Dynamics of Sill and Laccolith Emplacement in the Brittle Crust: Role of Host Rock Strength and Deformation Mode

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Abstract

Igneous intrusions in sedimentary basins exhibit a great diversity of shapes from thin sheets (e.g., dikes, sills and cone sheets) to massive intrusions (e.g., laccoliths and plugs). Presently, none of the established models of magma emplacement have the capability to simulate this diversity because they account for either purely elastic or purely plastic or purely viscous host rocks, whereas natural rocks are complex visco-elasto-plastic materials. In this study, we investigate the effects of elasto-plastic properties of host rock on magma emplacement using laboratory experiments made of dry granular materials of variable cohesion. Our results show how the deformation mechanism of the host rock controls the emplacement of magma: thin sheet sills form in high-cohesion materials, which dominantly deform by elastic bending, whereas massive intrusions such as punched laccoliths form in low-cohesion materials, which dominantly deform by shear failure. Our models also suggest that combined elastic/shear failure deformation modes likely control the emplacement of cone sheets. Our experiments are the first to spontaneously produce diverse, geologically relevant intrusion shapes. Our models show that accounting for the elasto-plastic behavior of the host rock is essential to filling the gap between the established elastic and plastic models of magma emplacement, and so to reveal the dynamics of magma emplacement in the Earth’s brittle crust.

1. Introduction

Igneous intrusions in the upper brittle crust exhibit diverse shapes ranging from thin sheets, such as dikes, sills, saucer-shaped sills, and cone sheets, to thick, massive intrusions such as laccoliths, baysmaliths, plutons, and plugs (Figure 1) (e.g., Breitkreuz & Petford, 2004; Bunger & Cruden, 2011; Delpino et al., 2014; Galland et al., 2017; Gilbert, 1877; Habert & de Saint-Blanquat, 2004; Phillips, 1974). Each intrusion shape reflects a distinct mode of magma emplacement (Galland et al., 2017). Such diversity illustrates the complex physical behavior of magma-host rock systems during magma emplacement, mostly as a result of the wide range of magma viscosities (100–1018) (Dingwell et al., 1993; Scaillet et al., 1997) and the wide range of elasto-plastic rheologies of the host rock (Galland et al., 2017; Ranalli, 1995).

To date, two end-member types of magma emplacement models exist. On one hand, models for sheet intrusions consider “relatively low,” often neglected, viscosity magma into purely elastic rock (Bunger & Cruden, 2011; Galland & Scheibert, 2013; Kavanagh et al., 2015; Malthe-Sørenssen et al., 2004; Michaut, 2011; Pollard & Johnson, 1973; Thorey & Michaut, 2016). On the other hand, models for massive intrusions consider very viscous magma intruding into a weak, usually plastic host rock (e.g., Brothelande et al., 2016; Merle & Borgia, 1996; Montanari et al., 2010; Roman-Berdier et al., 1995). Field observations (de Saint-Blanquat et al., 2006; Spacapan et al., 2017; Wilson et al., 2016) and seismic data (Jackson et al., 2013; Schmiedel et al., 2017, and references therein), however, indicate that both elastic and plastic deformation contemporaneously accommodate magma emplacement. In addition, plastic (Haug et al., 2017) and elasto-plastic (Scheibert et al., 2017) models suggest that inelastic deformation of the host rock might significantly affect the dynamics of magma propagation, even if inelastic deformation is restricted to an area of negligible size compared to that of the intrusion. Therefore, the end-member emplacement models are oversimplified with respect to natural rock behavior. As a consequence, none of them are able to reproduce the diversity of intrusion shapes and the modes of host rock deformation that accommodate their emplacement.

We present laboratory experiments of magma emplacement in realistic elasto-plastic brittle crust, which is modeled using dry granular materials of variable cohesion. Depending on the cohesion of the model rock, the modeled intrusion shapes and host rock deformation modes ranged from thin, sill-like sheets to...
massive, punched laccoliths, and from gentle bending to prominent shear failure, respectively. Our experiments, the first that spontaneously simulate both sheet and massive intrusions, bridge the existing end-member magma emplacement models and highlight the importance of the host rock strength and associated deformation mode.

2. Experimental Methods

We used dry Coulomb granular materials of variable strengths as model rock (Table 1) (Abdelmalak et al., 2016; Galland et al., 2006): (1) a high-cohesion (~600 Pa) end-member consisting of fine-grained crystalline silica flour (SF), simulating competent rocks such as limestone; (2) a low-cohesion (~100 Pa) end-member consisting of glass microspheres (GMs), simulating weak rocks such as shale; and (3) materials with intermediate cohesion consisting of homogeneous SF-GM mixtures with different proportions (Table 1). The cohesion of the SF-GM mixtures is linearly correlated with their mixing proportions (Abdelmalak et al., 2016). The model magma consists of molten vegetable oil of viscosity $\eta = 2 \times 10^{-2} \text{ Pa s}$ at 50°C (Galland et al., 2006).

The relevance of these materials and the scaling of the models are described in detail by Galland et al. (2014), Galland et al. (2006), and Abdelmalak et al. (2016) and in section 4.2.

The preparation procedure was the same as that of the layered models of Galland et al. (2009). The experiments were contained in a 40 × 40 cm box (Figure 2). They consisted of two layers of the same compacted granular material with a density difference of <5% (Galland et al., 2009). For the compaction procedure we first measured a mass of granular material for the lower layer (final thickness of the layer is equal to the inlet height), which we compacted using a high-frequency compressed-air shaker (Houston Vibrator, model GT-25). Subsequently, we placed a flexible net onto the surface of this first layer, and poured a second mass of granular material. The whole system was compacted until reaching the desired thickness of the overburden above the inlet (0.03 m). This procedure allows for a homogeneous, repeatable, and fast compaction of the granular materials, and ensures a good control of their density (<5% uncertainty), and so of their cohesion.

### Table 1

**Overview of the Experimental Series and Materials**

<table>
<thead>
<tr>
<th>Experiment no.</th>
<th>Model rock materials (GM/SF&lt;sup&gt;a&lt;/sup&gt; wt %)</th>
<th>Compacted density (kg/m&lt;sup&gt;3&lt;/sup&gt;)</th>
<th>Cohesion (Pa)&lt;sup&gt;b&lt;/sup&gt;</th>
<th>Experiment duration $t_e$ (s)</th>
<th>Intrusion shape</th>
</tr>
</thead>
<tbody>
<tr>
<td>Experiment 1</td>
<td>0/100</td>
<td>1100</td>
<td>559.91</td>
<td>101</td>
<td>Saucer-shaped sill</td>
</tr>
<tr>
<td>Experiment 2</td>
<td>50/50</td>
<td>1450</td>
<td>374.96</td>
<td>29</td>
<td>Saucer-shaped sill</td>
</tr>
<tr>
<td>Experiment 3</td>
<td>80/20</td>
<td>1605</td>
<td>313.1</td>
<td>7</td>
<td>Cone sheet</td>
</tr>
<tr>
<td>Experiment 4</td>
<td>90/10</td>
<td>1620</td>
<td>239.5</td>
<td>21</td>
<td>Laccolith</td>
</tr>
<tr>
<td>Experiment 5</td>
<td>100/0</td>
<td>1435</td>
<td>142.3</td>
<td>18</td>
<td>Laccolith</td>
</tr>
</tbody>
</table>

*Note. Galland et al. (2006) determined the angle of friction for GM and SF with 26° and 39°, respectively.*

<sup>a</sup>GMs = glass microspheres/SF = silica flour.  
<sup>b</sup>After Abdelmalak et al. (2016).
The flexible net simulated a weak interface between sedimentary layers, but quantifying its mechanical effects is challenging. We assume that the net approximately simulates strength contrasts in layered rocks, which can be as high as several orders of magnitude (Hoek et al., 1998; Schellart, 2000). The horizontal extent of the net in the experiments was significantly larger than the final intrusion diameters; therefore, the edges of the net had no effects on oil emplacement.

A volumetric pump injected the vegetable oil at constant flow rate ($40 \text{ mL min}^{-1}$), which corresponds to an average inlet velocity of $v_i = 0.026 \text{ m s}^{-1}$ through a $d = 5 \text{ mm}$ diameter inlet. We designed the experiments such that the net is located just at the inlet. This way the oil flows directly along the net; therefore, the experiments do not simulate the feeder-sill transition, and the inlet in the experiments is not to be compared to sill feeders in nature. The oil initially followed the flexible net as a stratigraphic weakness (resembling an initial sill stage) before propagating upward. The intrusion of the oil triggered deformation of the experimental surface, which was monitored using a photogrammetry setup consisting of four synchronized DSLR cameras (Figure 2) (Galland et al., 2016). This system produced high-resolution ($<0.1 \text{ mm}$) and high-precision ($\sim0.05 \text{ mm}$) topographic maps and horizontal displacement maps. The monitoring frequency was 1 Hz. The end of the experiment ($t_e$) coincided with the eruption of oil at the surface (Table 1). After the end of the experiment, the oil solidified, the resulting intrusion was excavated, and its shape was computed using the same photogrammetry system (Galland et al., 2016). Thus, this photogrammetry system allows direct comparison between the intrusion shape and the associated surface deformation. During cooling, the oil percolates into the granular host material. Visual observations show that this percolation takes many minutes; i.e., it takes a longer time than the duration of the experiments. We infer that the oil percolation is negligible and has no effect during the emplacement of the oil. Note that our models do not take into account the thermal effects of magma emplacement on the host rock rheology. These effects have been documented by Karlstrom et al. (2017) for the incremental growth of large, deep plutonic complexes. However, given that our models intend to simulate the emplacement of relatively small intrusions in the shallowest, cold parts of the brittle crust, we consider these mechanisms to be negligible.

In this paper, we present a series of experiments in which the cohesion of the model rock was variable from high-cohesion (100% SF) to low-cohesion (100% GM), with intermediate-cohesion mixtures (Table 1). Both the injection depth (3 cm) and the injection flow rate ($40 \text{ mL min}^{-1}$) were kept constant. The dimensional analysis of the models and their similarity to their geological prototypes are discussed in section 4.2.

### 3. Model Results

In experiment 1 (high cohesion, 100% SF; Table 1), the resulting intrusion was a typical saucer-shaped sill, with a circular inner sill of diameter $\approx10 \text{ cm}$ (Figure 3) emplaced along the net, connected outward to gently dipping ($15^\circ–35^\circ$) inclined sheets that flattened out to outer sills (Figure 3) (Polteau et al., 2008). The associated surface deformation in Figure 3a exhibited a wide, low-amplitude (amplitude-to-width ratio $h/w < 0.03$; Figure 4) bell-shaped dome, which exhibited gradual slope variations (maximum slopes $<3^\circ$), except locally prior to oil eruption at the outer margin of the dome (Figure 3a). At the end of experiment 1, the horizontal extent of the dome correlated with that of the underlying intrusion (Figure 3).

In experiments 2 and 3, the cohesion of the model rock was lower than in experiment 1 (Table 1). The resulting bodies were also saucer-shaped sheet intrusions. However, with decreasing cohesion the diameter of the inner horizontal sill decreased (5–4 cm; Figure 3b), and the dip angle of the inclined sheets increased (Figure 3b). In experiment 3, the inner sill was so small that the intrusion can be defined as a cone sheet...
Figure 3. (a) Experimental results of the overburden deformation. (top row) Ortho-rectified images of the final model surface. (middle row) Topographic maps of the final model surface. (bottom row) Topographic profiles of the surface uplift for three time steps. The experiment duration is given by $t_e$. White lines in topographic maps locate the profiles of the profiles and plots (note the different vertical scales in topographic maps and profiles). (b) Experimental results on the intrusion geometry. Topographic maps of the top surface of excavated intrusion in our experiments, from high-cohesion host (crystalline silica flour; experiment 1) to low-cohesion host (glass microspheres; experiment 5). (top row) Red lines locate the profiles displayed in bottom row. (bottom row) Plots of profiles across intrusions displayed in top row maps.
Experiment 4 and 5 (i.e., low cohesion) showed drastically different results. The intrusions were massive, not sheet intrusions. They exhibited roughly vertical plug-like shapes, with flat ~3 cm wide bottom and subvertical sidewalls (Figure 3b); i.e., the intrusions resembled punched laccoliths as shown in Figure 1 (Corry, 1988). The surface deformation displayed high-amplitude domes \( h/w < 0.45 \) with sharp, steep (\( >45^\circ \)) margins (Figure 3a). The tops of the domes were blocky (Figure 3a) and some blocks collapsed during the dome growth. After the start of eruption, we stopped the pump, and the dome kept collapsing as the oil was draining out from the underlying intrusion.

**4. Interpretations and Discussion**

**4.1. Interpretation**

Our results highlight a trend between the intrusion shapes and the cohesion of the overburden (Figure 3). The intrusions emplaced into the high- to intermediate-cohesion model rock (experiments 1–3) are sheet intrusions, whereas the intrusions in the low-cohesion model rock (experiments 4 and 5) are massive laccolith intrusions (Figure 3b). Even among the sheet intrusions our results show a gradual trend from wide, gently dipping saucer-shaped sills in experiment 1 (high cohesion) to narrow, steeply dipping cone sheet in experiment 3 (Figure 3b). In addition, in all experiments the injection occurs at the same depth \( H = 3 \text{ cm} \); however, the diameter \( L \) of the intrusions and so the aspect ratio \( H/L \) strongly vary as a function of the cohesion of the overburden: intrusions are narrow \( (H/L < 1) \) for laccoliths emplaced in low-cohesion host rock, whereas intrusions are wide \( (H/L > 1) \) for sills emplaced in high-cohesion host rock. These results clearly show that host rock strength is a key factor controlling magma emplacement.

An trend emanates between uplifted surface morphology (dome) and the cohesion of the overburden (Figures 3a and 4). In the high-cohesion experiment (experiment 1), the dome exhibits gentle slopes (1.8° to 2.7°) and curvatures and small \( h/w \) aspect ratios (Figure 4). We infer that the overburden dominantly deformed by elastic bending, as considered in theoretical (Bunger & Cruden, 2011; Galland & Scheibert, 2013; Michaut, 2011; Pollard & Johnson, 1973; Scheibert et al., 2017) and laboratory (Kavanagh et al., 2015; Pollard & Johnson, 1973) elastic models. An exception is the region near the point of eruption, where the dome edge was steeper (Figure 3a) and failure of the experimental surface occurred, similarly to the plastic models of Haug et al. (2017). In contrast, low-cohesion model rock experiments (experiments 4 and 5) feature domes with steep slopes and abrupt surface scarps marked by locally high curvatures (Figure 3a) and drastically higher (1 order of magnitude) \( h/w \) aspect ratios (Figure 4). We infer that the overburden dominantly deformed by shear failure. Finally, in experiments 2 and 3 (intermediate cohesion), the slopes at the surface ranged between those in experiment 1 and experiment 4 (Figure 3a). We infer that the overburden deformed by a combination of elastic bending and local shear failure. In conclusion, the cohesion greatly controls the deformation mode of intrusion host rock.

The points discussed above highlight a relation between the shapes of the intrusions and the deformation mode of their overburden. The massive laccoliths of experiments 4 and 5 are associated with dominant shear (mode II) failure in the low-cohesion overburden, in good agreement with the “punched laccolith” emplacement model (Corry, 1988; Gilbert, 1877) and the end-member plastic models of Roman-Berdiel et al. (1995). In contrast, the flat-lying, saucer-shaped sills of experiments 1 and 2 are associated with dominant elastic bending of the overburden, in good agreement with the end-member elastic models of Galland and Scheibert (2013), Malthe-Sørenssen et al. (2004), and Pollard and Johnson (1973). This suggests that the

![Intrusion type vs. host rock strength](https://example.com/figure4.png)  
Figure 4. Plot of the maximum uplift to diameter ratio \( h/w \) versus host rock strength (cohesion) in our models. Open symbols correspond to sheet intrusions (saucer-shaped sills and cone sheets), and full symbols correspond to massive intrusions (laccoliths). Colored bar to the right side of the plot indicates the natural ranges of \( h/w \) for sills (light green) and laccoliths (orange) estimated from field observations and seismic interpretations (Bunger & Cruden, 2011; Hansen & Cartwright, 2006; Schmiedel et al., 2017, and references therein).
Table 2  
Values and Equations for the Dimensional Analysis

<table>
<thead>
<tr>
<th>Values</th>
<th>Experiments</th>
<th>Nature</th>
</tr>
</thead>
<tbody>
<tr>
<td>( g ) (m s(^{-2}))</td>
<td>9.81</td>
<td>9.81</td>
</tr>
<tr>
<td>( T ) (m)</td>
<td>0.0034–0.018</td>
<td>10–3,000</td>
</tr>
<tr>
<td>( L ) (m)</td>
<td>0.04–0.24</td>
<td>500–10(^5)</td>
</tr>
<tr>
<td>( \rho_m ) (kg m(^{-3}))</td>
<td>890</td>
<td>2,500–2,700</td>
</tr>
<tr>
<td>( \eta ) (Pa s)</td>
<td>0.02</td>
<td>100–10(^7)</td>
</tr>
<tr>
<td>( \nu ) (m s)</td>
<td>0.0072–0.038</td>
<td>0.00034–0.1</td>
</tr>
<tr>
<td>( H ) (m)</td>
<td>0.03</td>
<td>1,000–5,000</td>
</tr>
<tr>
<td>( h ) (m)</td>
<td>0.0034–0.018</td>
<td>10–3,000</td>
</tr>
<tr>
<td>( w ) (m)</td>
<td>0.04–0.24</td>
<td>500–10(^5)</td>
</tr>
<tr>
<td>( \rho_f ) (kg m(^{-3}))</td>
<td>1,100–1,600</td>
<td>2,500</td>
</tr>
<tr>
<td>( C ) (Pa)</td>
<td>100–600 ( \times 10^6 )–10(^8)</td>
<td></td>
</tr>
<tr>
<td>( \phi ) (°)</td>
<td>26–39</td>
<td>25–45</td>
</tr>
</tbody>
</table>

Dimensionless parameters

\[
\Pi_1 = \frac{\nu}{gT^2} \\
\Pi_2 = \frac{\phi}{T} \\
\Pi_3 = \frac{Re}{\eta} = \frac{\rho_f wT^2}{\eta} \\
\Pi_4 = \frac{T}{H} \\
\Pi_5 = \frac{L}{T} \\
\Pi_6 = \frac{h}{T} \\
\Pi_7 = \frac{1 - \frac{L}{T}}{H}
\]

Our experiments are designed to model processes at basin scale; i.e., 1 cm represents ~1 km in nature. The physical parameters ruling magma flow in our experiments are (Table 2) (1) magma velocity in the intrusions \( v \), (2) magma viscosity \( \eta \), (3) magma density \( \rho_m \) and (4) the resulting diameter \( L \) and thickness \( T \) of the intrusion (Figure 1). Note here that the diameter \( L \) and the thickness \( T \) of the intrusion are not prescribed a priori, but they are experimental results. Note well, as mentioned above, that we do not consider the feeder-sill transition; therefore, we do not account for the inlet geometry/size in the dimensional analysis.

The physical parameters governing the brittle host rock and its deformation in our experiments are (Table 2) (1) thickness of the overburden \( H \); (2) rock density \( \rho_r \); (3) cohesion \( C \), i.e., host rock shear strength; and (4) angle of friction \( \phi \). Here the overburden thickness \( H \) corresponds to the injection depth in our experiments. We measure the host rock deformation through the uplift \( h \) and the diameter \( w \) of the dome (Figure 1), which are assumed in our analysis to represent realistic approximate values of the intrusion thickness \( T \) and intrusion diameter \( L \), respectively. Our results indeed confirm that millimeter-scale uplifts in experiment 1 to experiment 3 reflect millimeter-scale sheet intrusion thickness, whereas \( h > 1 \text{ cm} \) uplifts in experiment 4 and experiment 5 reflect \( >1 \text{ cm} \) massive intrusion thickness (Figure 3). Similarly, 20 cm dome diameter reflects the 18–20 cm intrusion radius, whereas the 4 cm dome diameter reflects the 3–4 cm intrusion diameter (Figure 3). Therefore, we consider \( h \approx T \) and \( w \approx L \), so that two parameters are considered in our dimensional analysis instead of four. The whole system is subject to gravity \( g \). Thus, according the Buckingham II theorem (e.g., Gibbings, 2011), 10 variables minus three variables with independent dimensions lead to seven independent dimensionless numbers (Table 2). These seven dimensionless numbers characterize the physics of the modeled processes; our models are similar to its geological emplacement of the magma dominantly occurs by mode I (tensile opening) failure of the host rock. We cannot rule out, however, that local shear failure occurred, e.g., at the inner sill-to-inclined sheet transition (Galland et al., 2009; Haug et al., 2017; Scheibert et al., 2017). In between these plastic and elastic end-members, our experiments suggest that the cone sheet of experiment 3 is associated with a combination of elastic bending and mode II failure of the overburden, which corroborates the conclusions of Phillips (1974) and Galland et al. (2014). In addition, our conclusion is very similar to those of Abdelmalak et al. (2012), who showed that the formation of V-shaped sheet intrusions in 2-D experiments can be controlled by local shear failure of the host rock. All these results highlight the first-order control of host rock damage on magma propagation, which is in good agreement with progressive damage models of volcano deformation in response to inflating magma body (Carrier et al., 2015; Go et al., 2017). Overall, we conclude that the deformation mode of the host rock is a primary parameter governing magma emplacement in the Earth’s brittle crust.

4.2. Dimensional Analysis and Scaling

In this section, we discuss the geological relevance of our laboratory results through the dimensional analysis of the modeled system and its similarity to their corresponding geological objects (Galland et al., 2017; Gibbings, 2011; Merle, 2015). The dimensional analysis is challenging because we need to account for (1) the elastic/inelastic deformation mode of the host rock, (2) the flow regime of the viscous flow, and (3) the coupling between the flowing viscous fluid and the deforming host. Because of the complexity of the physical system, there is no mathematical model from which scaling parameters can be extracted (see e.g., Burger & Cruden, 2011; Michaut, 2011); therefore, we apply the dimensional analysis procedure as described by Gibbings (2011). In addition, the discussion of the similarity of our models to its geological equivalent is not straightforward because of (1) the broad range of magma viscosities and (2) contrasting deformation mechanisms in the host (Galland et al., 2017). The procedure described below follows the main lines of those implemented by Merle and Borgia (1996), Galland (2012), and Galland et al. (2014, 2009).
equivalent if the values of these dimensionless ratios in both experiments and nature are similar. In the following paragraphs, we discuss successively the dimensionless parameters that account for (1) the host rock behavior, (2) the magma flow, and (3) the physical coupling between the host and the magma.

The first dimensionless parameter to scale the brittle host rock is the ratio of gravitational stress to cohesion \( \Pi_1 = \frac{\rho g H}{C_0} \) (Table 2). In our experiments, injection is at 3 cm depth, and the model rock cohesion varies from ~100 to ~600 Pa (Abdelmalak et al., 2016). The values of \( \Pi_1 \) thus range between 0.54 and 4.71. In nature, rocks have an average density of about 2500 kg m\(^{-3}\); their cohesion spans between \( 10^3 \) and \( 10^5 \) Pa (Schellart, 2000, and references therein). In sedimentary basins, intrusion depths are typically between 1 and 5 km. The values of \( \Pi_1 \) in nature thus range between 0.25 and 122.63 (Table 2). The model values of \( \Pi_1 \) are within the range of values of \( \Pi_1 \) in nature; therefore, the brittle deformation regime in our experiments is representative of those of the corresponding geological systems. Note that the values of \( \Pi_1 \) in our models correspond to the lower range of \( \Pi_1 \) in nature (Table 2); thus, our models do not account for gravity-dominated systems in nature (\( \Pi_1 >> 1 \)).

The second dimensionless parameter to scale the brittle host rock is the angle of internal friction \( \Pi_2 = \phi \) (Table 2). The angle of internal friction of the granular materials ranges between 26° and 39° (Abdelmalak et al., 2016; Galland et al., 2006). This is within the range of those measured for natural rocks (Table 2) (Schellart, 2000, and references therein).

The first dimensionless ratio to scale the magma flow regime is the Reynolds number \( Re = \Pi_3 = \frac{\rho v d}{\eta} \), i.e., the ratio between inertial and viscous forces within the fluid. We use \( \Pi_3 \) to describe the flow behavior within the intrusion. Therefore, we need to estimate the magma velocity \( v \) within the intrusion using a simple conversion formula \( v = \frac{d}{T} \), where \( d \) is the thickness of the inlet/feeder. In our experiments, \( v = 0.026 \) m s\(^{-1}\), \( d = 0.005 \) m, and \( T \) is variable from 0.0034 to 0.018 m; therefore, the values of \( Re \) in our experiments are in the range of 1.09–30.63 (Table 2). This shows that the magma flow in the model intrusions is laminar. In nature, the magma density ranges from 2500 to 2700 kg m\(^{-3}\) for felsic to mafic magma, respectively (Galland et al., 2017, and references therein). Magma velocity \( v \) in feeder dikes vary between 0.01 and 1 m s\(^{-1}\) (Battaglia & Bachèlery, 2003; Clemens & Mawer, 1992; Petford et al., 1993; Roman et al., 2004; Spence & Turcotte, 1985), thickness of feeder dikes \( d \) typically vary between 1 and 10 m, and the intrusion thickness \( T \) typically vary between 10 and 3000 m. Therefore, in nature the values calculated for \( Re \) range between \( 0.83 \times 10^{-8} \) and 81,000 (Table 2). Note that the highest values of \( Re \) would correspond to 3000 m thick mafic intrusions with magma velocity of 1 m s\(^{-1}\), which is unlikely. Therefore, a geologically relevant upper bound for \( Re \) may be significantly lower, which indicate that in most common cases, the magma flow within intrusions is laminar. Therefore, the laminar fluid flow in our model intrusions is representative of most magma flow regimes within intrusions in nature.

The second dimensionless ratio to scale the magma intrusion is the resulting geometric aspect ratio \( \Pi_4 = \frac{T}{L} \approx \frac{h}{w} \) of the intrusions. In our experiments, the values of \( \Pi_4 \) range between ~0.014 for sills and ~0.45 for laccoliths (Table 2 and Figure 4), which overlap with the values of \( \Pi_4 \) in nature ranging between 0.0001 and 6 (Table 2). Our model intrusions are thus overall geometrically similar to those in nature (Figure 4).

We consider now the dimensionless ratios that account for the physical coupling between the intrusions and the overburden. The first dimensionless parameter to account for this is the geometric ratio \( \Pi_5 = \frac{H}{L} \approx \frac{h}{w} \), which describes whether the intrusion is larger than its depth (\( \Pi_5 < 1 \), i.e., considered as a large shallow intrusion) or narrower than its depth (\( \Pi_5 > 1 \), i.e., considered as a narrow deep intrusion). Theoretical models of sill emplacement based on thin plate theory require that \( \Pi_5 < 0.1 \) (e.g., Galland & Scheibert, 2013; Michaut, 2011). In our experiments, the values of \( \Pi_5 \) range between ~0.125 for laccoliths and ~0.75 for sills (Table 2). The respective values of \( \Pi_5 \) for our model sills and laccoliths are in the range of their equivalent values in nature (0.01–10), which implies that our model intrusions are similar to their geological equivalent (Table 2). Note that most values of \( \Pi_5 \) are larger than 0.1, which implies that the established theoretical models are generally not applicable to thick sills and laccoliths.

The second dimensionless parameter \( \Pi_6 = \frac{\eta}{C_0} \frac{v}{T} \), which quantifies the balance between the viscous stresses within the flowing magma and the cohesion (strength) of the host rock. Note that this \( \Pi_6 \) ratio is the same as the \( \Pi_2 \) ratio defined by Galland et al. (2014). In our experiments, the values of \( \Pi_6 \) span between \( 1.34 \times 10^{-5} \) and 0.0022 (see Table 2). In nature, the broad range of magma viscosities in sheet intrusions and massive
intrusions due to, e.g., magma composition (mafic-felsic), leads to a wide theoretical range of values of $\Pi_6$ in nature between $1.1 \times 10^{-16}$ and 1 (Table 2) (Galland et al., 2014). Thus, the values of $\Pi_6$ in our models are in the same range as those of $\Pi_6$ in geological systems, showing that our models are similar to their geological equivalent in terms of balance between viscous stresses and host rock strength. Note that our experiments do not consider cooling effects: magma cooling induces viscosity increase which may locally affect the value of $\Pi_6$ in nature (Thorey & Michaut, 2016).

The final dimensionless parameter $\Pi_7 = 1 - \rho_m/\rho$, is the ratio of hydrostatic to lithostatic forces, i.e., buoyancy of the magma (Table 2). In our experiments $\Pi_7$ ranges from 0.19 to 0.44 (Table 2). If ($\Pi_7 < 0$), the magma is heavier than the host rock, and locally negatively buoyant; in contrast, if $\Pi_7 > 0$, the magma is lighter than the host rock, so that the magma is locally buoyant. In nature $\Pi_7$ is neutrally buoyant to negatively buoyant, whereas the oil in our experiments is positively buoyant (Table 2). Therefore, there is a discrepancy between the experiments and the geological systems. However, as sills and laccoliths are (sub)horizontal conduits, the effect of the buoyancy is very small, even negligible in the modeled systems as discussed in previous works (e.g., Galland et al., 2009).

To summarize, the systematic overlap of the values of the $\Pi$ numbers in both model and nature show that the processes modeled in our experiments are similar to magma emplacement processes in nature.

### 4.3. Discussion

As mentioned above, both oil percolation and intrusion drain-out occurred after the end of the experiments. Thus, the shapes of the excavated intrusions do not reflect their exact shapes during their emplacement. Consequently, we cannot use the final excavated intrusions to estimate their in situ aspect ratios ($T/L$) during their emplacement. Alternatively, we assume that the aspect ratios of the domes ($h/w$) measured at the surface of the experiments during the experiments are better estimates of the aspect ratios of the underlying intrusions. In the following paragraphs, we will discuss the aspect ratios $T/L$ of the intrusions in our experiments using the values of the aspect ratios $h/w$ of their associated domes.

Our laboratory models simulate for the first time the emplacement of both thin sheet (saucer-shaped sills and cone sheets) and massive (punched laccolith) intrusions, as illustrated in Figure 1. Our experiments reproduce spontaneously a large range of aspect ratios of natural sills and laccoliths, covering several orders of magnitude without prescribing the intrusion geometry (thin sheet versus massive; Figure 4). These results imply that our models capture the first-order mechanisms of magma emplacement in a realistically complex host rock. This was only possible by accounting for variable complex elasto-plastic mechanical properties of the host rock.

The geometrical aspect ratios $\Pi_4 = T/L$ for laccoliths in our experiments concur with values from field observations and seismic interpretation, i.e., 0.02–1 for laccoliths and thick sills (Figure 4) (Burger & Cruden, 2011; Cruden et al., 2017; de Saint-Blanquat et al., 2006; Delpino et al., 2014; Magee et al., 2017, and references therein; McCaffrey & Petford, 1997; Schmiedel et al., 2015). Thus, we conclude that our models are able to simulate the diversity of observed intrusion geometries in nature. The values of $\Pi_4$ for our model sills (0.014 to 0.02) are in the range, close to the upper bound, of those for sills in nature (0.0001 to 0.05) (Burger & Cruden, 2011; Cruden et al., 2017; Hansen & Cartwright, 2006; Schmiedel et al., 2017). This suggests that our models simulate the emplacement of magmas that are more viscous than mafic magmas, e.g., andesite to rhyolite sills (Breitkreuz et al., 2017, and references therein). Other possibilities are that neither the elastic properties of the silica flour, i.e., it is too soft with respect to natural rocks, nor do the properties of intrusion-induced dilation or compaction of the host material properly simulate those of natural rocks. However, constraining the elastic properties, and so the elastic scaling, of the silica flour is challenging.

The main difference between our models and the established elastic models of sills and laccoliths is the ability of the overburden to fail along shear fractures (Kavanagh et al., 2006; Pollard & Johnson, 1973). Several studies discuss that overburden failure is likely of primary importance for understanding magma emplacement mechanisms (e.g., Abdelmalak et al., 2012; Cruden et al., 2017; Guldstrand et al., 2017; Haug et al., 2017), and our models confirm that the elasto-plastic deformation modes of the overburden play an important role on magma emplacement. Similar to our results, the numerical models of Burger and Cruden (2011) also reproduced the natural range of laccoliths aspect $T/L$ ratios, up to 0.6 (Figures 1 and 4).
However, the mathematical formulation of their model is that of a thin stiff plate; the assumptions of which are (Ventsel & Krauthammer, 2001) (1) the thin plate formulation implies shallow intrusions, i.e., intrusion diameter should be larger than 10 times its depth ($L/H \geq 10$) (criterion used by, e.g., Galland & Scheibert, 2013 & Scheibert et al., 2017), and (2) the stiff plate formulation implies small deflection of the overburden, i.e., the overburden thickness should be 5 times greater than the intrusion thickness ($H/T \geq 5$). Combining the thin plate and stiff plate criteria implies that the intrusion diameter should be 50 times greater than intrusion thickness ($L/T > 50$ or $T/L < 0.02$). This critical value of $T/L$ is 1 to 2 orders of magnitude smaller than those of laccoliths in nature, and so those calculated from the thin, stiff plate models of Bunger and Cruden (2011). Hence, most of the laccolith calculations (>90%) presented by Bunger and Cruden (2011) are outside the mathematical assumptions of their mathematical model, and the resulting large deflections are expected to produce unrealistically large stresses due to local extreme bending of the overburden. The strong mismatch between the critical value $T/L < 0.02$ for the validity of thin stiff plate models and the values of $T/L > 0.1$ of natural laccoliths explains the mismatch between the elastic assumption of the models and the inelastic deformation observed in the field (e.g., Agirrezabala, 2015; de Saint-Blanquat et al., 2006; Spacapan et al., 2017; Wilson et al., 2016). This analysis confirms that the inelastic deformation observed in the field likely plays a significant role during magma emplacement; hence, inelastic deformation should be accounted in magma emplacement models.

Further, our experiments simulate conical intrusion shapes, which resemble cone sheet intrusions described in nature (e.g., Burchardt et al., 2013; Galland et al., 2014; Gudmundsson et al., 2014). Our results show their formation as transition stage between sills and laccolith under intermediate host rock strength conditions (Figure 3). These cone sheet intrusions have never been accounted for in sill or laccolith models. However, seismic data and the field observations show that such features are in fact present in volcanic basins (Burchardt et al., 2013; Gudmundsson et al., 2014; Sun et al., 2014). That means our models address the mechanism of magma emplacement in a more realistic way, simulating both (1) the established end-member shapes of intrusions (sheet and massive) and additionally, (2) the poorly understood occurrence of cone sheet intrusions.

In our experiments the oil viscosity was constant, although magma viscosity plays a major role on magma emplacement mechanism (e.g., Rubin, 1993). However, Galland et al. (2014) showed that the emplacement of magma in the brittle crust is neither governed by the flowing viscous magma nor the deforming host rock alone, but by the mechanical coupling between them as expressed by the dimensionless ratios $\Pi_0$, $\Pi_2$ of Galland et al. (2014); Table 2. This dimensionless number quantifies the ratio between the viscous stresses in the magma and the cohesion of the brittle host rock. This ratio shows that decreasing the cohesion of the host rock, as we did in our experiments, is equivalent to increasing the viscosity of the magma. Our results are corroborated by many field observations, as punched laccoliths (1) are commonly observed in weak rocks and/or (2) are made of magma of dominantly high viscosity. We are aware that the elastic properties of our silica flour and glass beads are unconstrained, such that elastic stresses in our models might not be properly scaled. Thus, we did not quantify the relative elastic or plastic contributions to the deformation of the overburden. We are also aware that the dilation and compaction properties of our granular materials might not be identical to those of natural rocks. Our experiments investigate one aspect only of the magma-host rock systems, and future work should investigate systematically the combined effects of magma viscosity and host rock strength. This will ensure filling the gap between the formerly distinct sheet intrusion and massive intrusion models of magma emplacement.

5. Conclusions

This paper describes the results of laboratory experiments of magma emplacement in the brittle crust with horizontal layering. We performed a parameter study to test the effect of the cohesion (strength) of the model host rock, while the depth of magma injection and magma injection rates were kept constant. The main results of our studies are the following:

1. Our laboratory models simulate—for the first time—the emplacement of both sheet intrusions (thin saucer-shaped sills and cone sheets) and massive intrusions (punched laccoliths) in a controlled manner.
2. Saucer-shaped sills form in high-cohesion host material, punched laccolith intrusions form in low-cohesion host material, and cone sheets form in intermediate-cohesion material.
3. Our models show that the diameter of the intrusion decreases with the cohesion of the overburden: at same depth, thin saucer-shaped sills emplaced in high-cohesion host rock are broader than massive laccoliths emplaced in low-cohesion host rock.

4. The overburden deformation modes accommodating saucer-shaped sill and laccolith emplacement are dominated by elastic bending and shear failure, respectively; the emplacement of cone sheets may be accommodated by mixed elastic bending and shear failure of the overburden.

Our laboratory experiments show that cohesion, as a measurement for the host rock strength and associated deformation mode, is an important factor controlling magma emplacement. In addition, our results show the necessity to account for both elastic deformation and shear failure of the host to realistically address the emplacement of magma in the brittle upper crust.

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