Permeability measurements of Campi Flegrei pyroclastic products: An example from the Campanian Ignimbrite and Monte Nuovo eruptions

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1. Introduction

Knowledge of permeability is of paramount importance for understanding the evolution of magma degassing during pre-, syn- and post-eruptive volcanic processes (Klug and Cashman, 1996; Rust and Cashman, 2004). Permeability results from the combination of various conduit processes related to magma vesiculation and crystallization that accompany magma rise from chamber/dykes along the conduit to the surface and/or to the formation of fractures due to shear fragmentation (Gonnermann and Manga, 2003; Tuffen and Dingwell, 2005). It is a key parameter in the transition from effusive to explosive volcanism (Eichelberger et al., 1986; Jaupart and Allegre, 1991; Woods and Koyaguchi, 1994; Kozono and Koyaguchi, 2009a,b; Degruyter et al., 2012), for example in the catastrophic failure from dome-forming eruptions to Vulcanian/Plinian behaviour (Lipman and Mullineaux, 1981; Herd et al., 2005). Permeability is also responsible for the quiescent gas loss from persistently degassing volcanoes (Oppenheimer et al., 2003), through formation of networks of continuously connected vesicles (Polacci et al., 2008), where gas percolates and can exit the volcanic system non-explosively (Burton et al., 2007). Quantifying permeability is therefore important for assessing the contribution of magmatic gases to the atmospheric global volatile budget (Gerlach, 2011).

Most permeability estimates existing to date refer to magmas of calc-alkaline compositions. Permeability measurements have been performed on pyroclastic products from andesitic to rhyolitic explosive eruptions (Klug and Cashman, 1996; Klug et al., 2002; Bernard et al., 2007; Bouvet de Maisonneuve et al., 2009; Wright et al., 2009; Degruyter et al., 2010a,b), on dome samples (Melnik and Sparks, 2002; Rust and Cashman, 2004; Mueller et al., 2005), and also on basaltic lavas and scoriae from effusive and mildly explosive volcanic activity (Saar and Manga, 1999; Mueller et al., 2005; Polacci et al., 2012). Permeability has also been measured in synthetic products from degassing experiments of volcanic material (Takeuchi et al., 2008; Bai et al., 2010, 2011). Mueller et al. (2005, 2008) compiled a thorough dataset that includes the only permeability values measured on trachytic volcanic rocks from a past sub-Plinian event: the Agnano Monte Spina (AMS) eruption (4.1 ka, de Vita et al., 1999) from Campi Flegrei (CF), an active, hazardous caldera west of the city of Naples, Southern Italy (Fig. 1). In summary, these studies have illustrated that permeability is overall a function of vesicularity; yet abundant scattering in the permeability-vs.-vesicularity data from the previous works implies that permeability

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depends also on other parameters of which the most important are vesicle size and throat size, vesicle shape and tortuosity (Wright et al., 2009; Degruyter et al., 2010a; Polacci et al., 2012).

Here we report the first permeability measurements performed on trachy-phonolitic pyroclastic products from the Campanian Ignimbrite (CI) and Monte Nuovo (MN) eruptions, two explosive eruptions from CF that have very similar compositions but strongly differ in intensity, magnitude and eruption dynamics (Piochi et al., 2008 and references therein). The obtained results significantly enlarge the existing limited database on the permeability of CF magmas providing an important contribution to understanding and modelling degassing and eruptive processes at this highly risky volcanic caldera.

2. Samples and summary of main textural features

We have chosen to study the permeability of CI and MN products for many reasons. First, the two eruptions represent the two end-members (Plinian and Vulcanian, respectively) in the eruption intensity/magnitude spectrum typical of CF eruptions. Second, CI is the largest magnitude explosive event of the Mediterranean region in late Quaternary (dating 39 ka, De Vivo et al., 2001), while MTN is the last eruption that occurred in the caldera in A.D. 1538 (Di Vito et al., 1987). Finally, there exists in the literature a lot of information on these eruptions about their stratigraphy, deposit characteristics, eruption dynamics, compositional, geochemical and textural data that allows us to interpret the results of this study in a robust scientific context (Rosi et al., 1996; Civetta et al., 1997; Rosi et al., 1999; Polacci et al., 2003; D’Oriano et al., 2005; Piochi et al., 2005; Pappalardo et al., 2007; Piochi et al., 2008, and references therein).

The samples used in this study and the outcrop locations where they were collected are listed in Table 1 and Figs. 1 and 2. The CI sample suite includes pumice products from both distal pyroclastic flow (CI pf Sn and CI pf Mo, Table 1) and proximal breccia (CI BM MP, Table 1) facies, the former collected at San Nicola La Strada and Mondragone, the latter at Monte di Procida (Fig. 1). The MTN samples are pumice and scoria products belonging, respectively, to the lower (LNM) and upper (UM1 and UM2) stratigraphic sequence of the MTN cone (Table 1) and were collected at Monte Nuovo Oasi Park (Fig. 1). All samples from the same location/outcrop were collected approximately at the same stratigraphic height.

The CI distal pyroclastic flow pumice clasts are poorly phorphyritic to aphyric (phenocryst content between < 10 vol% and < 3 vol%, Civetta et al., 1997; Pappalardo et al., 2007), highly vesicular (connected vesicularity ~71–89 vol%, Table 1) volcanic rocks with a mostly microlite-free (microlite content <2 vol%, Civetta et al., 1997; Pappalardo et al., 2007) glassy groundmass. They have foamy vesicle textures consisting of mostly spherical to sub-spherical to slightly deformed vesicles with micron-size walls around larger, interconnected vesicles (Fig. 2a). The CI pumice samples from the Breccia Museo proximal facies are phorphyritic (phenocryst content up to 15 vol%, Melluso et al., 1995), vesicular to highly vesicular (connected vesicularity ~52–73 vol%, Table 1) rocks consisting of a fine vesicle structure in a mostly microlite-free (microlite content <2 vol%, Melluso et al., 1995) glassy groundmass, and of large, interconnected sub-spherical vesicles around single crystals or crystal aggregates (Fig. 2b). The MTN scoria and pumice clasts are characterized by a microlite-bearing (crystal content between 25–32 vol%, D’Oriano et al., 2005; Piochi et al., 2008) groundmass with tabular, acicular, and dendritic alkali-feldspar microlites, as well as widespread magnetite microlites in the microlite-richer scoria (Fig. 2c). Connected vesicularity ranges between 48 and 62 vol% (Table 1), and vesicles have very irregular shapes being deformed by the ubiquitous microlites. Large, irregularly-shaped, interconnected microcrack-like vesicles are very common in the scoria samples (Fig. 2c) (Piochi et al., 2008).

3. Permeability measurements

Connected vesicularity (the ratio between the volume of connected vesicles and the core bulk volume) and permeability measurements were performed on nineteen samples, 11 from the CI pyroclastic flow and breccia facies and 8 from the MTN cone deposits (Table 1). All samples were drilled into cores of about 2.5 cm in diameter and between 1.6 and 3.6 cm in height; their external volume was measured via water displacement for more precision (for a description of the procedure refer to Bouvet de Maisonneuve et al., 2009). Samples where vesicle elongation was clearly recognizable were drilled both parallel and perpendicular to that direction. In Table 1 they are reported as Para and Perp, respectively. Connected vesicularity was obtained by quantifying the volume of the core without connected vesicles (volume of glass + crystals + isolated vesicles) with a Quantachrome He-stereopycnometer, then subtracting it from, and dividing it by, the external volume of the core.

Permeability was measured on the same samples used for connected vesicularity measurements with a Porous Material Inc gas permeameter at the University of Geneva (Switzerland). The samples were inserted in a cylindrical chamber lined with rubber; such rubber was subsequently pushed towards the sample by air coming from the chamber sides to make the sample air-tight. Gas (air) was injected into the permeameter from the base of the chamber at increasing pressure and flow rate. The relative error of log(k1) and log(k2) in Figs. 3 and 4 was estimated to ~2% from multiple measurements of the same sample. To take into account energy loss due to both viscous and inertial effects, the one-dimensional form of the Forchheimer equation (Ruth and Ma, 1992; Rust and Cashman, 2004) was fit to our experimental data to obtain the viscous (Darcian) k1 and inertial (non-Darcian) k2 permeabilities:

$$\frac{P_1^2 - P_0^2}{2PL} = \frac{\mu}{k_1} v + \frac{\rho}{k_2} v^2$$

(1)

where $P_1$ (Pa) and $P_0$ (Pa) are the high and low pressure on each side of a sample of length L, $P$ (Pa) is the gas pressure at which the gas flow is measured, $\mu$ (Pa s) is the gas viscosity, $\rho$ (kg/m³) is the gas density.
and \( \nu \) (m/s) is the gas filter velocity (which is the volumetric flow rate per total cross-sectional area of the sample orthogonal to fluid flow), taken at atmospheric conditions. The first and second terms on the right of the equation represent the contribution of the viscous friction and that of inertia and turbulence, respectively.

4. Results

Viscous permeability \( (k_v) \) of the investigated samples spans a wide range of values between \( 1.22 \times 10^{-14} \) to \( 9.31 \times 10^{-11} \) \( \text{m}^2 \) (Table 1, Fig. 3a). We observe that the most permeable samples \( (k_v = 1.57 \times 10^{-11} – 9.31 \times 10^{-11} \text{m}^2) \) are the scoria clasts from the upper units of MTN (Table 1); pumice samples from the proximal breccia facies of CI are instead the least permeable \( (k_v = 4.85 \times 10^{-14} – 5.05 \times 10^{-12} \text{m}^2) \), Table 1). Pumice clasts from CI distal pyroclastic flows have intermediate permeabilities overlapping with both those of the former and the latter. Inertial permeability \( (k_i) \) follows the same trend as viscous permeability: it increases as viscous permeability increases (Table 1, Fig. 3b), highlighting the direct correlation between these two parameters that can be nicely fitted by a power-law relationship, \( k_i = 8.6 \times 10^{-7} k_v^{0.75} \) (Fig. 3c). We note that, in oriented samples, \( k_v \) and \( k_i \) are, respectively, at least 1 order and 2 orders of magnitude higher in samples oriented parallel to vesicle elongation (Table 1, Fig. 3a, b).

5. Discussion and conclusions

5.1. Comparison with permeability data of other silicic eruptions

CI and MTN permeability data compare well with data obtained from andesitic to the most evolved calc-alkaline volcanic rocks \( (10^{-15} – 10^{-14} \text{m}^2), \) for references see Introduction). Comparisons between our measured CF permeabilities with permeability data measured in rhyolitic (frothy and tube parallel to vesicle elongation) pumices from the Kos Plateau Tuff (KPT) (Bouvet de Maisonneuve et al., 2009) and basaltic scoria from Ambrym volcano, Vanuatu Arc

![Fig. 2. Microtomographic axial slices of the Campi Flegrei samples used in this study: a) CI pyroclastic flow (CI pf, Sn + Mo) pumice, b) CI proximal (CI BM) breccia pumice, C) Monte Nuovo (MTN UM) scoria. Dark voids are vesicles, feldspar microphenocrysts are dark grey objects and oxides are bright spots; background groundmass is light grey. In (c) arrows indicate vesicle channels.](image-url)
Polacci et al., (2012), effusive and explosive volcanic products from Mueller et al. (2005), fallout pumice clasts from Little Glass Mountain (LGM), and obsidian flow samples from Little Glass Mountain (LGM) and Big Glass Mountain (BGM) (Rust and Cashman, 2004), are used to frame our results in a more global perspective (Fig. 4a, b). KPT frothy and tube rhyolitic pumices (crosses), Ambrym basaltic scoria clasts (x), effusive (open diamonds) and explosive (full diamonds) products from Mueller et al. (2005), and obsidian flow samples from Rust and Cashman (2004). Dashed line is parameterization from Degruyter et al. (2012). See text for further explanation.

5.2. Insights into conduit degassing

Similarly to mafic (Saar and Manga, 1999; Mueller et al., 2005, 2008; Polacci et al., 2012) and felsic (Klug and Cashman, 1996; Klug et al., 2002; Mueller et al., 2005, 2008; Bouvet de Maisonneuve et al., 2009; Wright et al., 2009) calc-alkaline natural magmas, we note that in evolved alkaline magmas i) vesicularity does not exert a first order control on permeability (e.g., the MTN samples are the most permeable but not the most porous), and that ii) sample geometry exhibits permeability anisotropy (CI BM sample cores parallel to vesicle elongation display a higher permeability than perpendicular cores) (Wright et al., 2006). Assuming that the direction of elongated vesicles in the conduit works as a proxy for that of magma rising along the conduit to the surface, we suggest that magma degassing is stronger vertically than laterally in the conduit. Previous studies have indicated that the region where degassing vesicle pathways are most likely to occur is near and/or at the conduit walls (Cashman et al., 2000; Llewellin et al., 2002; Polacci et al., 2003; Bouvet de Maisonneuve et al., 2009; Caricchi et al., 2002; Rust and Cashman, 2004).
where the shear strain rate is maximum and magma viscosity becomes lower during magma rise due to viscous dissipation effects (Polacci et al., 2001; Rosi et al., 2004). These two conditions promote efficient vesicle coalescence and vesicle deformation to form a continuous pathway for gas to flow out of the system quiescently in both silicic (Okumura et al., 2009) and mafic (Burton et al., 2007) magmas. The centre and the region between the centre and conduit walls are also affected by magma degassing, although to a potentially minor extent, which recent experimental studies on rhyolitic obsidian have indicated as the mandatory requirement to trigger explosivity during a volcanic eruption (Okumura et al., 2013).

5.3. Insights into inertial effects in vesicular magmas

Quantifying inertial effects inside magmatic foams during outgassing is important for the transition from effusive to explosive eruptions. In the case inertial effects become dominant, explosive eruptions become more likely (Degruyter et al., 2012). From the permeability measurements we can obtain some insight into the onset of inertial effects inside magmatic foams during outgassing. The Forchheimer number \( F_{oc} \), which is essentially a Reynolds number at the scale of the porous medium, provides an easy way to quantify this (Ruth and Ma, 1992). It is obtained from the quantities defined in Eq. (1),

\[
F_{oc} = \frac{\rho v_0 k_1}{\mu k_2}
\]

During volcanic eruptions the gas and magma phase are driven by the local pressure gradient across the conduit, which creates a difference in their velocity (Kozono and Koyaguchi, 2009a,b; Degruyter et al., 2012). This velocity difference creates an internal friction \( F_{mg} \) \((\text{Pa} \cdot \text{m})\) between the two phases, which transfers momentum from the gas to the magma phase. The magnitude of this friction is quantified by the right hand side of the Forchheimer Eq. (1) where

\[
v = v_g - v_m
\]

and \( v_g \) (m/s) is the gas velocity and \( v_m \) (m/s) the magma velocity. If \( k_1 \) and \( k_2 \) are large enough, i.e. the permeability is high and the inertial effects are low, this internal friction will be small and the gas and magma will be able to flow at two separate velocities resulting into an effusive eruption. In the case where \( k_1 \) and \( k_2 \) are small the gas will transfer its momentum to the magma phase such that the two phases remain coupled and an explosive eruption occurs. When Fo becomes larger than 1, gas transport will be governed by the inertial term in the Forchheimer equation (second term in Eq. (1)). Inertia thus becomes dominant when the difference in gas and magma velocity exceeds a critical value,

\[
v_g - v_m > \frac{\mu k_2}{\rho k_1}
\]

As the gas density and viscosity will be mostly similar during outgassing in volcanic eruptions, the ratio \( k_1/k_2 \) will govern the onset of inertia for a given gas volume flux (Rust and Cashman, 2004). The results of this study fall in line with previous measurements of \( k_1 \) and \( k_2 \) (Fig. 4b; Rust and Cashman, 2004; Mueller et al., 2005; Takeuchi et al., 2008; Bouvet de Maisonneuve et al., 2009; Yokoyama and Takeuchi, 2009; Bai et al., 2010). A correlation between \( k_1 \) and \( k_2 \) exists, which can be fitted by a power law (Fig. 4b; Rust and Cashman, 2004). The correlation found between \( k_1 \) and \( k_2 \) can be reproduced by a simplified parameterization of these quantities proposed by Degruyter et al. (2012),

\[
k_1 = \left( \frac{f_0 r_b^2}{8} \right)^{m}
\]

\[
k_2 = \frac{f_0 r_b}{\phi}^{(1.3m)/2}
\]

with \( \phi \) the connected vesicularity, \( f_0 \) the throat/vesicle ratio, \( r_b \) (m) the vesicle radius, \( N_d \) (m\(^{-3}\)) the vesicle number density, \( m \) the tortuosity, and \( f_0 \) the friction coefficient. An example fitted by the power-law relationship \( k_1 = 5.6 \times 10^{-6} \kappa_0^2 \) is shown by the dashed line in Fig. 4b for representative values \( f_0 = 0.2 \), \( N_d = 1 \times 10^{22} \text{ m}^{-3} \), \( m = 3 \), \( r_b = 50 \). The trend itself is explained by the dependence of the pore space on the same parameters. The spread around the trend line is due to differences in throat/vesicle ratio, tortuosity, vesicle number density and roughness or factor/friction coefficient across the different samples. Rust and Cashman (2004) found that for values of \( k_1/k_2 \) exceeding the critical value, inertia will be significant in conduit flow models. Degruyter et al. (2012) demonstrated that inertia affects the transition from effusive to explosive eruptions, and that such transition is controlled by the ratio of a characteristic \( k_2 \) and the conduit radius. A small value of this ratio results in explosive eruptions.

5.4. The role of viscosity on permeability

Our data show that magmas with the same composition have strikingly different connected vesicularity–permeability values (e.g. MTN and CI samples); however, strikingly different magmas exhibit very similar connected vesicularity–permeability values (e.g. CI and KPT samples). For this reason, we suggest that similarities in permeability data can be generated by a balance between melt viscosity and crystal content as these will have a strong control on vesicle size and throat size and tortuosity (Wright et al., 2009; Degruyter et al., 2010a; Polacci et al., 2012).

We test this hypothesis by calculating the effect of crystals and dissolved water content on the viscosity of the CI, MTN, KPT and Ambrym magmas following Giordano et al. (2010) (Fig. 5, Table 2). From Fig. 5 it clearly emerges that Ambrym basaltic scoriae have the lowest melt and bulk (melt + crystals) viscosity. We believe that the lower bulk viscosity of these basaltic products may allow the development of large, interconnected vesicles with large vesicle apertures.


(Polacci et al., 2012) delivering the highest permeability values amongst the samples plotted in Fig. 4a. It also emerges that, although crystal content plays a non-negligible (up to a maximum of 0.6 log units) effect on the bulk viscosity of the samples illustrated in this study, water content could have an equally relevant effect (Fig. 5). In fact, although at the same water content the melt viscosity of KPT rhyolites is higher than that of MTN and CI trachytic melts, comparable values of calculated viscosity ($\eta \approx 4/\pm 1$; as these are calculations and not measurements they have an uncertainty of the order of 1 log unit) are obtained by assuming a dissolved water content of 7 wt.% for the KPT magma (Bachmann, 2010) ($\eta \approx 3$), 3 wt.% for the MTN magma (Piochi et al., 2008) ($\eta \approx 4$) and 4 wt.% for the CI magma (Piochi et al., 2008) ($\eta \approx 3.5$), and very low water content for the Ambrym magma (Allard et al., 2009) ($\eta \approx 3$). In this view, we suggest that the high permeabilities of the MTN samples could be generated by a combination of high magnetite microlite content (Fig. 2c, Table 2) and the effect of dissolved volatiles on melt viscosity (Fig. 5). The effect of microlites is to promote both vesicle nucleation and growth on and around them, eventually enabling vesicles to coalesce and form a network of microcrack-like, permeable gas pathways (Piochi et al., 2008). CI pumice samples have permeabilities that are similar to those of KPT pumice products (Fig. 4a). While the former are essentially aphyric, their melt viscosity being still low enough (Fig. 5) to allow a high degree of vesicle coalescence and to form interconnected vesicle pathways (Fig. 2a, b), the latter displays the highest melt and bulk viscosity (Fig. 5). We propose here that crystallinity of the KPT magma was sufficiently high to counter balance the viscosity effect and allow vesicle to grow, coalesce and form permeable vesicle pathways well visible in tube pumices (Bouvet de Maisonneuve et al., 2009; Degruyter et al., 2010a). This process of vesicle coalescence enhancement by high crystallinity has been modelled numerically (Huber et al., 2012) and it has been documented to occur in crystal-rich magma analogues (Belien et al., 2010); it is likely to play a fundamental role on the development of permeability in any crystal-rich magma (e.g., basaltic magmas such as Stromboli (Bai et al., 2011), or silicic mushes (Huber et al., 2012)). However, the occurrence of this process in crystal-rich natural magmas deserves further textural and experimental investigation.

The cases provided in this study are just preliminary examples of the role that viscosity may have on magma textures and permeability. However, in order to determine if a correlation between viscosity and permeability exists, accurate determination of both the original dissolved water (and other volatiles) content and melt viscosity is therefore necessary to establish the role played by rheology in controlling magma physical properties and, consequently, magma textural evolution. Future studies on this topic should be focused on detailed investigations of magma textures and their evolution with time under different conditions of deformation and cooling.

Acknowledgments

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References


Civetta et al., 1997; for BM from Melluso et al. (1995); for MTN from Piochi et al., 2008; for KPT from Bouvet de Maisonneuve et al. (2009); for Ambrym from Polacci et al. (2012).

in order to minimize and maximize relative viscosity, calculations were made following the procedure in Giordano et al. (2010) at strain-rates of $10^{-3}$ and $10^{-6}$ s$^{-1}$, respectively.

<table>
<thead>
<tr>
<th>Sample</th>
<th>$T(°C)$</th>
<th>Crystal content%</th>
<th>Min relative viscosity (log unit)</th>
<th>Max relative viscosity (log unit)</th>
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<td>Cl pf Sn</td>
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<tr>
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<tr>
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<tr>
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<td>35</td>
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</tr>
<tr>
<td>KPT</td>
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<td>35</td>
<td>0.33</td>
</tr>
<tr>
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<td>0</td>
<td>8</td>
<td>0.00</td>
</tr>
</tbody>
</table>

$^a$ eruptive T for CIs, CIMo, BM and MTN from Piochi et al. (2008); $^b$ T for KPT from Bachmann, 2010; $^c$ T for Ambrym from Allard et al. (2009).


