Use of vegetable oil and silica powder for scale modelling of magmatic intrusion in a deforming brittle crust

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Abstract

In the upper crust, intrusive bodies adopt different configurations, depending on the tectonic setting. This paper describes a new technique for analogue modelling of such intrusions, specifically of low-viscosity magma into a deforming brittle crust. Proper dynamic scaling is an important consideration. If the crust has a Coulomb failure envelope, the model material should have a cohesion $C$ of 40 Pa to 7500 Pa. If the intruding magma is highly mobile, the model fluid should have a viscosity of $4 \times 10^{-9}$ Pa s to 75 Pa s.

For the model crust, we have used crystalline silica powder (SI-CRYSTAL), siliceous microspheres (SI-SPHERE) of grain size $<30 \mu m$, and a mixture (SI-MIX) of both materials. The mechanical properties of these powders have been obtained by shear and tension–shear tests. SI-CRYSTAL powder is cohesive ($C \approx 300$ Pa, angle of internal friction $\phi \approx 45^\circ$) and represents competent rock. SI-SPHERE powder is much less cohesive ($C \approx 10$ Pa, $\phi \approx 25^\circ$) and represents incompetent rock. SI-MIX has intermediate properties. The model magma is a vegetable oil, which solidifies at room temperature. Its viscosity when molten is $\eta = 2 \times 10^{-2}$ Pa s at 50 °C.

Using these materials, we have done some preliminary experiments, to investigate the intrusion of low-viscosity magma into sedimentary basins. In non-deformed settings, intrusions were saucer-shaped cone sheets. In horizontal extension, they were steep dykes. Finally, in horizontal shortening, intrusions were basal sills that branched into thrust faults.

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1. Introduction

On planet Earth, magmatic activity tends to concentrate near tectonic plate boundaries, where most volcanoes lie along active fault zones. At divergent boundaries or in areas of extensional tectonics, magmatism is often associated with normal faults [1–3].

Volcanoes also appear close to active faults at convergent margins. In the Central Andes, transverse volcanic alignments are associated with strike-slip faults [4,5]. In some areas, such as the Sevier fold-and-thrust belt of Western Montana, magmatic activity is associated with thrusting [6,7].

Because magmatic activity is common to a wide range of tectonic settings, intrusive bodies in the brittle crust have a wide range of shapes. In non-deformed sedimentary basins, they tend to be axisymmetric cone-
sheets (Fig. 1A; [8]) or laccoliths [9,10]. In extensional settings, vertical dykes form as hydraulic fractures, perpendicular to the least principal stress \( \sigma_3 \) (Fig. 1B; [11]). Both dykes and deformation concentrate in narrow rift zones. In compressional settings, intrusions also appear to be associated with deformation. A notable example is the syn-tectonic Boulder Batholith in Montana, which was emplaced during the formation of the Sevier fold-and-thrust belt [6,7]. The batholith is tabular, 5–10 km thick, gently dipping, and lies next to a major thrust (Fig. 1C; [6,7]).

Even if they have widely variable shapes, most intrusive bodies in the brittle crust involve flow of magma and faulting of host rocks. However, the geometrical
relationships between magmatic intrusions and tectonic structures, such as faults, are not always clear [12]. Moreover, although the mechanics of magmatic intrusion have long been studied, the interactions between magmatic and tectonic processes remain poorly understood, especially when the magma is of low viscosity. In particular, the following questions remain unanswered. In active tectonic settings, what are the general processes that govern magmatic emplacement in the brittle crust? How do low-viscosity magmas interact with tectonic deformation? Does magma propagate along fault zones? Could the presence of magma control the development of fault zones?

One way of answering such questions is through analogue modelling. M. King Hubbert was the first to model hydraulic fracturing, by injection of a liquid into a stressed elastic material (gelatine), which failed in axial tension [11]. This technique has been widely used to model dyke emplacement by hydraulic fracturing of homogeneous media [13–20]. Experiments on stratified gelatine have also addressed laccolith emplacement [10,21]. Hubbert and Willis found that hydraulic fractures always form perpendicularly to the greatest principal stress [11]. Since then, gelatine experiments have been used to model the load induced by a volcanic edifice on dyke distribution [21–23]. Unfortunately, gelatine fails easily in tension, but not in shear, so that it is inappropriate for studying faults at the scale of the crust.

In contrast, sand is a Coulomb material, which fails in shear and is therefore suitable for modelling brittle deformation of the upper crust [24–26]. For modelling viscous rock, such as the ductile lower crust, evaporites or even magma, silicone putty has proved to be suitable. Sand and silicone have been used to study interactions between magmatic intrusion and associated deformation of host rocks [27,28], emplacement of laccoliths [29], caldera formation and associated domes [30,31]. The effect of tectonic deformation on magmatic emplacement has been modelled in extension [32] and in strike-slip [33,34]. Unfortunately, silicone represents magma of high viscosity (more than 10^{10} Pa s) and it therefore emplaces by ballooning (pushing aside the host material). Thus models made of sand and silicone are not suitable for studying the emplacement of low-viscosity magma into a deforming brittle crust.

In order to model mechanical interactions between low-viscosity magma and tectonic deformation, it is convenient to use a Coulomb material, representing a brittle crust, and a low-viscosity liquid, representing low-viscosity magma. To our knowledge, nobody else has done experiments of this kind. The challenge is to find new analogue materials, which have suitable physical properties.

In the first section of this paper, we discuss scaling and practical constraints, which limit the range of suitable model materials. The scaling is based on mechanical principles, not thermal ones. This approach is valid, if the magma is of low viscosity, so that heat advects readily. In the second section, we describe the mechanical properties of fine-grained silica powders, representing the brittle upper crust, and a vegetable oil, representing low-viscosity magma. Finally, in the third section, we describe some preliminary experiments, which make use of these materials and which result in geologically realistic configurations.

2. Required properties of model materials

In terms of physical properties, model materials should satisfy two sets of conditions. The first set comes from correct scaling of physical parameters. The second set comes from experimental and practical constraints.

2.1. Scaling

A scale model should be geometrically, kinematically and dynamically similar to its natural prototype [24,25]. For geometrical similarity, there is a fixed length ratio between model and nature. For kinematical similarity, a time ratio is necessary. Finally, dynamic similarity requires a set of ratios for forces, stresses, and so on. Correct scaling allows the experimenter to reproduce geological processes at a convenient size and rate.

2.1.1. Setting a stress scale for the brittle crust

In most tectonic processes, motions are very slow and inertial forces are negligible. The main forces to consider are body forces (due to gravity) and surface forces (stresses). The following ratio between body forces and surface forces is dimensionless:

\[ \Pi_1 = \frac{\rho \times g \times l}{\sigma_1}. \]  

(1)

Here \( \rho \) is the rock density, \( g \) is the acceleration due to gravity, \( l \) is a length and \( \sigma_1 \) is a measure of tectonic stress. Being dimensionless, the ratio \( \Pi_1 \) should be identical in model and in nature. In other words, the ratio \( \Pi_1(\text{model})/\Pi_1(\text{nature}) \) should be unity, so that:

\[ \sigma_1^* \approx \rho^* g^* l^*. \]  

(2)

Here \( \sigma_1^* \) is the stress ratio, \( \rho^* \) is the density ratio, \( g^* \) is the gravity ratio and \( l^* \) is the length ratio between model and nature.
Values in nature are after [43, 45, 48]. Experimental values come from scaling (Table 1).

**Table 1**

<table>
<thead>
<tr>
<th>Chosen model ratios</th>
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<tbody>
<tr>
<td>$\rho^*$</td>
</tr>
<tr>
<td>Lower bound</td>
</tr>
<tr>
<td>Upper bound</td>
</tr>
</tbody>
</table>

Values are lower bounds (top row) and upper bounds (bottom row) on model ratios of density, length, stress, magma velocity, and magma viscosity.

For experiments in the Earth’s field of gravity, the gravity ratio is unity. The density of sedimentary rocks ranges between 2.0 and 2.7 g cm$^{-3}$ and the density of common granular model materials ranges between 1 and 1.5 g cm$^{-3}$. Thus the density ratio is between 0.4 and 0.75 (Table 1). In the laboratory, we would like to reproduce kilometre-scale structures (such as dykes, laccoliths or faults) in a model a few centimetres thick, so that the length ratio $l^*$ is between $10^{-5}$ (1 cm represents 1000 m) and $10^{-4}$ (1 cm represents 100 m). This implies that the stress ratio $\sigma_t^*$ ranges between $4 \times 10^{-6}$ and $7.5 \times 10^{-5}$ (Table 1), so that the model needs to be about 13,000 to 250,000 times weaker than its geological counterpart.

We assume that brittle sedimentary rocks fail according to a linear Mohr–Coulomb criterion [35], where the parameters are the cohesion $C$ and the angle of internal friction, $\phi$. As the cohesion has the dimension of stress, the ratio for cohesion $C^*$ is in the same range as $\sigma_t^*$. For competent rocks, the cohesion ranges from about 10$^7$ Pa for a limestone, to about 10$^8$ Pa for a marble ([26, 35] and references therein; Table 2). Thus the cohesion of the model material should be between 40 and 7500 Pa (Table 2). However, for incompetent rocks, the cohesion can be two orders of magnitude smaller [36]. If so, it should be between 0 and about 10$^2$ Pa.

For a sedimentary rock, the angle of internal friction, which is dimensionless, ranges between 26$^\circ$ and 45$^\circ$ ([26] and references therein). Thus the angle of internal friction of model materials should also range between 26$^\circ$ and 45$^\circ$.

Granular materials, such as sand, have been widely used in experiments, to model the brittle upper crust (e.g. [24, 37, 38]). Such materials are easy to handle, and are suitable for modelling Coulomb behaviour of the brittle crust.

### 2.1.2. Setting a viscosity scale for magma

Viscous stresses within magma tend to be much smaller than tectonic stresses [39]. Let us define the dimensionless ratio $\Pi_2$ between the viscous and tectonic stresses. The viscous stresses will be properly scaled if the ratio $\Pi_2$(model)/$\Pi_2$(nature) is unity, so that:

$$\frac{\Pi_2(\text{model})}{\Pi_2(\text{nature})} = \frac{\sigma_t^*}{\sigma_t^*} \approx 1.$$  

(3)

Here, $\sigma_t^*$ is the viscous stress ratio between model and nature. If the magma is a Newtonian fluid, $\sigma_t^*=2\varepsilon \eta$, where $\sigma_t^*$ is the deviatoric stress, $\eta$ is the viscosity of the fluid and $\varepsilon$ is the rate of shear. Thus (3) becomes:

$$\dot{\varepsilon} \eta^* = \sigma_t^*$$  

(4)

where $\dot{\varepsilon}^*$ is the strain rate ratio, $\eta^*$ is the viscosity ratio and $\sigma_t^*$ is the stress ratio. The strain rate ratio can be expressed in terms of the velocity ratio, $V^*$, and the length ratio, $l^*$. Hence (4) becomes:

$$\eta^* = \frac{\sigma_t^* l^*}{V^*}.$$  

(5)

We have already chosen a length ratio $l^*$ between $10^{-5}$ (1 cm represents 1000 m) and $10^{-4}$ (1 cm represents 100 m; Table 1). The stress ratio $\sigma_t^*$ is in the range between $4 \times 10^{-6}$ and $7.5 \times 10^{-5}$. This leaves the viscosity and velocity ratios to be defined.

For basaltic dykes, geophysical observations and theoretical studies have shown that the propagation velocity is about 0.1 to 1 m s$^{-1}$ [40–42]. For magma of rhyolitic or granitic composition, which has a higher viscosity, smaller velocities of about $10^{-2}$ m s$^{-1}$ have been estimated [43, 44]. Thus in nature, magma velocity ranges from $10^{-2}$ to 1 m s$^{-1}$ (Table 2). In the laboratory, a practical rate of injection is a few tens of ml per min, or about $10^{-7}$ m$^3$ s$^{-1}$ [39]. If the injection pipes are about 1 cm in diameter, the average liquid velocity is about $10^{-3}$ m s$^{-1}$, and the model ratio of velocity $V^*$ ranges between $10^{-3}$ and $10^{-1}$. Under these conditions,
from Eq. (5), the viscosity ratio ranges between $4 \times 10^{-10}$ and $7.5 \times 10^{-6}$ (Table 1).

In nature, magma viscosity depends on chemical composition, water content, temperature, and volume fraction of crystals. It can therefore vary widely, from 10 Pa s for basaltic melts, to $10^{18}$ Pa s for partially crystallised granitic magma [28,43,45]. In this paper, we consider low-viscosity magma of basaltic to rhyolitic composition. The viscosity can range, from 10 Pa s for a basaltic melt with a high water content, near its liquidus temperature, to about $10^7$ Pa s for a more siliceous and volatile-poor magma, containing crystals [43,46–48] (Table 2). Thus, from the values of $\eta^*$, we calculate a range of viscosity for the model magma, between $4 \times 10^{-9}$ Pa s for low-viscosity magma and 75 Pa s for high-viscosity magma (Table 2).

2.2. Experimental constraints

2.2.1. Finding a fluid that solidifies

Models that are made of sand and silicone are not transparent. One way of observing the structures that result from deformation is to section the models afterwards. This is feasible because the silicone has a relatively high viscosity ($\approx 10^4$ Pa s) and does not flow appreciably during the interval of observation.

In contrast, for our experiments, the viscosity of the model magma should be small ($<75$ Pa s, see Section 2.1). This means that the fluid flows very quickly in a normal field of gravity—too quickly for the models to be sectioned. An alternative is to use a fluid that solidifies at room temperature, but is molten at a somewhat higher temperature (less than 50 °C). Injected as a low-viscosity fluid, it freezes later in the experiment, conserving the structures. Thus, the model can be cut into observable cross sections.

2.2.2. Preventing the percolation of model magma

Magma in nature intrudes its host rock by moving along fractures, rather than through pore space. We wish to ensure the same mechanism of fluid transport in our models. However, many of the granular materials that have been used in analogue models are susceptible to percolation by fluids of low viscosity.

One way of preventing percolation in granular materials is to reduce the pore size. This is because the permeability is proportional to the square of the grain size. Another way of preventing percolation is to use materials that are not very wettable by the fluid in question. Thus, the fluid should be chemically and physically incompatible with the granular material.

2.3. Summary: required properties of model materials

Experiments on magmatic injection in a brittle crust will be properly scaled if the model materials have the following physical properties.

(1) The model crust should have a Coulomb failure envelope. Granular materials have the requisite properties.

(2) The cohesion of the model crust should be between 40 Pa and 7500 Pa, and its angle of internal friction should be between 26° and 45°.

(3) The viscosity of the model magma should be low (between $4 \times 10^{-9}$ Pa s and 75 Pa s).

For practical reasons, we add the following requirements.

(4) The model magma should solidify at room temperature.

(5) The host material should be fine-grained, to prevent percolation.

(6) Magma and host should be chemically and physically incompatible.

3. Suitable model materials

3.1. Fine-grained silica powder for the brittle crust

3.1.1. Basic properties

We have investigated the properties of two classes of silica powder (Table 3), a crystalline silica (SI-CRYSTAL), and empty microspheres (SI-SPHERE), produced by Verre Industrie.

SI-CRYSTAL has a grain size of about 10–20 μm (Fig. 2). The grains are angular (Fig. 3) and have a

<table>
<thead>
<tr>
<th>Material</th>
<th>Roughness</th>
<th>$d$ (μm)</th>
<th>$\rho$ (g cm$^{-3}$)</th>
<th>$\phi$</th>
<th>$C^*$ (Pa)</th>
<th>$T$ (Pa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SI-CRYSTAL</td>
<td>Rough</td>
<td>10–20</td>
<td>1.33±0.2%</td>
<td>40±2</td>
<td>288±26</td>
<td>88±17</td>
</tr>
<tr>
<td>SI-SPHERE</td>
<td>Smooth</td>
<td>30</td>
<td>1.56±0.18%</td>
<td>~24</td>
<td>1.5</td>
<td>Negligible</td>
</tr>
</tbody>
</table>

Parameters are grain size ($d$), density ($\rho$), angle of internal friction ($\phi$), extrapolated cohesion ($C^*$), and tensile strength ($T$).
tendency to interlock. Macroscopically, SI-CRYSTAL powder is cohesive, so that a free vertical face up to 15 cm high does not collapse under its own weight.

SI-SPHERE powder has a grain size of about 30 μm (Fig. 2). The grains are spherical and very smooth (Fig. 3). Macroscopically, SI-SPHERE powder is mechanically similar to sand and collapses under its own weight, forming a cone with a critical surface slope.

For both silica powders, the density and porosity depend strongly on the amount of compaction. The density of compacted and non-compacted SI-CRYSTAL powder and SI-SPHERE powder has been determined to the nearest 0.01 g cm$^{-3}$, by weighing various known volumes of material in a cylinder several centimetres wide. The calculated density is 1.331 ± 0.011 g cm$^{-3}$ for compacted SI-CRYSTAL and 1.557 ± 0.012 g cm$^{-3}$ for SI-SPHERE (Fig. 4). Because the plots of mass versus volume are almost perfectly linear, there is no evidence that the materials compact under their own weights.

For the range of stresses acting in our experiments, we assume that the granular model materials have a linear Coulomb failure envelope [49]:

$$
\tau = C + \mu \times \sigma_n.
$$

Here $\sigma_n$ and $\tau$ are the normal and shear stresses on the fracture surface, $C$ is the cohesion of the material.
and $\mu = \tan \phi$, where $\phi$ is the angle of internal friction.

3.1.2. Apparatus for failure tests

3.1.2.1. Shear box. We tested the powdered materials using a shear box similar to that of [50], as modified by [51]. Such an apparatus has been used by many authors to measure the mechanical properties of sands, glass microspheres or sugar [26,51–55]. The apparatus consists of two cylinders, which contain a sample of material (Fig. 5). The upper cylinder hangs from four strings above a fixed lower cylinder, and it moves laterally under the traction exerted by a mass $M$ hanging over a pulley (Fig. 5). A gap between both cylinders controls the location of the shear fracture and avoids friction between the two cylinders (Fig. 5).

The height $H$ of material above the gap determines the normal load $\sigma_n = \rho \times g \times H$ on the shear plane, where $\rho$ is the density of the material and $g$ is the acceleration of gravity. Before each test, the upper cylinder (which is in two halves) is temporarily opened, to suppress friction between the granular material and the cylinder wall.

Such friction may significantly decrease the vertical load $\sigma_n$ on the fracture plane (the silo effect; [56]). The opening of the cylinder is possible only because SI-CRYSTAL powder is cohesive and does not collapse.

During the test, sand is poured into a container hanging from a pulley until failure occurs (Fig. 5). The known mass $M$, divided by the cross-sectional area of the cylinder, provides a measure of the shear stress that leads to failure. After each test and before any new one, the powder in the cylinders is replaced by a fresh sample.

Frictional resistance to motion in the apparatus has been estimated by measuring the mass that is required to displace the upper empty cylinder a few millimetres. The corresponding shear stress is two to three orders of magnitude smaller than the shear stresses measured on a test sample.

3.1.2.2. Tension–shear box. In cohesive materials, such as compacted SI-CRYSTAL powder, open fractures occur if normal stresses $\sigma_n$ are truly tensile. We have set up a simple apparatus to measure the stresses on surfaces that fail in tension and shear simultaneously (Fig. 6).

One half of a split cylinder is fixed to a rectangular plate (Fig. 6). The other half is on a mobile semicircular plate. The assembled cylinder (of diameter $D$) is filled with compacted model material to a known height $H$. A vertical gap, between the two plates and the two half-cylinders, controls the location of the fracture surface (hachured in Fig. 6). As for the shear box, a container is filled with sand until failure occurs. The container pulls a string at an angle $\alpha$ to the fracture surface (Fig. 6). For $\alpha = 90^\circ$, the fracture is in tension; for $\alpha = 0^\circ$ it is in shear; and for intermediate angles, it is in both tension and shear.

The components of tensile and shear stress acting on the fracture surface are:

$$
\begin{align*}
\sigma_n &= \sigma \sin \alpha = \frac{M \sin \alpha}{H \times D} \\
\tau &= \sigma \cos \alpha = \frac{M \cos \alpha}{H \times D}.
\end{align*}
$$

When $\alpha = 0^\circ$, $\sigma_n = 0$ Pa and the rupture stress equals the true cohesion $C$ of the material. When $\alpha = 90^\circ$, $\tau = 0$ Pa and the rupture stress equals the tensile strength $T$ of the material.
Frictional resistance to motion in the apparatus has been estimated by measuring the mass that is required to displace the empty mobile cylinder by a few millimetres. The corresponding shear stress is one to two orders of magnitude smaller than the shear stresses measured on a sample.

3.1.3. Results of failure tests

3.1.3.1. Compacted SI-CRYSTAL powder. A Mohr failure envelope for compacted SI-CRYSTAL powder has been obtained using both the shear box (full circles, Fig. 7) and the tension–shear box (open circles and triangles, Fig. 7). Shear tests have been done for several values of $\sigma_n$, between 0 and 1200 Pa. The best linear fit to the data for shear fractures (full and open circles in Fig. 7) gives an extrapolated cohesion of $C'=288\pm26$ Pa, somewhat smaller than the measured value of $374\pm40$ Pa (open circles in Fig. 7). The slope of the straight line provides the angle of internal friction of the material, $\phi=40^\circ$, with an error of about $\pm5\%$ (Table 3).

The tension–shear data (white triangles) prolong the shear test results, and fit a quadratic Griffith failure.

Fig. 4. Plots of mass versus volume for three model materials. Slopes of fitted lines provide best estimates of densities. (A) Compacted SI-CRYSTAL, density $\rho=1.33$ g cm$^{-3}$; (B) Compacted SI-SPHERE, $\rho=1.56$ g cm$^{-3}$; (C) Molten vegetable oil, $\rho=0.89$ g cm$^{-3}$.
The tensile strength of the compacted SI-CRYSTAL powder has been measured at $T = 88 \pm 17$ Pa (Table 3).

3.1.3.2. Non-compacted SI-CRYSTAL powder. We measured the failure stresses for nominally non-compacted SI-CRYSTAL powder using the tension–shear box, in which the vertical load is small. This avoided unwanted compaction. The measured cohesion is $C \approx 100$ Pa (compared with $C \approx 300$ Pa for the compacted powder), and the tensile strength is $T \approx 50$ Pa (compared with $T \approx 100$ Pa; Fig. 8). Our results show that non-compacted SI-CRYSTAL powder is weaker than the compacted variety. Compaction is therefore a critical parameter in the mechanical behaviour of SI-CRYSTAL powder.

3.1.3.3. SI-SPHERE powder. The mechanical properties of a fine-grained material, which has the same specifications as our SI-SPHERE powder, have been measured previously [57]. We assume that the cohesion and angle of internal friction of both materials are identical ($C = 1.5$ Pa, $\phi = 24^\circ$, Table 3).

3.1.3.4. Mix of SI-CRYSTAL and SI-SPHERE powders (SI-MIX). The grain sizes of SI-CRYSTAL and SI-SPHERE powders are almost the same (Table 3), so that a mix of these two materials is homogenous. The mechanical properties of the mix depend on the proportions of SI-CRYSTAL and SI-SPHERE powders. Whereas extension in cohesive SI-CRYSTAL and non-cohesive SI-SPHERE powders results in open fractures and shear fractures, respectively, extension in a material made of 20 vol.% of SI-CRYSTAL and 80 vol.% of SI-SPHERE results in open fractures at the surface that continue downward as shear fractures. In cohesive Coulomb materials, the height of an open fracture $h$ is approximately proportional to the tensile strength $T$ of the material, and thus to the cohesion $C$ [58]. In SI-MIX powder, $h \approx 3$ cm, and in SI-CRYSTAL powder, $h \approx 15$ cm, so that $C_{\text{MIX}} \approx C_{\text{CRYST}}/5 \approx 60$ Pa.

3.2. Vegetable oil for low-viscosity magma

The vegetable oil that we have tested is produced by ASTRA under the trademark Vegetaline and is commonly used in cooking. It is solid at room temperature and melts to low-viscosity oil when the temperature exceeds $\approx 31$ °C (Fig. 9). The density of the oil, measured using the same method as for granular materials, was found to be $0.89$ g cm$^{-3}$ at 50 °C, with an error of 0.45% (Fig. 4). As the oil solidifies, its volume decreases by less than 4%, which is negligible for our purposes.

Curves of stress as a function of strain rate have been obtained for various temperatures (Fig. 9), using a commercially available rotary viscometer (Rheo, model RV8). The viscometer measures the stress $\sigma$ for a constant imposed strain rate $\dot{\gamma}$. The resulting curves are almost perfectly linear, showing that the vegetable oil is a Newtonian fluid. The slope gives the viscosity $\eta$ with an error of less than $\pm 0.5\%$ (Fig. 9). We have measured the viscosity for various temperatures in the range from 30 °C to 75 °C, for which the material is totally molten (Fig. 9). As the temperature drops below 31 °C, the material solidifies in part and the viscosity increases strongly. In what follows, we consider data only for temperatures above 31 °C (Fig. 9).

The data fit the relationship of Arrhenius:

$$\eta = A \times \exp(E_a/RT).$$

Here $A$ is a constant, $E_a$ is the activation energy, $R$ is the universal gas constant and $T$ is the temperature.
(in degrees Kelvin). The best-fitting values are $A = (2.3 \pm 0.8) \times 10^{-7}$ and $E_a = 30380 \pm 850$ J (Fig. 9). Between 31 °C and 75 °C, the viscosity decreases from $4 \times 10^{-2}$ Pa s to $10^{-2}$ Pa s. At 50 °C, the viscosity is $\eta \approx 2 \times 10^{-2}$ Pa s.

3.3. Conclusions

Because SI-CRYSTAL and SI-SPHERE have Mohr–Coulomb failure envelopes, they are both suitable for modelling the brittle upper crust [24,25,49].
The cohesion of SI-CRYSTAL powder ($C \approx 300$ Pa) is in the required range for correct scaling and is much higher than that of the granular model materials that have been used in most previous studies \[26,52,53,57\]. In contrast, the cohesion of SI-SPHERE powder is very small. The ratio is similar to that between a competent rock (such as limestone or sandstone) and an incompetent rock (such as shale) \[36\]. The SI-MIX material has intermediate mechanical properties, so that it can model rocks of low cohesion.

The viscosity of the vegetable oil ($\eta \approx 2 \times 10^{-2}$ Pa s at 50 °C) is also in the required range for correct scaling. It is much lower than that of the stiff silicones that have been used in previous studies.

### 4. Preliminary experiments on intrusion and deformation

#### 4.1. Experimental technique

We have done three sets of preliminary experiments, to test the suitability of fine-grained silica powder and low viscosity vegetable oil for experiments on magmatic injection into a brittle crust. The apparatus (Fig. 10) is an improvement on a previous version \[39\]. A pack of silica powder, representing brittle crust, lies in a rectangular box, 60 cm long, 40 cm wide, and 20 cm deep. In all the experiments, the boundary conditions on the model are purely kinematic. Given a correct scaling of material properties, this ensures that forces and stresses are within acceptable ranges at the model boundaries. At one end of the box, a steadily moving piston causes deformation within the silica pack. A basal plate fixed to the piston results in a velocity discontinuity that controls the locations of the first faults. Independently, a pump injects molten vegetable oil at a steady predetermined rate through an orifice at the base of the pack.

We describe three experiments (A, B and C, Fig. 11), representing different tectonic contexts. The models were constructed in successive layers. In Experiments A and C, each layer was built to a pre-selected volume and thickness, by weighing a batch of silica powder of known density, pouring it into a sieve, and shaking it through. Once the layer was in place, we flattened the surface, by scraping it down to a datum level. The lowermost layer was of SI-CRYSTAL, 1.13 cm thick. The next was of SI-SPHERE, 0.33 cm thick. This procedure was repeated 3 times. Finally, we added an uppermost layer of SI-CRYSTAL, 1.13 cm thick, and compacted the multilayer using a compressed-air vibrator. The total thickness of the multilayer was then 5.5 cm (Fig. 11). Experiment B was made of a homogeneous SI-MIX. It was constructed in successive layers, between which thin layers of blue sand were poured, so as to form internal markers. The final pack was compacted, using a compressed-air vibrator. The total thickness of the model was then 9 cm.

In Experiment A, there was injection, but no deformation; in Experiment B, injection was synchronous.
with horizontal extension; and in Experiment C, injection was synchronous with horizontal shortening (Table 4).

At the end of each experiment, the model was sprinkled with water, so as to render the silica powders cohesive, and then it was sliced longitudinally (Fig. 11).

4.2. Experiment A: injection only

In Experiment A, vegetable oil was injected steadily through a central orifice, 5 mm in diameter, at the base of the model. The oil formed an almost axisymmetric intrusive body around the injection point (Fig. 11A). After 410 s of injection, the oil reached the surface (not seen in Fig. 11A). Within the host material, the oil percolated no more than a few millimetres. The intrusive body consisted of a basal sill, which branched upward and outward into a cone sheet (saucer-shaped sill). Above the injection point, the silica powder was jacked up, forming a smooth dome. Differential uplift at the periphery of the dome was by vertical annular shear [39]. Such features are typical of cone sheets, as observed in the field [59,60] or on seismic profiles (Fig. 1; [8]). Theoretical studies also predict that internal pressure in a magma chamber results in an uplifted dome with a cone sheet at its edge [60–62].

In detail, the dip and thickness of the cone sheet changed from one layer to the next (Fig. 11A). Across the SI-CRYSTAL layers, the intrusion was thick and flat-lying, whereas across the SI-SPHERE layers it was thin and steep, sharply offsetting the layering. Because the SI-CRYSTAL layers were cohesive, we infer that the

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Fig. 9. Rheological properties of vegetable oil. Data are from a rotary viscometer. (A) Plot of stress versus strain-rate for three different temperatures. Linear relationship shows that vegetable oil is Newtonian. Slope is measure of viscosity. (B) Inferred viscosity versus temperature. Arrhenius relationship (top right) provides good fit to data points above 31 °C. Inferred activation energy ($E_a$) is about 30 kJ mol$^{-1}$.
intrusion traversed them in a series of open hydraulic fractures [39]. In contrast, across the non-cohesive SI-SPHERE layers, we suspect that the intrusion propagated along vertical shear fractures at the periphery of the uplifting dome. Similarly, in experimental models of extension within a multilayered model of flour and sugar, open fractures were observed to develop in the cohesive flour, and shear faults in the less cohesive sugar [63]. In nature as well, sheeted intrusions have variable dips, probably as a result of mechanical stratification in the host rocks [64,65].

4.3. Experiment B: injection and simultaneous extension

In Experiment B, the model was made of SI-MIX powder. Extension caused a rift to nucleate at the leading edge of the basal plate (Fig. 11B). Open fractures at the surface continued downward as normal faults. First to form were two major rift-bounding faults. Then a series of minor faults formed within the rift. Injection started after 620 s. About 140 s later, some oil reached the surface (not seen in Fig. 11B). In cross section, the intrusive body was a thin steep dyke, lying approximately perpendicular to the extension direction. Such an arrangement is typical of hydraulic fractures [11].

From bottom to top of the model, the dyke varied in dip. For the first two centimetres, it was almost vertical. Above that, it followed one of the normal faults (Fig. 11B). We infer that the lower part emplaced by hydraulic fracturing in non-faulted material, whereas the upper part took advantage of the pre-existing fault. Such an interaction between dykes and normal faults has been observed in rift systems in Iceland [1].

4.4. Experiment C: injection and simultaneous shortening

In experiment C, deformation resulted in a doubly verging thrust system. In the early stages of the experiment, the model shortened by about 9%, without injection. Thrusts nucleated at the leading edge of the basal plate (Fig. 11C). Then, as deformation proceeded, oil was injected rapidly, for a short time interval of 20 s. This resulted in an intrusive body, a few millimetres thick, comprising a basal sill and a splay along one of the thrusts (Fig. 11C).

The structures in Experiment C are similar to those obtained by deformation of quartz sand if there is no injection [37]. According to theoretical studies, a thrust wedge developing in a Coulomb material will have a triangular profile. The apical angle of the wedge depends on the angle of internal friction, the cohesion, and the frictional properties at the base [66]. The internal structures also depend on whether the model is
Fig. 11. Cross sections of three models in different tectonic settings. (A) Experiment A (injection without simultaneous deformation). Model consists of layers of SI-SPHERE and SI-CRYSTAL. Section was cut after wetting. Photograph shows layering. Vertical tension cracks result from wetting. Line drawings show layering and intrusion (dark grey). Injection point (black arrow) was near centre of model. Intrusion consisted of basal sill, branching upward into axisymmetric cone sheet. Numbers (see Table 4) denote volume of injected oil ($v_0$) and horizontal strain ($\epsilon$). (B) Experiment B (injection and simultaneous extension). Model was made of homogeneous SI-MIX (SI-CRYSTAL and SI-SPHERE). Line drawings show interpreted faults (dashed lines). Extension resulted in rift zone. Normal faults nucleated at edge of receding basal plate. Notice large offsets of layers across faults. Injection resulted in steep thin dyke. (C) Experiment C (injection and simultaneous shortening). Model consists of layers of SI-SPHERE and SI-CRYSTAL. Section was cut after wetting. Photograph shows layering. Vertical tension cracks result from wetting. Line drawings show layering and intrusion (dark grey). Injection point (black arrow) was near centre of model. Thrust faults nucleated at edge of advancing basal plate. Injection was for short period only. It resulted in basal sill, branching upward along thrust fault.
In sandbox models, as in our experiments, the deformation results in folds as well as faults [57,67]. In contrast with Experiment B, the intrusion in Experiment C was asymmetrical and seemed to follow the thrust faults. Similar relationships have been observed for the Boulder Batholith of Montana, which intruded a developing fold-and-thrust belt, along a major thrust fault (Fig. 1; [6,7]).

5. Discussion

5.1. Percolation of oil through silica powder

We have seen that proper scaling requires a fluid of low viscosity. However, we wish to avoid percolation through the host material. A granular material, such as sand, has a larger porosity than most rock, and this makes it susceptible to percolation. However, SI-CRYSTAL and SI-SPHERE powders have grain sizes of less than 30 μm. Their intrinsic permeabilities are therefore small. In addition, silica powders are strongly hydrophilic and this makes them resistant to wetting by oils and greases. As a result, percolation is negligible during propagation of a hydraulic fracture. However, it may reach several millimetres into the host material, by the time the intrusion has solidified.

5.2. Modelling tectonic processes

In nature, extensional or compressional tectonics result in rift or thrust systems, respectively. Our models, although simplified, reproduce the main features of such systems. Moreover, the alternating layers of SI-SPHERE and SI-CRYSTAL account for some of the mechanical heterogeneity of a stratified sedimentary sequence. In nature, deformation of a multilayer results in faults and folds. Similar structures can be obtained experimentally, if the mechanical properties vary from one layer to the next, while the cohesion is small [57]. Thus SI-CRYSTAL, SI-SPHERE and SI-MIX powders are suitable for modelling tectonic processes in the brittle upper crust.

5.3. Modelling dyke and sill emplacement

In Experiments A and B, the oil appeared to have intruded by hydraulic fracturing of the silica powder. The strong cohesion of SI-CRYSTAL and SI-MIX powders allows fractures (such as hydraulic fractures) to open, without collapsing. Injection of a low-viscosity fluid therefore results in sheeted intrusions (Fig. 11), including dykes, sills or laccoliths. In nature, the thicknesses of the sheeted intrusions range between 1 and 100 m. Given that the scale ratio \( l^* \) between model and nature ranges from \( 10^{-5} \) to \( 10^{-4} \), the expected thickness of model intrusions should range between \( 10^{-5} \) and \( 10^{-2} \) m. The values observed in our experiments (typically, a few mm) fall into this range. In this sense, models made of SI-CRYSTAL, SI-MIX and vegetable oil appear to be more realistic than those made of stiff silicone and cohesion-less sand, in which the fluid emplaces by inflation (“ballooning”), forming rounded plutons, several centimetres thick.

In contrast, SI-SPHERE powder is not cohesive enough for open fractures to form. This explains why, in Experiment A, fractures within the layers of SI-SPHERE powder were mainly shear fractures, whereas those in the SI-CRYSTAL layers were mainly tensile (Fig. 11).

The intrusive body in Experiment A is similar to the saucer-shaped sills observed on seismic profiles (Fig. 1). Such bodies appear to be common structures in sedimentary basins [8], but their mechanisms of emplacement are not clearly understood. To our knowledge, our experiment was the first to have produced a saucer-shaped intrusion. Thus the silica powder and vegetable oil appear to be suitable for studying the processes of emplacement of saucer-shaped sills within sedimentary basins.

In nature, hot magma intrudes a cooler host rock. Similarly, in our experiments, molten vegetable oil intruded a cooler host material. In both nature and experiment, if the rate of heat loss by conduction is less than the rate of heat gain by advection, the intruding liquid will not freeze. Conversely, if the former is greater than the latter, the liquid will solidify. A measure of the

<table>
<thead>
<tr>
<th>Type</th>
<th>( v_p ) (mm s(^{-1}))</th>
<th>( Q ) (ml s(^{-1}))</th>
<th>( t ) (s)</th>
<th>( e ) (%)</th>
<th>( v_0 ) (ml)</th>
</tr>
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<tr>
<td>Experiment A</td>
<td>Injection only</td>
<td>0</td>
<td>0.17</td>
<td>420</td>
<td>0</td>
</tr>
<tr>
<td>Experiment B</td>
<td>Injection + extension</td>
<td>0.03</td>
<td>0.25</td>
<td>760</td>
<td>5.5</td>
</tr>
<tr>
<td>Experiment C</td>
<td>Injection + shortening</td>
<td>0.04</td>
<td>1.22</td>
<td>1225</td>
<td>−9</td>
</tr>
</tbody>
</table>

Values are for piston speed \( (v_p) \), rate of injection \( (Q) \), duration \( (t) \), extension \( (e) \) and volume injected \( (v_0) \).
ratio between gain and loss of heat is the Péclet number, which is defined as:

\[ Pe = \frac{w \times V \times C_p \times \rho}{k} \]  

(9)

Here \( C_p \) is the heat capacity and \( k \) is the thermal conductivity of the magma. If \( Pe > 1 \), the rate of heat gain by advection is higher than the rate of heat loss by conduction.

We have estimated the Péclet number for a single intrusion in both nature and experiment (Table 5). In nature, the number ranges from \( 4 \times 10^4 \) to \( 4 \times 10^9 \), and in our experiments, from 33.8 to 338. All these values are much greater than unity. Thus in nature and experiment, under the above conditions of steady flow, the liquid cannot solidify.

5.4. Modelling intrusion and simultaneous deformation

Silica powders and vegetable oil are suitable materials for modelling, on the one hand tectonic processes (resulting in normal and reverse faults), and on the other hand the emplacement of low viscosity magma in a brittle upper crust (resulting in dykes, sills and laccoliths). In Experiments B and C, the oil was injected during extension and compression, respectively. In extension, oil emplaced perpendicularly to the least principal stress along sub-vertical dykes, and interacted with normal faults. Such features are similar to those observed in nature, for example in Icelandic rift systems (Fig. 1B; [1]). In compression, oil propagated along horizontal sills and along thrust faults. Such relations between intrusive bodies and thrusts appear to be geologically realistic. In the Boulder Batholith of Montana, the main intrusive body seems to be associated with a major thrust (Fig. 1C; [6,7]).

In nature, the time scales for tectonic processes are typically much longer than those of intrusion. The characteristic time interval for the emplacement of single intrusive bodies, such as dykes, sills, laccoliths or plutons, is of the order of months [68] to tens of thousand of years [69], whereas tectonic processes typically last for millions of years. In Experiments B and C, the period of injection was also much shorter than that of deformation. Thus, these experiments were relevant to the emplacement of a single intrusive body.

Because the flow rate can be set independently of the piston velocity, our apparatus can also be used to explore the cumulative effects of multiple or long-term injections during tectonic deformation. This work is in progress [39,70].

6. Conclusions

We have developed an experimental method for modelling the intrusion of low-viscosity magma into a deforming brittle crust. The method relies on new model materials, rather than the conventional sand, stiff silicone, or gelatine.

The first part of this paper dealt with governing equations and scaling. It resulted in a list of required physical properties for the model materials.

(1) The model crust should have a Coulomb failure envelope.
(2) The cohesion of the model crust should be between 40 Pa and 7500 Pa, and its angle of internal friction should be between 26° and 45°.
(3) The viscosity of the model magma should be between \( 4 \times 10^{-9} \) Pa s and 75 Pa s.

For practical reasons, we add the following requirements.

(1) The model magma should solidify at room temperature.
(2) The host material should be fine-grained, to prevent percolation.
(3) The model magma and host should be chemically and physically incompatible.

On this basis, we have chosen fine-grained silica powders (SI-CRYSTAL and SI-SPHERE), to model the brittle crust, and a vegetable oil, to represent low-viscosity magma. The cohesion of SI-CRYSTAL powder is about 300 Pa and that of SI-SPHERE powder

<table>
<thead>
<tr>
<th>Table 5</th>
<th>Thermal parameters in nature and experiment</th>
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<tbody>
<tr>
<td>( w ) (m)</td>
<td>( V ) (m s(^{-1}))</td>
</tr>
<tr>
<td>Nature</td>
<td>( 1 \times 10^3 )</td>
</tr>
<tr>
<td>Experiment</td>
<td>( 10^{-1} )–( 10^{-2} )</td>
</tr>
</tbody>
</table>

Values are for intrusion thickness (\( w \)), magma velocity (\( V \)), magma density (\( \rho \)), heat capacity (\( C_p \)), thermal conductivity (\( k \)), and Peclet number (\( Pe \)). Values in nature are after [45,72,73]. Value of \( C_p \) for vegetable oil was determined with calorimeter.
is negligibly small. The vegetable oil has a viscosity of about $2 \times 10^{-2}$ Pa s at 50 °C. In addition, it solidifies at room temperature, preserving the shapes of intrusive bodies. Finally, the vegetable oil percolates no more than a few mm into the fine-grained silica powder.

In the new experimental apparatus, a mobile piston deforms the model and a volumetric pump injects oil through a basal orifice. The rates of deformation and injection can therefore be set independently.

In preliminary experiments, we have investigated the effects of tectonic setting on the intrusion of low-viscosity magma into sedimentary basins. In a non-deformed setting, the intrusive bodies were saucer-shaped cone sheets. In horizontal extension, they were steep dykes. Finally, in horizontal shortening, intrusive bodies were basal sills that branched into thrust faults. In all settings, the main mechanism of intrusion was hydraulic fracturing.

In these experiments, the time scales for injection were much smaller than those of tectonic processes. Each experiment produced a single intrusive body, such as a dyke, sill or laccolith. As injection and deformation are independent in our experimental apparatus, the time-scale for oil injection could also be set in the same range as that of tectonic processes, so as to model the emplacement of a composite magmatic complex, such as a batholith or a volcano [39,70].

In general, fine-grained silica powder and vegetable oil are suitable materials for modelling the injection of low-viscosity magma into a brittle crust.

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