image 5452r (-1°, 340°); (b) Image 0713r (-79°, 124°); (c) Image 0739r (-81°, 132°); (d) Image 9352r (-28°, 218°). All these images are from the Galileo SSI data. Scales are indicative as these images are non-projected. Orange arrows show the direction of the sunlight.
“AMES StereoPipeline” parts of the flowchart in Fig. 2. After calibration, correction, and projection of the image using ISIS 3 (https://isis.astrogeology.usgs.gov) (see details in supplementary materials, section 1), two main tools can be used with ASP in order to produce DEMs: the Stereo tool, based on the stereoscopic analysis, and the Shape from Shading (SfS) tool, based on the photometric principle.

The Stereo tool produces robust DEMs because this method is based on the correspondences between pixels of two (or more) images (Beyer et al., 2018). This process requires at least two images of the zone, each taken from a different point of view (at least a few degrees of difference) and illuminated from a similar direction. These two criteria are very limiting in the case of Galileo SSI data: due to the limited number of high-resolution images, the image pairs that satisfy these two criteria are extremely rare. For this reason, we could not use the Stereo tool, and we focus on the SfS. The Stereo tool is expected to give more robust results, but SfS produces 3 to 5 times more resolved DEMs (Nimmo and Schenk, 2008) which allows a more precise study of the small scale elevation variations. Also, even though SfS is only able to give relative heights, and cannot be used to infer an absolute elevation, this is not limiting in our case as we only want to know the height of cryovolcanic features relative to the surrounding terrain.

To produce the DEMs presented here, we use the SfS tool that is based on the photometric principle (Alexandrov and Beyer, 2018). SfS uses the variations in light intensity of each pixel of an image to deduces the surface relief. In fact, the brightness of each surface facet, which is the surface square represented by a pixel, depends on the angle between the sunlight incidence direction and the facet normal. Given the knowledge of the solar elevation and azimuth, SfS computes the slope of each pixel of the image from its brightness. Then, it integrates the pixels slope to give the shape of the terrain. Numerically, this process is done by minimizing the following cost function (Alexandrov and Beyer, 2018):

$$
\int \int [I(h(x,y)) - T.A.R(h(x,y))]^2 + \mu^2 \parallel \nabla^2 h(x,y) \parallel^2 + \lambda^2 [h(x,y) - h_0(x,y)]^2 \, dx \, dy
$$

with $h(x,y)$ the function describing the heights of the terrain, $I(h(x,y))$ is the camera image interpolated at pixels obtained by projecting into the camera 3D points from the terrain $h(x,y)$, $T$ is the image exposure, $A$ is the terrain albedo (took constant on the whole image), $R(h(x,y))$ is the reflectance, $h_0(x,y)$ is the initial guess terrain (here a flat DEM provided by the user), and $\mu$ and $\lambda$ are two positive coefficients chosen by the user and controlling respectively the smoothness and the weight of the terrain initial guess.

The first term of the cost function constrains the brightness and ensures that the simulated light intensity
fits with the intensity recorded by the camera. This term depends on the reflectance model used, which can be chosen by the user. The Lambertian, LunarLambert and Hapke models are available to model the icy surfaces. According to recent photometric study [Belgacem et al., 2020], we use the Hapke model with the following parameters: $\omega=0.9$, $b=0.35$, $c=0.65$, $B_0=0.5$, $h=0.6$. The influence of the photometry on the DEMs is developed in supplementary materials (section 2). The relative uncertainty due to photometry has been estimated to $\pm 10\%$ in the measured volume (see supplementary materials).

The second term of the cost function controls the smoothness of the output DEM by minimizing the second-order derivative of the slope on each point. The smoothness coefficient $\mu$ is chosen by the user. Theoretically, the higher $\mu$ is, the smoother the DEM is, making the small-scale details less visible and flattening higher relief features. Nevertheless, we noticed that very small values of $\mu$ also produce flattened DEMs. To avoid an extremely flattened DEM, we typically choose $1<\mu<10$. We tested several values of this coefficient for each image and concluded that there is no ideal value of $\mu$ that can be used for all the images. In fact, the smoothing effect controlled by $\mu$ depends on the terrain roughness and therefore differs for each image studied (Alexandrov and Beyer, 2018). For each image, we need to test several values of $\mu$ to keep the most appropriate one. The precise effects of the variation of $\mu$ on the produced DEM is shown in the supplementary materials (section 2). For small features, we found that the DEM is always consistent with the feature heights estimated from shadows (see supplementary materials section 2), in a large domain of the smoothness parameter $\mu$. The relative uncertainty due to the smoothness coefficient has been estimated at $\pm 5\%$ in the estimated volume (see supplementary materials).

Finally, the third term of the cost function describes the difference between the calculated DEM and the initial guess $h_0(x, y)$ given to SfS. Concretely, $h_0$ is an input DEM that SfS uses to start the minimization of the cost function. This DEM is provided by the user and used by SfS to constrain the generated DEM. This option can be used to enhance the quality of a previously generated DEM (using Stereo for example). Here, we do not have an initial guess of the terrain elevation and SfS is the only tool used to generate the DEMs. Hence, we use a flat terrain at 0 elevation as the initial guess, and we set the parameter $\lambda=0$. By doing this, SfS does not give importance to this flat initialization DEM during the minimization iterations.

Finally, the total relative uncertainty from the DEM is $\pm 15\%$ in the estimated volume (see supplementary materials).
2.3. Volumes estimation

In order to estimate the cryolava volume, the first step is to calculate the total volume of the smooth feature from the DEM and this is detailed in the next section. With the example of image 5452r, we show in section 2.3.2 that it is necessary to invoke subsidence to explain the morphology of the smooth features. To estimate the real cryolava volume, it is necessary to subtract the estimated volume of the underlying topography (ridges in our case, coupled with terrain subsidence). The hypothesis and methods used to estimate the topography are detailed in section 2.3.3.

2.3.1. Gis processing: simple approach

We first calculate the volume of the smooth features using QGis (QGIS Development Team 2019) by a simple approach. The idea is to subtract the pre-existing terrain beneath the cryolava flow to the feature itself, to only keep the cryolava material and calculate its volume. A few steps are necessary to obtain this result and are summarized in the “QGis” part of the flowchart given in Fig. 2. First of all, the smooth feature needs to be delimited. We draw the shape of the feature and exclude the ridges that might intersect it. Then, the pre-existing terrain beneath the cryovolcanic feature is inferred from the surrounding terrain. In order to estimate it, we use the interpolation tool in QGis to create a layer that approximates the terrain elevation under the feature. Once the pre-existing terrain elevation is subtracted from the cryovolcanic feature DEM, we sum the height of all the pixels composing the feature and multiply it by the area of a pixel, so we obtain the volume of the cryolava flow. This volume is called “measured volume” (noted $V_{measured}$) in the following sections.
2.3.2. Subsidence and thermal erosion: example of image 5452r

Image 5452r was taken by the Galileo spacecraft during its fourth orbit with a resolution of approximately 30 m/px. It shows a very smooth circular feature. This feature was first presented by Head et al. (1998) and then described by Pappalardo et al. (1999) as a “smooth deposit, probably emplaced as a cryovolcanic eruption of low-viscosity material, perhaps liquid water”. This interpretation is now accepted in most of the literature because of the feature morphology (Fagents, 2003; Miyamoto et al., 2005).

Some impact craters are visible on the smooth feature and especially a large one, almost centered. We assume that all craters have been created after the emplacement of the smooth feature and have no relation with its formation. Firstly, because other craters are also present around the smooth feature and do not seem to interact with the older terrain (no melting, no particular ejecta...), and secondly because if the smooth feature was created by melting after the impact, the impact crater itself should not be visible anymore.

Image 5452r and its DEM generated with SfS is shown respectively in Fig. 3a and 3b. We determine the smooth feature edges (in blue on Fig. 3b) using the DEM. In fact, in the particular case of image 5452r, the sunlight comes from the east side of the image, which is the direction perpendicular to the surrounding ridges and thus makes the east-west oriented lobes hardly visible on the raw image. Nevertheless, the edges of the feature are well delimited on the DEM, more precisely in the valleys between the ridges. We also choose five points located in valleys adjacent to the feature (see the black crosses on Fig. 3b). These points are used to estimate the pre-existing terrain elevation beneath the feature (see dashed lines in Fig. 4). By interpolation, we obtain a surface that is subtracted from the smooth feature. By doing this, we can calculate the feature volume itself. DEMs of the three other images used in this study are given in supplementary materials, section 3.

We propose geomorphological interpretations of this smooth feature. First of all, the smooth feature, mostly in white on the DEM, has an elevation of around 0 m, which is higher than the surrounding valley bottoms, that have a negative elevation. At the locations indicated by the blue arrows in Fig. 3 one can see the feature edges, filling the bottom of the valleys. The smooth feature is thus possibly made of low-viscosity material added on the pre-existing terrain, flowing between the ridges.

Moreover, on the north, south, and west of the smooth feature, we can see that some ridges are not covered by the smooth material, which indicates that the putative flow had a relatively low viscosity (Miyamoto et al., 2005).

Finally, at the central region of the smooth feature, ridges are not visible. This is not trivial because
the surrounding ridges have a positive elevation, higher than the smooth feature, so a flowing liquid is not expected to cover them. Two scenarios can be put forward to explain this effect and are illustrated in Fig. 4 (see the two topographic profiles numbered 1 and 2 in Fig. 3b, 3c and 3d). The emplaced material could generate a local flexure (subsidence) of the crust (model A) or it could melt/erode the surface as it flows (model B). Both cases, or a mix of them (model C), might explain a lower topography of the pre-existing terrain preferentially near the center.

Model A detailed in Fig. 4a could result from the local flexure of the ice crust due to the presence of a liquid reservoir at depth centered on the feature. The required condition to create a few kilometers of large depletion is a thin elastic layer of less than a few hundred meters under the reservoir (Manga and Michaut, 2017). With this formation model, the real pre-existing terrain beneath the smooth feature has an elevation lower than the pre-existing terrain estimated with the DEM, which is the mean level of the surrounding valleys. This has to be taken into account in the volume measurement (see section 2.3.3).

In the case of model B, as shown in Fig. 4b, the ridges are subject to thermal erosion only. This could happen if a warm liquid flows onto the surface (see thermal erosion experiment from Kerr, 2001). In this case, the flowing liquid would be composed of a mixture of cryolava and molten terrain. After freezing, the molten terrain would be back to an ice state with a density similar to its original state, so finally, the net volume change due to thermal erosion is null. In the case of image 5452r, the smooth feature is around 0m elevation, which is also the mean elevation of the surrounding terrain because of the apriori terrain chosen for the DEM generation. This would mean that in case of melting and freezing of the terrain, no material coming from the interior was added onto the surface and thus only heat transfer is responsible for the feature emplacement. Such shallow surface ice melting is not in agreement with interior models (Sotin et al., 2002; Mitri and Showman, 2008; Vilella et al., 2020) so terrain subsidence is more likely to explain the smooth feature morphology, and we do not consider model B hereafter.

Model C is a combination of thermal erosion and local subsidence, as is illustrated in the sketch from Fig. 4c. Like in the case of model A, the real pre-existing terrain is lower than the estimated one, implying a difference between the flow volume measured from the DEM and the real cryolava volume coming from the reservoir. This has to be taken into account and is detailed hereafter.
Figure 3: (a) Image 5452r from Galileo SSI and (b) the associate DEM produced with SfS. The blue line indicates the limit of the flow-like feature. Blue arrows show the lobate zones interpreted as filling the valleys between the ridges. The two profiles numbered (c) 1 and (d) 2 are respectively perpendicular and parallel to the surrounding ridges. Schematic representations of sections of the feature along these two profiles are shown in Fig. 4. Black crosses indicate the chosen points used to calculate the reference level of the pre-existing terrain beneath the smooth feature (see Fig. 4).
Figure 4: Schematic views of the topographic profiles 1 and 2 respectively perpendicular and parallel to the surrounding ridges (see Fig. [3]). The grey zone represents the pre-existing ridge terrain whereas the light blue zone stands for the smooth feature. The dark blue points delimitate the edge of the smooth feature corresponding to the dark blue line in Fig. [3]. The black crosses show the bottom of the nearest ridges used as reference elevations to interpolate the pre-existing terrain level used to calculate the feature volume (represented here by the black dotted line). The actual pre-existing terrain of the feature may be different than the interpolated one and is plotted with a dotted grey line. Three mechanisms are sketched: (a) subsidence of the surface under the feature (model A), (b) melting or thermal erosion of the pre-existing terrain (model B) and (c) a mix of these two processes (model C). A difference between the interpolated pre-existing terrain and the real one exists in the case of Model A and C.
2.3.3. From measured volumes to cryolava volumes

On Europa, most of the terrains are ridged (see images in Fig. 1 and Greeley et al., 2000). Thus, the simple methodology proposed in section 2.3.1 may overestimate the volume of actual erupted material as it also takes into account the volume of the ridges. As we are interested in the actual volume of cryolava erupted onto the surface, we propose to use a volume factor $\alpha_V$ which expresses the actual cryolava volume $V_{\text{cryolava}}$ with respect to the apparent volume $V_{\text{measured}}$ measured with the DEMs:

$$V_{\text{cryolava}} = \alpha_V V_{\text{measured}}$$  \hspace{1cm} (2)

The calculation of $\alpha_V$ is illustrated in Fig. 5. We use a topographic profile AB extracted from the ridged plain near the smooth feature from image 5452r (see Fig. 5a) to estimate $\alpha_V$. The topographic profile along this section is shown in Fig. 5b.

In Fig. 5c, we show a simplified situation where the cryolava flows between the ridges without any subsidence of the terrain. We calculate the cross-sectional area of the smooth feature measured with the simple approach $A_{\text{measured}}$, which extends from the smooth feature top (elevation $\sim 0$ m) to the mean elevation of valley bottoms (elevation $\sim -7$ m). This area is filled in grey in Fig. 5c. We also calculate the area $A_{\text{cryolava}}$ of the cross-section filled with cryolavas after the eruption, represented in dotted in Fig. 5c. We finally calculate $\alpha_V = A_{\text{cryolava}}/A_{\text{measured}}$ using these two areas. This 2D approximation is justified by the homogeneity of the ridged terrains.

The situation presented in Fig. 5c might not be realistic because, as discussed in the previous section, a cryolava flow on the surface cannot produce the observed features without local subsidence of the terrain (see Fig. 4). Such subsidence may impact significantly the $\alpha_V$ factor, and this is what we estimated in Fig. 5d and 5e. We simulate the subsidence of the terrain of cross-section AB to calculate the associated $\alpha_V$.

We study two extrema to estimate the range of possible cases: (i) shallow subsidence of only the necessary height for the ridges to be embayed in the smooth feature ($\sim 5$ m, see Fig. 5d), (ii) deeper subsidence of 40 m, that is the maximum subsidence modeled by Manga and Michaut (2017) for subsurface reservoirs less than 10 km large (see Fig. 5e). Thank to the graphical representation of a terrain subsidence of Fig. 5, we obtain $\alpha_V \simeq 0.8$ for the shallow subsidence and $\alpha_V \simeq 5$ for the maximum subsidence. These two factors will be taken into account further.

One should also note that the putative melting or thermal erosion of the terrain during the liquid flow
should not modify the $\alpha_V$ value. Indeed, this only could transfer some eroded material from the underlying terrain to the smooth feature, with a null net weight balance (as explained in section 2.3.2). For this reason, the results calculated further are valid with or without melting or thermal erosion of the surface.

3. Eruption model

3.1. Cyclic eruptions

In [Lesage et al. (2020)], we tested the feasibility of a cryomagmatic eruption model proposed previously by Fagents (2003). We modeled a cryomagma reservoir as a spherical cavity in the Europa’s ice crust, filled with liquid cryomagma. Because of the temperature gradient between liquid cryomagma and cold surrounding ice, cryomagma freezes, from the cavity wall toward its center, and we model the solidification front position as a function of the time by solving the Stefan problem. The density contrast between the liquid and solid cryomagma generates an overpressure in the reservoir, and thus tangential stress on the wall. When the overpressure in the reservoir is high enough, the wall breaks and a fracture can propagate toward the surface (Lister and Kerr, 1991; Rubin, 1993). Then, cryolava can flow onto the surface until the overpressure in the reservoir has been released. Thanks to this model, we calculated the cryolava volume emitted during an eruption, as well as the fluid velocity during the eruption and the total duration of this event. The model is derived for two cryomagma compositions: 1) pure liquid water and 2) a briny mixture of 81 wt% $\text{H}_2\text{O} + 16$ wt% $\text{MgSO}_4 + 3$ wt% $\text{Na}_2\text{SO}_4$ that is predicted to be the Europa’s ocean and ice composition (Kargel, 1991).

These results were obtained as functions of the cryomagmatic reservoir parameters (such as its volume $V$ and depth $H$) and its environment (temperature gradient in the ice crust, ice crust composition). In [Lesage et al. (2020)], a single cryovolcanic eruption was modeled. Nevertheless, a single reservoir might trigger several eruptions during its lifetime, as shown in Fig. 6. At the end of an eruption, the liquid in the fracture freezes, which closes the reservoir, but the solidification continues and the freezing front progresses toward the reservoir center, which pressurizes the cryomagma and can lead to a second eruption. A single reservoir might hence be able to erupt several times. To take this effect into account, we compare the smooth feature volumes with the products of single eruptions but also with the total erupted volume during a cryomagma reservoir lifetime.

Fagents (2003) predicted that cyclic eruptive events could result in morphologies where multiple flow lobes are present. The cyclic eruption model presented here might explain the lobate feature that can be seen at the center of the smooth feature from image 9352r (see Fig. 1). We do not clearly see several lobate
Figure 5: Illustration of the calculation of the $\alpha_V = \frac{V_{\text{cryolava}}}{V_{\text{measured}}}$ factor. (a) We make a section in a ridge terrain and we extract a topographic profile AB from this section. (b and c) We measure the cross-sectional area of the smooth feature obtained with the simple approach $A_{\text{measured}}$ (in grey), i.e. the difference between the mean elevation of the smooth feature’s top and the elevation of the interpolated pre-existing terrain times the length AB. We also measure a more realistic cross-sectional area which should be filled with cryolava $A_{\text{cryolava}}$ (dotted). This cross-sectional area is defined as the difference between the mean elevation of the smooth feature’s top and the elevation of a more realistic pre-existing terrain, i.e. a ridged terrain with a subsidence of 5 to 40 m, times the length AB. We then calculate $\alpha_V = \frac{A_{\text{cryolava}}}{A_{\text{measured}}}$. We investigate the impact of a putative subsidence of a terrain on $\alpha_V$. We simulate a subsidence of the terrain in 2 cases: (b) a shallow subsidence, where the ridges tops are showing on the surface (-5 m, profile A’B’), and (c) a maximum subsidence value after [Manga and Michaut, 2017] (40 m, profile A”B”).
features on the other features, which may be explained by the limited resolution of the Galileo images. As an example, for image 5452r, the resolution is approximately 25 m/px. The solar incidence angle is $\sim 75^\circ$, so only features higher than $\sim 11$ m can be seen on the image with their projected shadow, assuming that the shadow is projected on flat terrain. Taking into account the complex processes expected to affect liquid water flowing onto Europa’s surface, as the possibility of endogenous cryolava flow and the competition between freezing and vaporizing (see Allison and Clifford 1987; Fagents 2003; Quick et al. 2017 and section 3.2 for details), it is hard to determine whether the smooth features could result preferentially from a single or multiple eruptions, so here we consider these two possibilities.

As an illustration, Fig. 7a shows the volume erupted at the surface for the full activity lifetime of a cryomagmatic reservoir. To obtain these results, we iterate the cryomagmatic eruption model for a $10^9$ m$^3$ reservoir located 4 km beneath the surface and filled with briny cryomagma. Each step in Fig. 7a represents a complete eruptive cycle, i.e. cryomagma freezing, pressurization, and eruption. We consider that a cryomagmatic reservoir can erupt as long as the liquid volume remaining in the reservoir is greater than the volume of liquid stored in the fracture linking the reservoir to the surface. Moreover, only the reservoir fraction that remains liquid at the end of each freezing is taken into account to calculate the erupted cryolava volume.

Fig. 7b shows the volume of cryolava erupted at the surface during each eruption. It follows a decreasing
exponential trend (note that y-axis is logarithmic). We can predict the volume $V_{\#i}$ erupted during each eruption with an exponential law. For the first eruption, the erupted volume is equal to the increase in the frozen cryomagma volume. The freezing cryomagma volume is equal to $nV$ where $n$ is the cryomagma fraction necessary to freeze to trigger an eruption \cite{Fagents2003, Lesage2020} and $V$ is the reservoir initial volume. The cryomagma volume that does not freeze is defined as $V_{0i} = V(1 - n)$. The newly formed ice has a volume $n\frac{\rho_l}{\rho_s}V$ where $\rho_l$ is the liquid cryomagma density and $\rho_s$ is the solid cryomagma density, so the liquid part of the reservoir is compressed to a volume $V_{0f} = V(1 - n\frac{\rho_l}{\rho_s})$. Finally the volume of ice added in the reservoir after freezing is \cite{Fagents2003, Lesage2020}:

$$V_{\#0} = V_{0i} - V_{0f} = nV\left(\frac{\rho_l}{\rho_s} - 1\right)$$  \hspace{1cm} (3)

Then, for the second eruption, we calculate $V_{1i}$ and $V_{1f}$ from the remaining cryomagma after the first eruption, which has a volume $V_{1i} = V_{0f}(1 - n)$. Then the erupted volume $V_{\#1}$ is:

$$V_{\#1} = V_{1i} - V_{1f} = V_{0f} (1 - n) - V_{0f} \left(1 - n\frac{\rho_l}{\rho_s}\right) = V_{\#0} \left(1 - n\frac{\rho_l}{\rho_s}\right)$$  \hspace{1cm} (4)

Eq. (4) can be generalized for all the following eruptions, so finally the erupted volume at eruption $\#i$ can be written:

$$V_{\#i} = nV\left(\frac{\rho_l}{\rho_s} - 1\right) \exp\left(\ln\left(1 - n\frac{\rho_l}{\rho_s}\right)^{\#i}\right)$$  \hspace{1cm} (5)

Fig. 7 demonstrates that the time between each eruption increases before shortening. In fact, this freezing time is a competition between two phenomena: firstly, the thermal transfer between the warm reservoir and the cold surrounding ice slows down over time as the surrounding ice temperature increases progressively, so the thermal wave propagates slower; secondly, the liquid volume decreases, which reduces the frozen layer thickness required to trigger an eruption. For the example given in Fig 7, the first phenomenon dominates from eruptions $\#1$ to $\#12$, then the second one becomes preponderant.

Fig. 8 shows the total activity lifetime of a reservoir, considering that one cryomagmatic reservoir may
erupt several times. The reservoir activity lifetime is calculated for 5 different reservoir volumes ranging from $10^6$ to $10^{11}$ m$^3$ and 6 different depths from 1 to 10 km under the surface. Lifetime increases with the reservoir volume, for both pure and briny cryomagma, ranging from 0.4 years to $10^5$ years. In addition, reservoir lifetime is approximately 10 times larger for reservoir at 10 km depth in comparison to 1 km depth. Reservoir lifetime depends on the reservoir depth because of the temperature gradient in the ice crust: reservoirs near the surface are located in a colder environment, which makes them freeze faster.

It is also possible to predict the volume of liquid erupted at the surface during the whole reservoir lifetime. Theoretically, if the reservoir remains active until all its volume $V$ is filled with ice, mass conservation imposes:

$$V \rho_l = V_{\text{erupt}} \rho_l + V \rho_s$$  \hspace{1cm}(6)$$

where $V_{\text{erupt}}$ is the total liquid erupted, $\rho_l$ is the density of the liquid cryomagma, $\rho_s$ is the density of the ice formed by the frozen cryomagma, and $V$ is the initial reservoir volume. The total volume of liquid erupted after the eruption cycles is a function of the density contrast between the cryomagma solution and the corresponding ice:

$$V_{\text{erupt}} = V \left(1 - \frac{\rho_s}{\rho_l}\right)$$  \hspace{1cm}(7)$$

In our case, we also need to take into account that the eruption stops when $V_l < V_{\text{fracture}}$ where $V_l$ is
the volume of liquid remaining in the reservoir, and \( V_{fracture} \) is the volume of liquid stored in the fracture. Finally, we obtain:

\[
V_{erupt} = V \left( 1 - \frac{\rho_s}{\rho_l} \right) - 2V_{fracture} \quad (8)
\]

In most of the case \( V_{fracture} \) is negligible regarding the volume of the reservoir so that the final erupted volume does not depend anymore on reservoir depth.

3.2. Vaporized fraction of water

The putative flow of water-based liquid on Europa takes place in a low pressure and temperature environment. The pressure at Europa’s surface is near \( 10^{-6} \) Pa (Hall et al., 1995) and the mean temperature is around 110 K (Spencer, 1999), so that the liquid water erupting at the surface is subjected to the competition between freezing and boiling. Because of the \( \approx 160 \) K difference between the liquid and the environment, several authors previously proposed that an ice crust forms rapidly on top of the flow (Allison and Clifford, 1987; Fagents, 2003; Quick et al., 2017). Allison and Clifford (1987) studied the flow of liquid water on Ganymede’s surface. As they suggested, a \( \sim 0.5 \) m thick ice crust is enough to prevent the flow from boiling,
allowing the underlying liquid to flow onto the surface, precising that it would only apply to the interior of the flow, and not to its expanding edges. There, the flow is more likely to look like a mix of boiling water and ice blocks being pushed by the liquid (Allison and Clifford 1987). The water fraction being vaporized at the surface is a key parameter to link the volume of the smooth features and the initial erupted volume of liquid water, and it is necessary to evaluate it. This quantity was roughly estimated by Porco (2006), who calculated the vaporized fraction of water during the opening of cracks on Enceladus. One should note that this study does not take into account the formation of an ice crust as discussed above, so the amount of material vaporized used in this study can be considered as an upper bound. Porco (2006) assumes that the latent heat of fusion $L_f$ generated by the freezing part of liquid water is used as latent heat of vaporization $L_v$ by the vaporized part of the fluid, so they calculate that a fraction $x = L_f / (L_f + L_v) = 0.13$ of liquid is vaporized.

This simple calculation from Porco (2006) gives an idea of the quantity of liquid turned into vapor when it reaches the icy moon’s surface, but to have a better knowledge of the transformation occurring in the two-phase region it is necessary to use the phase diagram of water. To assess the water behavior in extreme environments such as Europa’s surface, it is relevant to use a temperature-entropy (T-s) phase diagram (Lu 2009).
and Kieffer (2009) as the flow might be considered as isentropic (adiabatic and reversible) as discussed by Kieffer and Delany (1979). An isentropic process is a vertical line on a T-s phase diagram, so it is easy to deduce the properties of a flow after this representation, and one can directly read the mass ratio of gas/solid on such a diagram. Fig. 9 shows the T-s diagram of water adapted from Lu and Kieffer (2009). On this diagram, the blue arrow is the isentropic depressurizing process from liquid water at the triple point (T=273 K, P=612 Pa) to the exit state, i.e. the conditions at Europa’s surface (low pressure and low temperature). It indicates that at the end of the process, only 10 to 15% of the liquid water erupted at the surface is vaporized; the major part of the water freezes to solid.

These results show that, without taking into account the formation of an ice crust on top of the flow, the vaporized fraction of water should range in between 10 to 15% of the total erupted cryolava. The ice crust formed on top the of the flow plays a role in protecting the well-developed flow from boiling, however, the flow edges are still subject to boiling and freezing competition. We thus consider a vaporized fraction of 13±3% after Porco (2006) and Lu and Kieffer (2009) to calculate the liquid volume that may be at the origin of the smooth features, keeping in mind that this calculated volume is a lower limit of the erupted volume. Also, the addition of salts or impurities in the cryomagma could slightly modify this result as it lowers the vapor pressure on Europa from ~600 to 300-500 Pa depending on the salt content (Quick et al., 2017).

Another parameter that could affect the density contrast between the liquid and solid cryomagma phases is the formation of hydrates during the solidification process. In our previous work (Lesage et al., 2020), we considered a briny cryomagma composed of the following mixture: 81 wt% H$_2$O + 16 wt% MgSO$_4$ + 3 wt% Na$_2$SO$_4$, which is the Europa’s ocean and ice composition predicted by Kargel (1991). Phase diagram from McCarthy et al. (2007) shows that hydrates of MgSO$_4$ and Na$_2$SO$_4$ form for concentrations above respectively 17.3 and 4 wt%. Thus, in the cases considered in this work, hydrates should not form in the freezing cryomagma.

4. Results

4.1. Measured and erupted volumes

Table 1 summarizes the volumes of the smooth features measured on the four images shown in Fig. 1. The measured volumes are extracted directly from the DEM (see the method in section 2.3.1 and flowchart in Fig. 2). One should note that a measured volume is not equal to the liquid volume erupted at the surface to create the corresponding smooth feature. To obtain the erupted volume, we multiply the measured volume by
a $\alpha_V$ factor to take into account the ridges on the ancient surface and putative subsidence (see section 2.3.3 and Fig. 5). We calculate the results for two extreme values of $\alpha_V$: $\alpha_V = 0.8$, which is the case of shallow subsidence, and $\alpha_V = 5$, which describes the maximum subsidence possibly induced by a liquid subsurface reservoir of 5 km radius after Manga and Michaut (2017). Then, we multiply the volume obtained by a factor $1/0.87$ to take into account the vaporization of the erupted liquid. Finally, we multiply this volume by a factor $\rho_s / \rho_l$ ($1130/1180$ for briny cryomagma or $920/1000$ for pure water) to estimate the liquid volume before expansion due to phase change. This process is summarized in the following equation:

$$V_{erupted} = \alpha_V V_{measured} \frac{1}{(1-x)} \frac{\rho_s}{\rho_l}$$

(9)

where $V_{erupted}$ is the fluid volume erupted from the reservoir, $V_{measured}$ is the smooth feature volume measured from the DEM using the simple approach, $\alpha_V$ is a volume factor to take into account the underlying terrain (see section 2.3.3), $x = 0.13$ is the fluid vaporized fraction, $\rho_l$ is the density of the liquid cryomagma and $\rho_s$ is the density of the corresponding ice.

We propagate the uncertainties from the DEM, in order to take into account the two main uncertainty sources i.e. the smoothness coefficient and the reflectance model (and the associate albedo). Calculation of the uncertainties using the deviation of the mean are detailed in supplementary materials, section 2. We found an uncertainty of $\pm 15\%$ on the volumes measured from the DEM. Moreover, we add a $\pm 3\%$ uncertainty on the calculation of the erupted volume due to the uncertainty on the vaporized fraction (see section 3.2). The DEM of each smooth feature and the selection of the edges to extract its volume is detailed in supplementary materials, section 3. Finally, we obtain the results given in Table 1.

4.2. Cyclic or single eruption products

In Lesage et al. (2020), we obtained the volume of cryolava erupted at the end of a single eruptive event as a function of on the reservoir volume and depth, for two different cryomagma compositions: pure water and a briny mixture of water and salts: 81 wt% H$_2$O + 16 wt% MgSO$_4$ + 3 wt% Na$_2$SO (Kargel 1991). These volumes are shown in Fig. 10 for reservoir depth ranging from 1 to 10 km and reservoir volume ranging from $10^8$ to $10^{12}$ m$^3$ which corresponds to reservoir radius between $\sim 0.3$ and 6.2 km. We compare these results with the volumes measured on the four DEMs produced here in Fig. 10 (uncertainties on the erupted volumes calculated from the DEMs are shown). One can see in Fig. 10 that the eruption of a $3 \times 10^9$ to $10^{11}$ m$^3$ reservoir (0.9 to 2.9 km radius) is necessary to explain the emplacement of the smooth features from a
single cryovolcanic event for shallow subsidence of the terrain ($\alpha_V = 0.8$). For deeper subsidence ($\alpha_V = 5$), a $2 \times 10^{10}$ to $10^{12}$ m$^3$ reservoir (1.7 to 6.2 km radius) is required. The range of erupted volumes is nearly identical for these two compositions as the $\rho_{\text{ice}}/\rho_{\text{liq}}$ factor does not differ that much between pure water or a water-based mixture. One should note that larger reservoirs are expected to create deeper subsidence of the surface (Manga and Michaut, 2017), so that it is worth to consider both of these solutions.

In addition, we also want to consider in this study the case of several eruptive cycles instead of one single eruption (see section 3.1). In this case, the same reservoir will produce several eruptions during its whole lifetime. According to Fig. 8, this kind of reservoir may remain active during a lifetime ranging from $5 \times 10^3$ to $10^5$ years. Using Eq. (8), we can predict the cryolava volume that is erupted during the total reservoir lifetime. This volume depends more on the $\rho_{\text{ice}}/\rho_{\text{liq}}$ ratio than for a single eruption, as a consequence the reservoir volumes required to produce the smooth features depend slightly on the cryomagma composition. The erupted volume necessary to produce the observed features depend mostly on the cryomagma composition, and the $\alpha_V$ ratio.

We summarize in Table 2 the erupted volumes obtained using the minimum and maximum measured volumes given in Table 1, taking into account the uncertainties. One should note that these volumes do not depend on the reservoir depth (see Eq. (8)).

5. Discussion and conclusions

We identified four images from Galileo SSI data presenting smooth features that may possibly be formed during one or several eruptions of cryolava at Europa’s surface. We produced DEMs of those smooth features
Figure 10: Volume erupted at the surface during one cryovolcanic eruption after the model from Lesage et al. (2020). The smooth feature volume measured on the four analyzed images (see Fig. 1) correspond to reservoir volumes ranging in between the two dashed lines for $\alpha_V = 0.8$ and between the two solid lines for $\alpha_V = 5$ (see Table 1).

Table 2: Size of the reservoir required to erupt the cryolava amount necessary to generate the smooth features observed depending on the cryomagma composition and the $\alpha_V$ ratio. We give these results for the smallest and largest features, respectively from image 5452r and 9352r. We use the following composition for the briny cryomagma: 81 wt% H$_2$O + 16 wt% MgSO$_4$ + 3 wt% Na$_2$SO (Kargel, 1991).
using the Shape from Shading tool of the AMES Stereo Pipeline, and we measured their volumes with their associated uncertainties. The major uncertainties on the DEM generation and volume measurement come from the parameters of SfS, but the effect is relatively reduced for small scale features. We estimated the uncertainty on the volumes to be of the order of ±15% (see supplementary materials, section 2). The volume measurements of the four smooth features selected gave results ranging from $5 \times 10^7$ to $5 \times 10^8$ m$^3$.

Shape from Shading tool allowed us to generate DEMs from single images, with a high precision at low scales ([Nimmo and Schenk](#) 2008). Nevertheless, SfS presents its own limitations. First of all, it does not manage albedo or photometric heterogeneities in a single image. In fact, the reflectance model chosen by the user is applied to the whole image, but the surface properties can differ at regional or local scales, as shown by [Belgacem et al.](#) (2020). We show that the reflectance model and the associated coefficients chosen (such as albedo) are not expected to create uncertainties on volume higher than ±10% for small scale features. Nevertheless, as discussed by [Jiang et al.](#) (2017) for Martian images, too much heterogeneous surface properties in one image can lead to non-convergence of the algorithm. From the images we selected, four were converging (images presented in Fig. 1), but it was not the case of image 8613r which exhibits a smooth material darker than the brighter surrounding terrain, maybe due to an albedo heterogeneity. We could not provide a DEM of this image. It could be interesting to investigate the possibility of considering several zones with different surface properties in an image: this could solve the problem of DEM generation of image 8613r.

Geomorphological interpretations of the DEMs are consistent with the formation of the smooth features by flowing of a low viscosity fluid at the surface: smooth features own a very flat topography, filling the preexisting terrain valleys. Moreover, double ridges, at an elevation higher than the smooth features, seem to disappear beneath the smooth material, especially near the center of the feature. We argue that this could result from a flexure of the elastic lithosphere but thermal erosion is not required. To link the volume of the smooth features that we measure on the DEM using a simple approach and the actual volume of the cryolava erupted during their emplacement, we propose to take into account a range of possible subsidence values induced by the liquid reservoir underneath, as it is modeled by [Manga and Michaut](#) (2017). We studied the two extreme cases of shallow subsidence of 5 m, or maximum subsidence of 40 m. In the first case, approximately $4 \times 10^7$ to $4.4 \times 10^8$ m$^3$ of cryolavas are required to erupt in order to emplace the smooth features observed. In the second case of deeper subsidence, approximately $2.5 \times 10^8$ to $2.7 \times 10^9$ m$^3$ of cryolava are required. These volumes depend slightly on the cryomagma composition, but this does not change the
results trend. These extreme values may be better constrained given the knowledge of the subsidence height under the smooth features, but unfortunately, it is not possible to infer it. Larger reservoirs are expected to create deeper subsidence of the surface (Manga and Michaut 2017), so it is worth considering both of these solutions.

After our previous model (Lesage et al. 2020), we can predict the cryolava volume erupted at the surface at the end of a cryovolcanic eruption as a function of the reservoir volume and depth. These volumes are compared to the volumes measured on the four Galileo images. We found that a $10^9$ to $10^{11}$ m$^3$ (0.6 to 2.8 km radius) liquid reservoir is required to explain the emplacement of these smooth features from a single eruption in the case of shallow subsidence. For deep subsidence of 40 m, the required reservoir volumes are 10 times higher (reservoirs up to 6 km in radius). In the case of cyclic eruptions from the same reservoir, for a pure water reservoir and at the end of its activity lifetime, a $4 \times 10^8$ to $4 \times 10^9$ m$^3$ (0.5 to 1 km radius) liquid reservoir is needed in case of shallow subsidence, and a $2 \times 10^9$ to $2 \times 10^{10}$ m$^3$ (0.8 to 1.6 km radius) one in case of a 40 m subsidence. The cryomagma composition slightly changes these results: a two to three times higher reservoir volume is required for a briny cryomagma than for pure water (up to 2.4 km radius). The total lifetime of such reservoirs ranges from $5 \times 10^3$ to $10^5$ years.

The recent detection of chlorides such as NaCl (Trumbo et al. 2019) or Mg-bearing chlorinated species (Ligier et al. 2016) on Europa’s surface may indicate that the cryomagma composition could be different from the one used here (81 wt% H$_2$O + 16 wt% MgSO$_4$ + 3 wt% Na$_2$SO, see Kargel 1991). Two main quantities might be impacted by the cryomagma composition: (i) the freezing time-scale of the reservoir and (ii) the erupted volume at each eruption. The freezing time-scale mainly depends on the eutectic temperature of the solution, which is slightly lower for chloride aqueous solutions than for sulfate ones (Quick and Marsh 2016). Thus, the freezing time scale must be slightly larger for chloride solutions. On the other side, the erupted volume of cryomagma depends on the density contrast between the cryomagma solution and the corresponding ice. Here we tested a sulfate-based cryomagma and the extreme case of pure water. The erupted volumes obtained with these two compositions are quasi identical for a single eruption and differ of at most a factor two for several eruptions during the whole reservoir lifetime. Hence, the cryomagma composition is not expected to modify the results’ trend, at least for a reasonable salt content. Nevertheless, Eq. (5) provides a very simple way to predict the cryomagma volume erupted from a subsurface reservoir depending on the cryomagma and ice densities. It is then possible to adapt this model to any cryomagma composition.
We demonstrated in this study that cryomagmatic reservoirs of $\sim 10^{7}$ to $10^{12}$ m$^3$ located a few kilometers under Europa’s surface may possibly be at the origin of the smooth features seen on the Galileo images 5452r, 0713r, 0739r and 9352r. This information could help the two upcoming missions JUICE (ESA) and Europa Clipper (NASA) to select the interesting zones to target in priority to search for biosignatures. In order to link the volume of the observed smooth features with the characteristic size and depth of the source reservoir it would be necessary to precise the amount of subsidence of the pre-existing terrain. Higher resolution images expected from the future missions will provide more precise DEMs of the surface and more information on the sub-surface. This would help to better constrain the putative cryomagma reservoirs’ dimensions.

6. ISIS pre-treatment

ISIS tool needs pre-processed and calibrate images to work on. After a raw Galileo image and its label text file are downloaded from the PDS data volumes, the first step is to make them usable by ISIS in converting them in a .cub file, which is the ISIS standard format. A .cub file contains the image and the data from the label text file. Then, it is necessary to add the SPICE data to the cub file using ISIS, which contains all the information linked to the spacecraft and imager position and orientation at the moment of the shot. To enhance the quality of the image, we use the ISIS calibration tool “gllssical” plus the “trim” function which corrects the noise visible on the outermost pixels of the image. Finally, the image is map-projected with the “cam2map” tool.

7. Uncertainties estimation

Here we estimate the uncertainties on the DEMs as a function of the parameters used to generate it, in form of a sensitivity study. Unfortunately, absolute DEM uncertainties are not possible to calculate on Europa, since there is no ground truth. We also tried to produce DEMs using the photogrammetry technique but the angle of separation and/or the illumination condition did not allow us to generate a DEM, at least for the images chosen here. Nevertheless, thanks to shadows, it is possible to measure the height of some features and use it as a comparison with the DEM.

From Eq. (1) in the main article, we identify two major factors that could modify the produced DEM depending on the user’s choices: $\mu$ the smoothness parameter and $R(h(x,y))$ the reflectance model and the associated parameters ($\lambda$, the a priori weight, is fixed as $\lambda = 0$ and so it has no importance on the produced
DEM). These parameters have to be optimized and will determine the uncertainties we have on the DEM produced. The image resolution also seems to have an effect on the DEM produced, so we also tested the importance of this bias. Finally, we generate two DEMs of the same terrain seen on two different images to test the process repeatability.

We choose the image 0526r (see Fig. 11) from Galileo SSI to perform all these tests because of the heterogeneity of the terrains and the high variety of objects visible on it. This image does not exhibit cryovolcanic features, but objects with different scales, which is useful to test the parameters available with the SfS tool. To test the validity and the quality of the resulting DEMs as a function of the parameters used, we need to have an idea of the height of the objects seen on the DEMs. In order to produce the most accurate DEMs, the shadow measurement technique is used for each image in this study. We compare the heights measured with the shadows visible on one image to the heights on the corresponding DEM and optimize the smoothness parameter to fit it.

7.1. Measurement of heights with shadows

To measure the approximate height of some reference objects that we choose, we use the shadows visible on the image. An object of height \( h \) projects a shadow of length \( l \) (in the direction of the sun lightning) with:

\[
h = l \tan(\alpha)
\]  

(10)

where \( \alpha \) is the solar elevation.

We measured the shadow of objects of three different sizes on image 0526r to compare them to our “test” DEMs (see Fig. 11): the greatest double ridge at south, a medium-scale crater and a small double ridge at west. We obtain a height of \( 170 \pm 30 \) m for the part of the big double ridge showed in Fig. 11, a depth of \( 120 \pm 20 \) m from the crater rim to its center, and a height of \( 50 \pm 20 \) m for the small double ridge. These values are compared to the topographic profiles obtained during the optimization tests detailed hereafter.

7.2. Smoothness parameter

The parameter \( \mu \) weights the smoothness of the final DEM. In ASP, the default value of \( \mu \) is 0.04, but the smoothness effect is highly dependent on the target surface properties (Alexandrov and Beyer, 2018) and it may be necessary to test several values of \( \mu \) to find an acceptable one, i.e. one that generates a non-noisy image that still shows the low-resolution details. To test the smoothing effect, we computed the DEM of
Figure 11: Image 0526r from Galileo SSI. The lightning direction is indicated with the yellow arrows. We use the shadows in the crater and near the two double ridges showed by the blue arrows in order the measure the height of these three objects.
image 0526r using different smoothness parameters $\mu$ ranging from $10^{-3}$ to $10^2$. We obtained the DEM showed in Fig. 12a with $\mu = 1.6$. Fig. 12b, 12c, and 12d are topographic profiles of the three objects selected previously to measure their height obtained with six values of $\mu$ ranging from $10^{-3}$ to $10^2$.

As expected, the DEM obtained is sensitive to the parameter $\mu$ chosen. We observe that a too high value such as $\mu \gtrsim 10$ leads to a flattened DEM at every wavelength as it is shown on Fig. 12c and 12d, which is the expected effect of smoothing. Moreover, surprisingly, a low value such as $\mu \lesssim 1$ also flattens the high-wavelength reliefs, as one can see on profile (b) from Fig. 12. Indeed, it seems that a too low value of $\mu$ does not allow to integrate an accurate shape of the terrain from the facet slopes calculated. To avoid these two extreme effects, we need to use a parameter $\mu$ around 1 or 2 in the case of image 0526r. Here, we take $\mu = 1.6$, a value for which we obtain the DEM the most consistent with the heights measured previously (see sec. 7.1).

We also noticed that some arbitrary values of the smoothing parameter, such as $1 < \mu < 1.5$ or $\mu = 2$ for image 0526r, produce a very noisy DEM, with artifacts propagating in the direction of the sunlight. This kind of wavy noise generated by the ASP is interpreted by [Jiang et al. (2017)] as an indeterminacy on a facet normal around the Sun’s direction. In some cases, the intensity of a facet only depends on the local incidence angle, which is not sufficient to determine the orientation of the facet normal around the sun’s direction. Because of this indeterminacy, the facet orientation is likely to be biased, and this bias can propagate across the DEM which generates the wavy noise observed. We cannot explain why this effect is produced with some specific values of the smoothness parameter, neither predict these values. Nevertheless, we can easily detect this kind of noise and choose another value of $\mu$ to generate our DEMs.

The several tests conducted on each image show that it is not possible to use the same value of the smoothness parameter $\mu$ for all the images. It is necessary to adjust this value for each image. It is also not possible to guess the most optimized value before conducting tests. To produce each DEM used here, we search for a value of $\mu$ which produces a non-noisy DEM, and which avoids the flattening effects. To select the most appropriate value of $\mu$ for one image, we measure the approximate height of at least two different objects using the shadows technique (see section 7.1) to use it as reference heights. Then we generate several DEMs with different values of $\mu$ and compare them with the reference heights. We keep the value of $\mu$ that generates the most consistent DEM with the shadow measurements. The value of $\mu$ selected for each image is given in section 8.

We demonstrated that the smoothness parameter affects the DEM topography, and as in this study we
focus on volume estimation, we thus need to estimate the uncertainty on these measures coming from the
smoothness parameter. It is not sufficient to look at the uncertainty on the elevations since an object could
have the same volume on two DEMs but a different mean elevation. It is thus necessary to look at the
objects cross-sections as a function of the smoothness parameter. Fig. 12a and 12d, show respectively the
cross-section of the left-side of the double-ridge, and the cross-section of the small crater, that are respectively
the greatest and the smallest objects visible on this zone. To have an idea of the uncertainty on the measured
volumes associated to the smoothness parameter, we estimate the uncertainty on the cross-sections (as an
example, two are showed in grey in Fig. 12b and 12d, and where obtained with $\mu = 1$). We estimate cross-
sectional areas with $\mu$ ranging from $10^{-3}$ to 10 (note that $10^2$ is excluded since the resulting DEM is too
smooth). We use the maximum and minimum cross-sectional areas to calculate the percentage deviation of
the mean, representing the uncertainty:

$$\frac{\Delta CS}{CS} = \frac{CS_{\text{max}} - CS_{\text{min}}}{\frac{CS_{\text{max}} + CS_{\text{min}}}{2}}$$

(11)

where $CS$ is the cross-sectional area. To infer the uncertainty on the volumes from the uncertainty on the
cross-section area, we should also take into account the uncertainty on the feature limits that we determine
with the relation:

$$\frac{\Delta V}{V} = \frac{\Delta A}{A} + \frac{\Delta H}{H}$$

(12)

where $V$ is the feature volume, $A$ is the feature area and $H$ is the feature volume. $\frac{\Delta CS}{CS}$ gives the uncertainty on
the heights $\Delta H$ of the object but does not take into account the uncertainty on the object area. Nevertheless,
$\frac{\Delta A}{A}$ is expected to be minor compared to $\frac{\Delta H}{H}$ since the features are well resolved on the image. Thus, we
consider that $\frac{\Delta V}{V} = \frac{\Delta CS}{CS}$.

We apply Eq. (11) to the double ridge and the small crater from the DEM of image 0526r: for the left
side of the double ridge, we measure a maximum cross-section of 239,000 m$^2$ for $\mu = 1.6$ and a minimum
cross-section of 174,000 m$^2$ for $\mu = 0.1$, which leads to an uncertainty of $\pm 15\%$. For the small crater, the
cross-section is a maximum for $\mu = 1.6$ with an area of 33,000 m$^2$ and a minimum for $\mu = 0.1$ with an area
of 29,500 m$^2$, which gives an uncertainty of $\pm 5\%$.

Topographic profiles on Fig. 12a, 12b and 12c show that the smoothness parameter has a greater effect
on the big objects than on the small ones. This effect is seen on every DEM we produced: smallest structures
of each DEM are quasi not impacted by the smoothness parameter chosen (as long as we take a reasonable
value). As we are interested on small objects, the volume uncertainty coming from the choice of $\mu$ should not exceed $\pm 5\%$.

7.3. Reflectance model

The SfS tool can be used with three different reflectance models, that are Hapke, Lambertian, and Lunar Lambert, and for which the user can select the parameters. The DEMs used here are generated using the Hapke model, which is the most commonly used to model the surface of the icy moons (Belgacem et al., 2020). Nevertheless, we tested the influence of the model and parameters selected on the resulting DEM.

The Lambertian model is not suitable for the icy surfaces, because it generates a lot a wavy noise on the DEMs (Jiang et al., 2017). We focus on the Hapke and Lunar Lambert models. We test here the influence of the model and albedo parameter used. We generated the same DEM of image 0526r in four different cases: a Hapke model with an albedo of 0 and 1, and a Lunar Lambert model with an albedo of 0 and 1. The other parameters are let to the default values since we did not observe a significant difference with other photometric parameters. We compare in Fig. 13 the topographic profiles obtained for cuts AB, CD, and EF (see Fig. 12).

Surprisingly the model and albedo used have a very minor influence on the DEM obtained. At most, one can observe a relative difference of a hundred meters for the highest object (double ridge shown on profile Fig. 13a). However, in this study, we only measure relatively small features, and Fig. 13b and 13c show that the reflectance model and the albedo only modify these objects relative height by a few tens of meters at most. For this reason, we decided to choose the most commonly used model in the literature, which is the Hapke model with an albedo of 0.9 (Belgacem et al., 2020).

The reflectance model effect must be considered in the uncertainties calculation. As above, we calculate the uncertainty on the cross-sections on the left side of the great double ridge, and on the small crater (grey areas in Fig. 12b and 12d). For the left side of the double ridge, we have a maximum cross-section of 239,000 m$^2$ for the Hapke model with $A = 1$ and a minimum cross-section of 121,000 m$^2$ for the LunarLambert model with $A = 1$, which gives an uncertainty of $\pm 33\%$. For the small crater, the cross-section is maximum for the Hapke model with $A = 0$ with an area of 33,000 m$^2$ and minimum for the LunarLambert model with $A = 1$ with an area of 27,000 m$^2$, which gives an uncertainty of $\pm 10\%$. Again, topographic profiles from Fig. 13 show that small objects are much less impacted by the photometric model. We limit our study to low elevation objects so the volume uncertainty is under $\pm 10\%$. 
Figure 12: (a) DEM of image 0526r from Galileo SSI with three topographic profiles of: (b) a high and wide double ridge, (c) a little double ridge and (c) a medium-sized crater. The DEM has been generated using six different values of the smoothness parameter $\mu$ ranging from $10^{-3}$ to $10^2$. DEM presented in (a) is calculated using $\mu = 1.6$. 
Thus, the cumulated uncertainties due to the smoothness parameter and photometry is, at worst, ±15% on the volumes.

7.4. Resolution

Even though the image resolution is not a parameter chosen by the user, it may have an effect on the produced DEM. To test this effect, we changed the resolution of image 0526r by reducing its number of pixels. An example is given in Fig. 14a where the DEM is produced with an 8 times lower resolution than the original one. We tested resolution factors ranging from 1 to 1/10.

We plotted the topographic profiles AB and EF (see Fig. 14b and 14c) to measure the cross-section of the double ridge and the small crater as functions of the resolution factor. Globally, we noticed that the cross-section of these objects becomes higher at low resolution. We found cross-section ranging from 185,000 to 270,000 m² for the left side of the double ridge, which gives an uncertainty of ±19%, and from 29,000 to 67,000 m² for the small crater, which gives an uncertainty of ±40%. This time, we can see that small objects are more affected by the resolution loss. Nevertheless, for reasonable resolution factors above 1/8, we can see in Fig. 14b and 14c that topographic profiles are not affected so much by the resolution change: we only have an uncertainty of ±2% on the small crater for a resolution factor between 1 and 1/4.
7.5. **Repeatability**

We finally test the repeatability of the process tool by generating DEMs of overlapping images and we compare the products obtained. The two images used (0526r and 0539r) have been taken during the same orbit, under the exact same conditions. The DEMs are computed with the same input parameters: the Hapke model with $\omega=0.9$, $b=0.35$, $c=0.65$, $B_0=0.5$ and $h=0.6$, the smoothness parameter $\mu=1.6$ and the initial DEM constraint weight $\lambda=0$. In Fig. 15, we show DEMs generated with two overlapping images 0526r and 0529r and compare two topographic profiles from them.

Fig. 15 shows that the DEM of image 0539r (Fig. 15b) is much noisier than the DEM of image 0526 (Fig. 15a). DEM of image 0539r presents the wavy noise evoked in section 7.2 and usually avoid by selecting an appropriated smoothness parameter. This shows that even for very similar images taken under exactly the same conditions, the DEM production needs to be optimized by selecting an appropriate smoothness parameter, certainly due to roughness differences between the two images.

We calculate the cross-sectional areas of the left side of the double ridge and of the small crater as previously. For the double ridge, we obtain a cross-section area of 253,000 m$^2$ for image 0526r and a cross-sectional area of 228,000 m$^2$ for image 0539r, which easily fits with a ±15% uncertainty. For the small crater, we obtain a cross-sectional area of 27,900 m$^2$ for image 0526r and a cross-sectional area of 28,300 m$^2$ for image 0539r, which again is well under the uncertainties we predicted. This validates the uncertainty range.
chosen. Moreover, this example confirms that the parameters chosen (µ and the reflectance model) have a minor effect on small objects.

8. DEMs produced and volume interpolation

In this section we show the four DEMs generated with SfS and used to measure the smooth features volumes.

8.1. Image 5452r

Image 5452r (see Fig. 16 left) is a well-known Europa’s surface image showing a circular smooth feature. Even if it is not obvious on the raw image, a few noisy pixels on the right side of the smooth feature were propagating noise on it, this is the reason why we cropped the image to generate the DEM (see Fig. 16 left). We choose the value µ = 0.02 to generate this DEM. Indeed, µ > 0.02 was leading to a noisy DEM. With the shadow measurement, we find that the small ridge crossing the feature at south-east is ∼ 20 m high, which is consistent with the calculated DEM. The measure of the volume of the smooth feature delimited by the blue line gives (5.7 ± 0.9) × 10^7 m^3.

8.2. Image 0713r

Image 0713r is showed in Fig. 17. On this image, one can see a smooth feature relatively young, crossed by only two fractures. The double ridge crossing the image north-west of the smooth feature is approximately 40±10 m high, which is also the value found on the DEM generated with µ=0.01. We calculated the volume of the smooth feature delimited by the blue line in Fig. 17 (right) and found a volume of (6.6±1.0) × 10^7 m^3.

8.3. Image 9352r

Image 9352r exhibits a smooth feature that seems younger than all the surrounding ridges and faults (see Fig. 18). DEM showed in Fig. 18 is obtained with µ=0.01. The high-elevation angular object south-west of the smooth feature is approximately 120±20 m after a shadow measurement and ∼80m high on the DEM, but the other smoothness parameters give a noisy DEM. This should not affect the small objects topography, as shown previously (see section 7). The measure of the volume of the smooth feature delimited by the blue line on Fig. 18 gives (4.2±0.6)×10^8 m^3.
Figure 15: DEMs of (a) image 0526r and (b) the overlapping image 0539r from Galileo SSI data. Topographic profiles from these two DEMs are obtained from cuts (c) of the double ridge and (d) of the small crater. Topography from image 0526r is plotted in black, topography from image 0539r is plotted in grey.
Figure 16: Image 5452r from Galileo SSI (left) and the associate DEM produced with SfS (right). The flow-like feature is delimited by the blue line, and the black crosses show the points taken to interpolate a pre-existing terrain beneath the smooth feature.
Figure 17: Image 0713r from Galileo SSI (left) and the associate DEM produced with SIS (right). The flow-like feature is delimited by the blue line, and the black crosses show the points taken to interpolate the pre-existing terrain.
Figure 18: Image 9352r from Galileo SSI (left) and the associate DEM produced with SIS (right). The flow-like feature is delimited by the blue line, and the black crosses show the points taken to interpolate the pre-existing terrain.
8.4. Image 0739r

Image 0739r (see Fig. 19 left) exhibits a smooth feature, with a small lobate elevated object on it (see the white arrow on Fig. 19 right). The smooth feature seems to be crossed by several double ridges, and more precisely by a big double ridge north-west of the feature and a small one south-east of the feature, but it is hard to tell if the smooth material expands on the other side of these ridges. For this reason, we only take into account the smooth material contained between these two ridges.

From shadow measurements, the big double ridge north-west of the smooth feature is expected to be 70±10 m high, and the small double ridge south-east of the smooth feature is expected to be 35±10 m high. The DEM showed in Fig. 19 (right) is generated with \( \mu=0.5 \). On this DEM we can measure heights of respectively 60 and 30 m for the big and the small double ridge, which values are similar to the ones measured from the shadows. We finally find a volume of \((2.7±0.4) \times 10^8\) m\(^3\) for the smooth feature.
Figure 19: Image 0739r from Galileo SSI (left) and the associate DEM produced with SfS (right). The white arrow shows a lobate feature on the smooth feature. The flow-like feature is delimited by the blue line, and the black crosses show the points taken to interpolate the pre-existing terrain.
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