The Survival of Mafic Magmatic Enclaves and the Timing of Magma Recharge

Philipp Ruprecht[1], Adam C. Simon[2], and Adrian Fiege[3]

[1]Department of Geological Sciences and Engineering, University of Nevada, Reno, NV, USA.

[2]Department of Earth and Environmental Sciences, University of Michigan, Ann Arbor, MI, USA.

[3]Department of Earth and Planetary Sciences, American Museum of Natural History, New York, NY, USA.

Corresponding author: Philipp Ruprecht (pruprecht@unr.edu)

Key Points:

- Common survival times for mafic enclaves in felsic volcanic systems are centuries to millennia extending timescale records from minerals
- Mafic enclaves record only syn-eruptive processes in hot magmatic systems
- Mafic enclaves in plutonic systems may represent recharge histories of 10,000 to 100,000 years

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Abstract

Many intermediate to felsic intrusive and extrusive rocks contain mafic magmatic enclaves that are evidence for magma recharge and mixing. Whether enclaves represent records of pro-longed mixing or syn-eruptive recharge depends on their preservation potential in their intermediate to felsic host magmas. We present a model for enclave consumption where an initial stage of diffusive equilibration loosens the crystal framework in the enclave followed by advective erosion and disaggregation of the loose crystal layer. Using experimental data to constrain the propagation rate of the loosening front leads to enclave "erosion" rates of 10^{-5} to 10^{-8} cm/s for subvolcanic magma systems. These rates suggest that under some circumstances, enclave records are restricted to syn-eruptive processes, while in most cases enclave populations represent the recharge history over centuries to millennia. On these timescales mafic magmatic enclaves may be unique recorders that can be compared to societal and written records of volcano activity.

Plain Language Summary

Two major questions in volcano research are how magma chambers are built through time and how they are disrupted to cause volcanic eruptions. One piece of evidence that chambers are assembled by episodic magma addition from below (called "recharge") comes from mingled magmas, where mingling is expressed by the presence of two or more chemically distinct magmas. In particular, the more primitive magma in such mingled magmas is commonly present as discrete blobs, called mafic magmatic enclaves. These enclaves are often interpreted as evidence for recharge-triggered volcanic eruptions. However, they may also form during recharge episodes that are not associated with volcanic eruptions and instead only feed and sustain the magma chamber. Here we develop a model that estimates how long mafic magmatic enclaves survive in a chemically-distinct magma chamber to better understand how information drawn from enclaves informs the two major questions above. We find that under most common conditions, they survive for centuries to millennia. Therefore, the presence of enclaves is not explicitly evidence for a recharge-triggered eruption without studying them in greater detail. That detail can then potentially provide information regarding both the run up to eruption as well as magma assembly over centuries and millennia.

1 Introduction

Magmas have long been recognized as open systems; a notion supported by abundant signatures in the crystal record and magma (i.e., whole rock) chemistry (e.g., DePaolo, 1981; Davidson et al., 2007; Ruprecht & Wörner, 2007). The most direct evidence is the macroscopic presence of mafic magmatic enclaves (also referred to as quenched mafic inclusions) and crustal and mantle xenoliths (e.g., Bacon & Metz, 1984; Clynne, 1999; Ruprecht et al., 2012). Mafic magmatic enclaves evince incomplete mixing and hybridization where viscosity contrasts during the mixing of felsic and mafic magmas preclude stirring and stretching to the crystal scale and the removal of any macroscopic mixing evidence (Sparks & Marshall, 1986; Ruprecht et al., 2012). However, once mafic magmatic enclaves form, it remains an important question whether they get consumed through time, and if so, how consumption progresses. What is the characteristic timescale associated with enclave-size reduction that controls their long-term presence? The timescale of enclave-size reduction determines if enclaves document predominantly (1) an integrated record of recharge magmas into felsic magma systems or (2) pre- and syn-eruptive changes in intensive parameters of magmatic systems. In the latter case, long-term assembly and end-member contributions can only be inferred from bulk chemistry and individual crystal chemistry.

Past work addressing the physical processes of enclave assimilation focused on the survival of macroscopic (ultra-)mafic components in magmas and their incorporation in basaltic magmas. That work suggested that mixed in components get consumed within hours to days of their introduction (Sachs & Stange, 1993; McLeod & Sparks, 1998). Thermal conditions in the hot basaltic magmas and extensive stirring due to the low viscosity of the melts ensure near instantaneous removal of diffusional gradients in the melt. The removal of compositional and thermal gradients drives melting and dissolution, which effectively erases physical evidence of compositionally distinct components. In felsic host magmas, thermal conditions and magma dynamics are also important for enclave survival. For example, mafic magmatic enclaves in plutons provide evidence that enclaves can survive a super-solidus history of a pluton. Yet, plutons often also show extreme macroscopic homogeneity suggesting that homogenization and enclave removal has to occur to some degree given the life time of millions of years for those systems (Coleman et al., 2004). In eruptive magmatic systems that are still stored at elevated temperatures (well above the solidus for periods of time), sufficient energy may be available to partially melt and disaggregate enclaves.

A renewed interest has emerged to understand mafic enclave survival in felsic host magmas in response to the growing research that targets magma process timescales, such as mixing, ascent, and eruption (Turner & Costa, 2007). Here, we develop a model for enclave size-reduction combined with data from experiments that juxtapose basaltic andesite and dacitic magmas to explore what controls mafic enclave survival.

2 Field observations related to enclaves and their formation

There are two processes that need to be distinguished when discussing the survival of enclaves: 1) what are the conditions needed for them to form?, and 2) once enclaves form, what is needed to preserve or destroy them? The focus of this paper is on the second question as their formation is controlled by compositional and thermal contrasts (ΔC, ΔT) between recharge and host magma (Marshall & Sparks, 1986), and the dynamics of mixing (Andrews & Manga, 2014; Hodge & Jellinek, 2012; Ruprecht et al., 2012). Ruprecht et al. (2012) argued that while ΔC-ΔT is fundamentally important, the dynamics and physicochemical interaction of mafic with felsic magma leads to time-dependent changes in magma viscosity that can promote enclave formation or allow for effective hybridization with a spectrum between these end-members. In particular, mineral chemistry reveals that, e.g., host magmas can contain enclaves, which contain multiple crystal populations of one or more mineral phases, as well as individual crystals that were themselves disaggregated from enclaves and are preserved in the host (Beard et al., 2005; Martel et al. 2006; Humphreys et al. 2009; Ruprecht et al., 2012). Thus, magmas range from completely hybridized (i.e., no enclaves) to partially hybridized (i.e., host and/or recharge magma have mixed and do not retain end-member compositions, while also containing enclaves) to no microscope/crystal-scale mixing and only mingling in the form of enclaves. In addition to the presence of mafic phenocryst phases being dispersed in host magmas, high anorthite (An) plagioclase microlites interpreted to be of mafic origin (Martel et al., 2006; Humphreys et al., 2009; Ruprecht et al., 2012) suggest that disaggregation is an effective process in removing the macroscopic evidence for mixing. An additional important observation in the microlite record is that their high An cores tend to be rounded, reflecting resorption prior to rim growth of low An plagioclase following the dispersal in the felsic magma (see figure 1 in Martel et al. 2006).

Enclave textures can vary drastically and so do the variations in composition and temperature associated with the end-member magmas that drive enclave formation. The range in ΔC and ΔT associated with the two mixing magmas and the relative volume contribution during mixing give rise to a diverse physicochemical and fluid dynamic response that leads to variations in overall crystallinity, a diversity in preserved crystal sizes, as well as the presence and absence of spatial gradients from interiors to enclave rims (e.g., quenched glassy rinds versus more crystalline enclave centers). In general, the significant temperature drop a mafic magma experiences as it comes in contact with cooler felsic host magma generates rapid crystallization of a fine matrix dominated by plagioclase with a subordinate amount of pyroxene, olivine, and oxide (Bacon & Metz, 1984). Second boiling within the enclaves as the enclave crystallizes drive vesiculation and additional plagioclase crystallization leading to many enclaves being almost completely crystallized with interstitial melt pockets making up <40 vol.% of the enclave (Browne et al., 2006). Whether the melt within the enclaves quenches during mixing depends on whether the “race” to the glass transition temperature during cooling is faster than chemical changes to the melt composition related to crystallization, which will lower the glass transition temperature. This race is partially controlled by the thermal evolution during mixing, which is a function of the absolute temperature difference between the mixing magmas and their relative proportion (Ruprecht & Bachmann, 2010). Quenching of the mafic melt is possible if host magmas are close to eutectic temperatures and dominate the mass balance; only in those cases can mafic to intermediate composition melts be quenched and fall below the glass transition temperature (Giordano et al., 2008). The presence of quenched margins in erupted mafic magmatic enclaves may point to fast transport to the surface where quenching can progress rather than quenching in the magma reservoir. Such fast transport is also supported by diffusion profiles in minerals (e.g. Humphreys et al. 2009; Ruprecht and Cooper, 2012). However, if recharge is volumetrically significant, then temperatures of the mixtures are well above any glass transition temperature and crystallization proceeds with the microlite-rich enclaves gaining internal strength as the rheologic lock up is exceeded due to high crystallinity. This latter case is a common occurrence of enclaves and is the focus of this contribution.

3 Physico-chemical processes of enclave consumption

Given the internal strength of a mafic, high-crystallinity enclave (<40% interstitial melt) that develops a crystal framework (Martin et al. 2006), enclave consumption is not simply a function of continued stirring and stretching in the host magma. Instead, the breakup of enclaves requires an interplay of phase change, thus weakening of the internal strength, combined with magma flow driving shear and disaggregation. Previous models focused on the wholesale melting of xenoliths combined with melt flow removing diffusive boundary layers (Sachs & Stange, 1993). However, this process removes any crystal evidence through melting and dissolution of the mafic magma, a condition that is not met for most mixed and mingled magmas that contain abundant enclaves. Instead, individual crystals that originated from a mafic end-member commonly remain dispersed in the host (Clynne, 1999; Browne et al., 2006; Ruprecht et al., 2012). Thus, enclave consumption is the combined process of (a) partial dissolution of microlites and microphenocrysts combined with volatile exsolution that loosens and weakens the crystal framework and (b) melt flow and shear that leads to the detachment of individual crystals or smaller crystal aggregates from the main enclave. Such removal mechanisms may be texturally difficult to identify in natural samples as a few microns to tens of microns can be sufficient for efficient loosening of the crystal framework.

The disaggregation of any aggregate, whether it is silicate minerals or other phases that are part of a connected cluster of particles, can occur by one of two modes: 1) rupturing where the new aggregates are reduced in size by a factor on the order of 2 and 2) erosion where shear and lift forces overcome the attractive forces for individual particles and enclave-size reduction is controlled by the rate at which individual particles are loosened (progressive dissolution into the enclave) and by the relative movement of enclave and surrounding melt (Powell & Mason, 1982; Ottino et al., 1999). Loosening of the particle framework happens in response to chemical disequilibrium between the host magma and the mineral assemblage in the mafic enclave. In particular, the plagioclase microlites that grow in response to cooling and second boiling during enclave formation are vulnerable to partial dissolution. Given that they make up most of the framework that holds enclaves together, it is their dissolution that ultimately leads to the erosion of the enclave and the release of mafic phenocrysts to the host magma (Fig. 1).

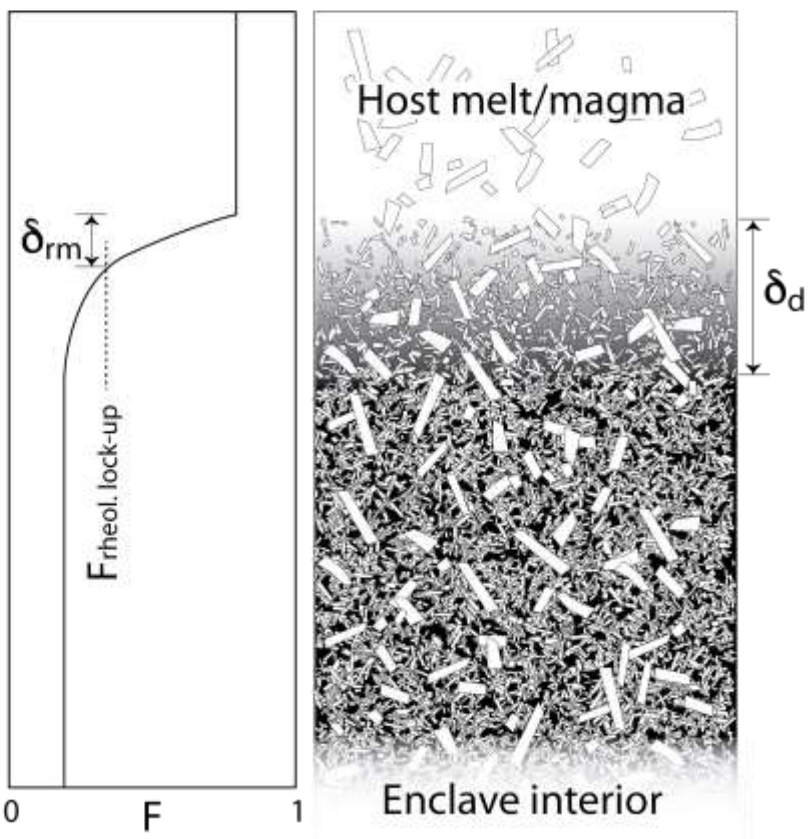

Figure 1. Conceptual model of enclave consumption and general model describing our underlying experiments (Figure 2A). If temperature conditions are such that enclave minerals (most importantly plagioclase, which is the only phase shown for simplicity) are melted, a boundary layer (δ_d) forms that is diffusion controlled and advances following $\sqrt{(Dt)}$. Within a convective regime, the boundary layer δ_d will be reduced by δ_{rm}, which is the instantaneous removal of material with crystallinity below the rheologic lock-up. Mafic plagioclase (and other)

phenocrysts will be added to the host melt. F is the melt fraction with the rheologic lock-up melt fraction ranging between 0.4 and 0.6.

Our model for enclave consumption is therefore twofold and starts after enclaves have formed and established a textural framework that includes phenocrysts, microlites, melt, and volatile bubbles in response to the local thermal equilibration of mafic and felsic magma. Dissolution advances into the enclave, which increases the interstitial melt fraction in the enclave above the rheologic lock-up (>0.4-0.6; Marsh, 1989). This is a diffusive process with a characteristic square-root relationship between length- and time-scale, and has been described previously (Tsuchiyama, 1985; 1986; Sachs & Stange, 1993; McLeod & Sparks, 1998). The physical removal of the emerging low-crystallinity boundary layer then occurs in a second step that is instantaneous as soon as the melt fraction decreases below rheologic lock-up. The exact conditions that govern the switch between diffusion-controlled plagioclase microlite dissolution and advection-driven removal of the boundary layer remains poorly constrained as such boundary layer problems have yet to be studied in much greater detail (Ottino et al., 1999). We assume diffusion operates first for timescales T_L such that individual plagioclase microlites become sufficiently loose to be removed from the enclave or crystal aggregate. Once T_L is reached, enough loosening of the crystal network has occurred and the boundary layer is removed by advection. The advective-driven size reduction is therefore a function of lengthscale δ_{rm} associated with T_L. The lengthscale of enclave consumption is thus best described by a diffusive-advective model:

$$\boldsymbol{x(t) = k\sqrt{t}\ \ \text{for}\ t < T_L\ ;} \qquad \textbf{Eq. 1}$$

$$\boldsymbol{\delta_{rm}(T_L) = k\sqrt{t}\ \ \text{for}\ t = T_L\ ;} \qquad \textbf{Eq. 2}$$

$$\boldsymbol{x(t) = k_L t\ \text{for}\ t > T_L\ ;} \qquad \textbf{Eq. 3}$$

With $k_L = \delta_{rm} / T_L$, where T_L is the time to reach a localized (crystal-scale) melt fraction that exceeds the rheologic lock-up and k_L is the dissolution rate when the advective regime takes over.

4 Experimental constraints on microlite dissolution and advective-controlled erosion rates

Our model is motivated by recently published time series experiments that explore the physico-chemical processes at mafic-felsic magma interfaces (Fiege et al., 2017; for more details of these experiments see also the supporting information). The experiments were conducted at 1,000 ˚C and are especially relevant for cases where the mass balance ratio of mafic recharge to host magma is large. The experiments exhibit the development of a systematic dissolution front in the mafic magma that is extensively crystallized with microlite-size plagioclase and subordinate mafic minerals and oxides (Fig. 2A). Analysis of the advancement of this dissolution front reveals a square root relationship (consistent with Eq. 1 of our model) that holds for a range of potential lock-up melt fractions (0.4-0.6; Fig. 2B).

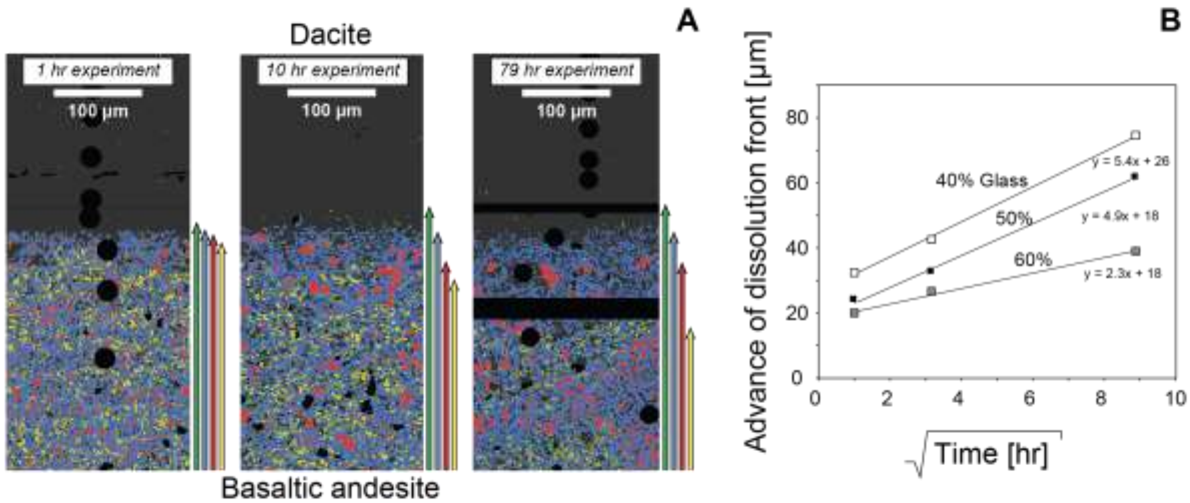

Figure 2. A) False color wavelength-dispersive X-ray maps from the timeseries experiments of Fiege et al. (2017). Gray: Silicate glass, green: spinel, blue: plagioclase, red: orthopyroxene, yellow: clinopyroxene. The arrows next to each map indicate the presence of the respective mineral phase in the basaltic andesite. B) Estimated advance of the dissolution front within the basaltic andesite of the three timeseries experiments. Mineral fractions change according to simple diffusion-controlled scaling. The basaltic andesite becomes progressively glass-rich through time documented by the advancing front of 40, 50, and 60% glass. The non-zero intercept is either a result of imprecise locating of the interface or due to heating rate effects. For more information on the image processing and associated uncertainties see extended data presentation in the supplement.

The crystal dissolution rates determined from these experiments are faster than experiments that measured the dissolution rate of a large plagioclase crystal at high temperature. (Tsuchiyama, 1985; Fig. 3a). Comparison is difficult for two reasons. First, previous experiments looked at 1D dissolution of individual, large plagioclase crystals in diopside-albite-anorthite melts at >1200 ˚C, while our experiments are poly-mineral aggregates dominated by plagioclase with a large surface area of melt-plagioclase contact (Fig. 2A). Second, our experiments are performed at lower temperatures, more realistic for natural systems, while being placed in large chemical disequilibrium. When we estimate the evolution of the 100% melt front, our rates are comparable to the ones from Tsuchiyama (1985). However, we argue above (section 3) that 100% dissolution is not required for enclave consumption.

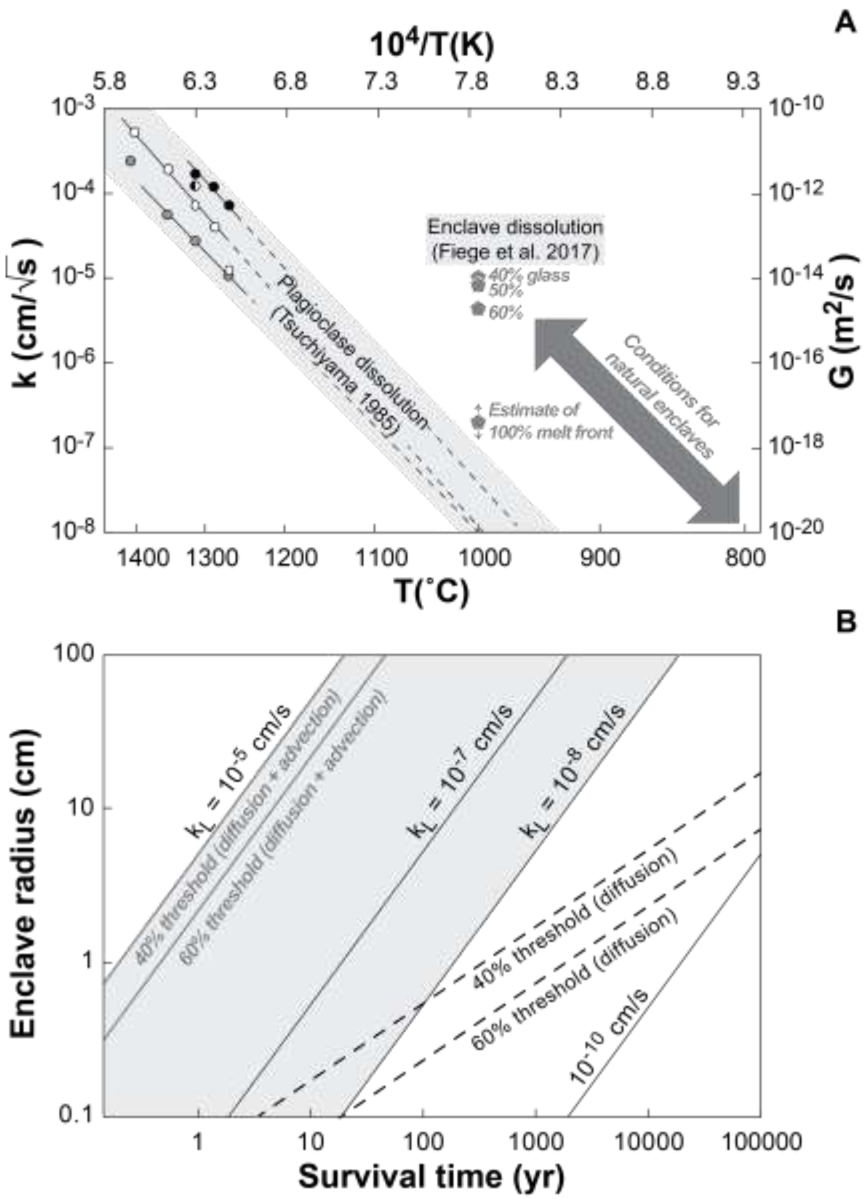

Figure 3. A) Comparison of poly-mineral experiments that include plagioclase dissolution (Fiege et al. 2017) with single crystal dissolution rates by Tsuchiyama (1985). Gray pentagons are the slopes in figure 2. The advancement of the 100% melt front is estimated from experiments in Fiege et al. (2017), but spatial scales for these experiments are too short to constrain the rates with low uncertainties. A conversion to a conventional diffusion rate *G* is provided. B) Enclave erosion and survival times for a range of advective "erosion" rates. Assuming that advection becomes important within hours of diffusion-controlled dissolution provides an estimate for the rates of advective removal (k_L). Conditions for natural systems suggest advective removal rates between 10^{-5} to 10^{-8} cm/s. For comparison, enclave survival in the absence of advection (dashed lines) is shown for rates derived from experiments by Fiege et al. (2017).

These experimental results allow us to explore the timescale(s) of enclave survival that significantly exceed the timescales for pure melting under hot basaltic conditions, and provide some constraints on enclave survival and preservation. Our results (Fig. 2 and 3a) indicate that the diffusive front for 40-60% melt advances at 10^{-5} to 10^{-6} cm/√s (equivalent to 10^{-14} to 10^{-16} m^2/s when cast as a more conventional diffusion/dissolution rate G). Such rates are likely limited to high temperatures that are reached for rare cases where mafic input is large and the system is thermally re-equilibrating slowly to more intermediate temperature conditions. It therefore represents a case for the more rapid consumption of mafic magmatic enclaves. More moderate conditions, where magmas may experience temperatures and plagioclase dissolution at 800-900 ˚C are more common for many andesitic to dacitic systems (Murphy et al., 2000; Holtz et al., 2005; Ruprecht et al., 2012) and those conditions may persist for longer timescales. Extrapolating rates from the experiments by Tsuchiyama (1985) and Fiege et al. (2017) suggest diffusion-controlled rates of 10^{-7} to 10^{-8}cm/√s (→ $G = 10^{-18}$ to 10^{-20} m^2/s) for temperatures of 800-900 ˚C. If one assumes that advective processes take over within hours of diffusion-controlled dissolution, we can estimate the advection-controlled removal rate k_L to vary between 10^{-5} to 10^{-8}cm/s for common andesitic to dacitic systems that frequently erupt mingled magmas with cm- to dm-size enclaves (Fig. 3b and Eq. 3). Such rates imply that enclaves consumed by an erosive process survive no longer than 100 to 1000 years. Any additional size reduction process, e.g., by rupturing, which is sometimes recorded in volcanic and plutonic systems and results from melt infiltration and the presence of large stresses (Laumonier et al., 2014), further reduces the survival times. Of course, enclave survival is also a function of enclave sizes (Fig. 3). Our model implies that if erosion is the dominant process, survival times are directly proportional to enclave size. Thus, systems with very large mafic magmatic enclaves (e.g., 1 m radius) may survive significantly longer. However, field evidence in the form of partially-ruptured enclaves, abundant specifically in larger enclaves, suggests that size-reduction by rupture is enhanced in the larger enclaves and, therefore, even those may quickly get reduced to sizes where erosion dominates.

5 Discussion

Once conditions are met for enclave formation, the question is whether they will survive past the lifetime of the magmatic system or whether they become part of a hybridized mixture through time. Those conditions may be met during many recharge events, which are likely to occur on the order of every tens to hundreds of years (Ruprecht & Wörner, 2007). Moreover, residence times of long-lived magma bodies in the crust often exceed 100 kyr (Reid et al., 1997). Thus, if survival exceeds the lifetime of the magmatic system, erupted magmas should be full of different enclave populations in magmatic systems that juxtapose evolved host and primitive recharge magmas in the crust. Even if recharge magmas are similar over such timescale, it is plausible to envision large diversity in enclave textures and compositions. While diversity is present in enclaves in many evolved lavas, they typically show only the presence of a few different populations (Clynne, 1999; Browne et al., 2006). One potential explanation is that enclaves are removed from the magma system through time. The survival of dispersed enclaves in magmas is important because if enclaves survived indefinitely, they could be used to understand the long-term assembly of magmatic systems. If they are lost relatively quickly from the rock record then enclave populations may provide important information on just the pre-eruptive changes in the magmatic system.

Enclave removal may occur through settling. While some field evidence in plutons suggests that enclaves may settle under some recharge conditions and internal dynamics of the magma body (Wiebe & Collins, 1998), plutonic records are inconsistent with efficient wholesale removal and deposition. Despite the greater density of mafic magmatic enclaves relative to the surrounding evolved magma, any minor convection in a viscous magma will keep them in suspension over long times as they either drift in the magma or operate as passive tracers (Burgisser et al., 2005). Further, in water-rich magmatic systems, enclaves are often vesiculated, and the exsolved volatile phase imparts buoyancy to the enclaves and inhibits settling. We therefore argue that only the largest enclaves can easily be lost by settling. The majority of cm- to dm-size enclaves remain dispersed in the host magma for extended times and interact with host magma with which they are not in equilibrium.

Under some thermal conditions enclaves may become macroscopically largely unrecognizable because they deform viscously into thin sheets during magma transport. Such flattening has been observed in nature (e.g. in the Adamello batholith; John & Blundy, 1993). However, the formation of magmatic fabric that erases enclave records require substantial flow (Paterson et al. 1998) and therefore is not an effective mechanism to completely erase a macroscopic record of mafic magmatic enclaves. This was recently tested numerically (Burgisser et al., 2020). Deformation is most effective during initial mafic-felsic interaction and enclave formation (Hodge & Jellinek, 2012; Andrews & Manga 2014) after that viscous deformation may sometimes lead to textures that resemble flow banding, but it is unlikely to completely erase the macroscopic record of mixing and enclave formation throughout the rock.

Alternatively to settling and viscous deformation, the thermodynamic disequilibrium in which enclaves find themselves may drive complete melting and dissolution. Mineral chemistry is often still significantly out of equilibrium with respect to an evolved melt (e.g., high An content plagioclase) and such minerals can respond to this disequilibrium by melting and dissolution; this is particularly common for plagioclase and even visible in microlites (Martel et al. 2006). However, if dissolution and melting was the lone process in removing the enclave record, then no crystals of the recharge magmas should survive, which is inconsistent with field observations (Clynne, 1999; Beard et al., 2005; Browne et al., 2006; Humphreys et al., 2009; Ruprecht et al., 2012). Instead, we argue that enclave survival times are controlled by a combination of dissolution and physical disaggregation. Here, dissolution is especially effective on the small microlites with large surface-to-volume ratios that experience a significant size reduction and that can be liberated easily from an enclave or any other crystal-rich aggregate, while preserving the larger phenocrysts. The stage of loosening by dissolution is important as it promotes the complete disintegration of enclaves to individual minerals. If disaggregation alone operates on the enclaves, then we would expect that micro-enclaves persist much longer as stresses on the enclave during stretching and stirring diminish with the crystal cluster size. While micro-enclaves in the form of glomerocrysts and crystal clusters have been described in various studies, they are subordinate to the dispersal of individual microlites (Martel et al., 2006; Humphreys et al., 2009; Ruprecht et al., 2012).

As a result, survival times in volcanic systems may be as short as a few years ($k_L \sim 10^{-5}$ m/s; mingling under hot conditions and small enclave sizes). Thus, in very hot systems, enclaves potentially only record the lead up to- and syn-eruptive history. In more moderate subvolcanic conditions our model suggests centuries to millennia for their complete removal ($k_L \sim 10^{-7 \text{ to } -8}$ m/s). Such survival times are consistent with a partial record of recharge preserved by mafic magmatic enclaves. Most intermediate to evolved magmatic systems that erupted magmas with mafic magmatic enclaves therefore provide more than syn-eruptive process information. Instead, multiple populations of enclaves may constrain compositional diversity that is being added to the magma system over centuries and millennia instead of a complex history of syn-eruptive magma assembly. By detailed bulk and mineral analysis of these populations we may be able to study the lead up to an eruption in greater detail as individual populations may represent different time markers in the lead up history. As a result, they also potentially extend temporal records from crystals to longer timescales as they add the timescale of disintegration to mineral equilibration. Moreover, such timescales suggest that for many magmatic systems mafic magmatic enclaves represent an integrated record over multiple eruptions and therefore they may be uniquely sensitive to providing constraints on the cycling in between eruptions. However, mafic magmatic enclaves are unlikely to provide a meaningful record of the entire recharge history for long-lived magma systems.

Whether reactive processes at the interface of mafic magmatic enclaves described here are also important in the plutonic record is complicated by the prolonged cooling recorded in plutons. The reactive process occurs shortly after mingling and if it is not completed (i.e., enclaves are disintegrated), any reactive front will be overprinted in the plutonic record and reactive boundary layers will be difficult to preserve or to infer. In some cases glassy rinds do survive (Wiebe 2016) and suggest that the mafic-felsic mass ratio and temperature difference are so that glass transition temperatures are reached. However, there are also examples of reactive boundary layers in plutonic settings regarding mafic enclaves. They involve gradual changes in texture and chemistry, as well as rinds rich in, e.g., biotite (Chen et al., 2009; Farner et al., 2014; Michel et al., 2017). Thus, extrapolating dissolution rates to temperatures of long-term storage conditions of plutons is difficult. Advective erosion rates are likely much smaller (potentially k_L $< 10^{-10}$ m/s). We can only speculate on timescales of enclave survival in plutonic systems. For advective erosion rates of 10^{-10} m/s, we predict that many episodes of enclave formation are erased over a plutons prolonged live and only enclaves produced through recharge within the last 10,000 to 100,000 years are preserved.

6 Conclusions

Given the current view that magma systems grow incrementally by a complex interplay of recharge, differentiation, assimilation, and melt segregation (Bachmann & Bergantz, 2004; Coleman et al., 2004; Hildreth et al., 2004), it suggests that either not all recharge is mafic and some systems experience recharge only in the form of evolved magmas, or that mafic recharge is often hybridized effectively and only individual crystals provide testimony to the open system behavior. Nonetheless, given that mafic recharge is central – why do we not see more evidence for mafic magmatic enclaves? They are present in some lavas, but just as common is their absence. Some plutons have no enclaves, whereas others contain abundant enclaves, and even others only have zones of mafic magmatic enclaves. This suggests that they are only partially retained – that processes lead to their removal. Our model is consistent with this notion. If partial retainment of enclaves is the dominant mode of preservation, enclaves lend themselves as unique components in magmatic systems to study the magma assembly and build up to eruptions on timescales of centuries to millennia, complementing the short record often retained in mineral diffusion studies (Costa & Chakraborty, 2004; Shamloo & Till, 2019) and long-term integrated record of plutons (Paterson et al., 2016). The presence of enclaves cannot be explicitly used as evidence for a recharge-triggered eruption without additional constraints. While timescales from individual crystals can be reconciled with modern continuous monitoring signals, we suggest that detailed investigation and extraction of timescales from enclave populations can be reconciled with historic and societal records of volcano activity.

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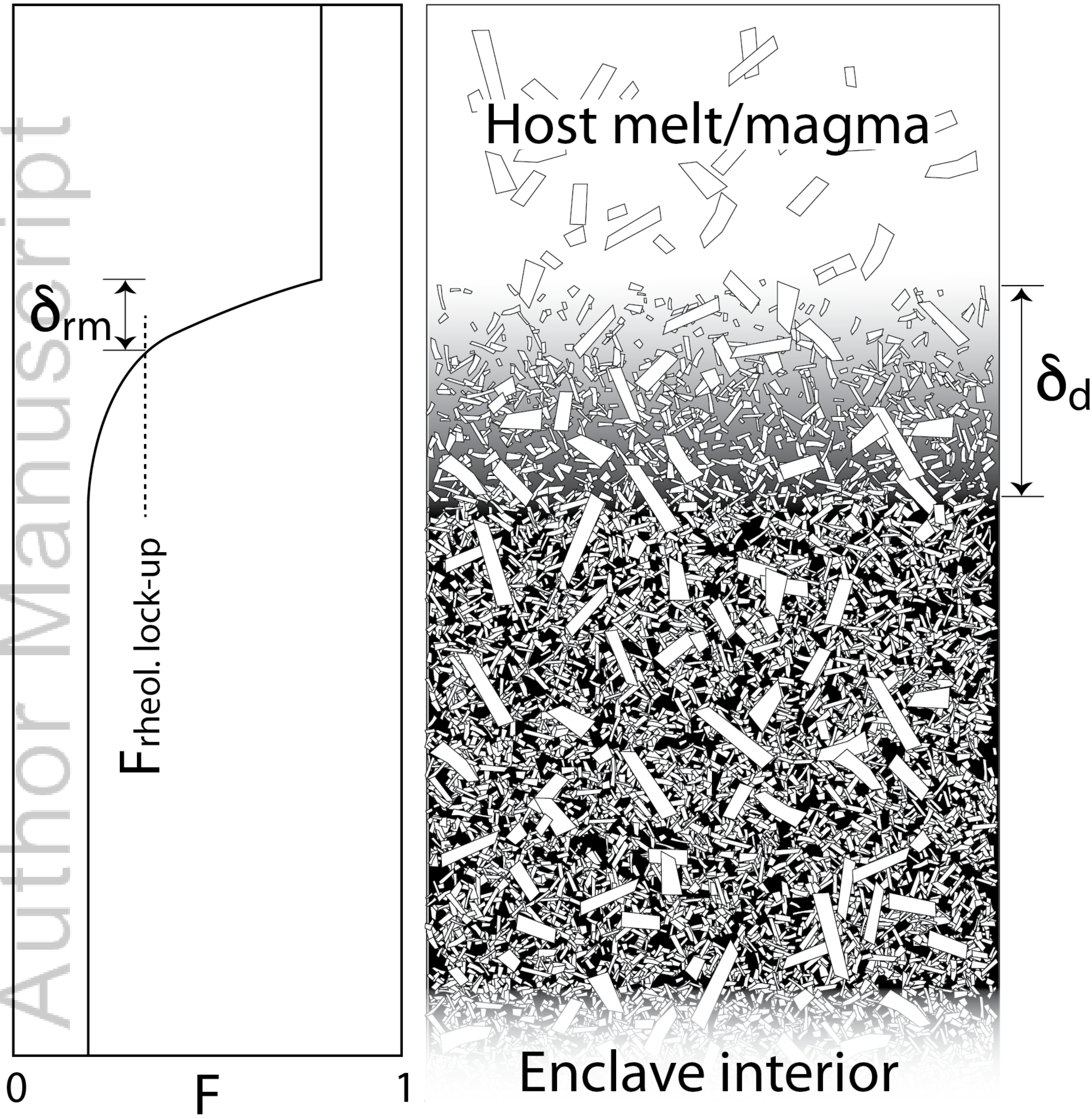

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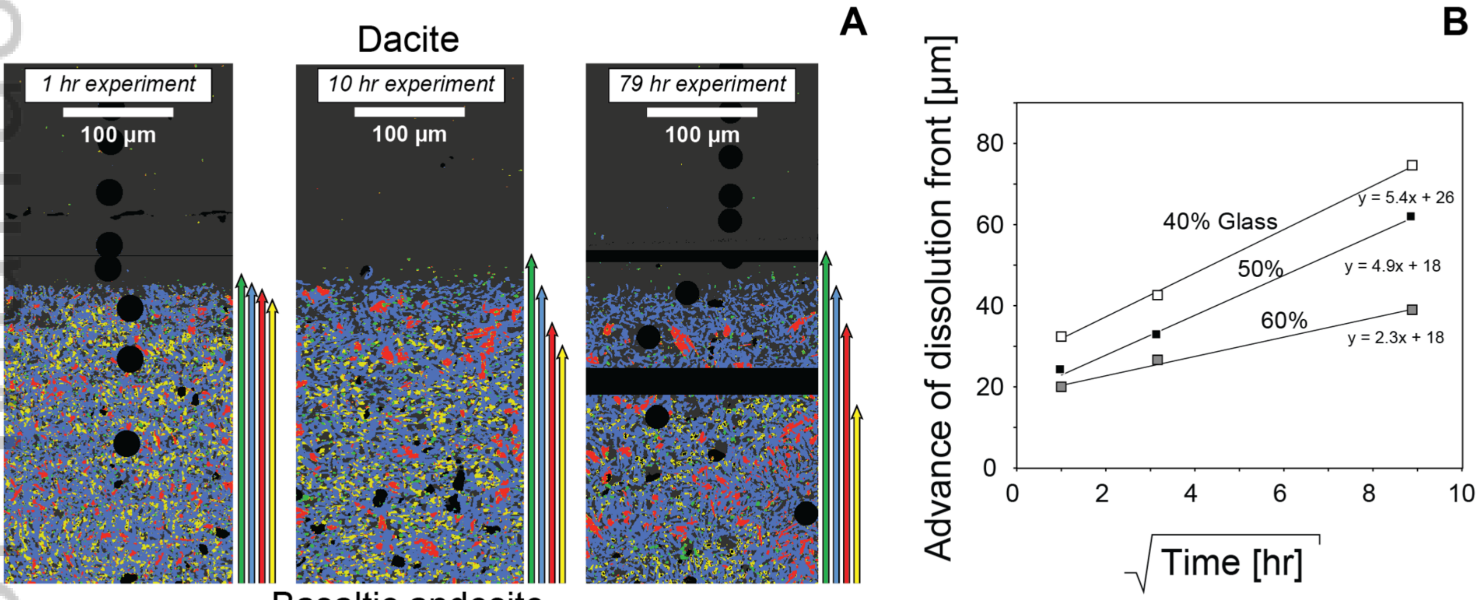

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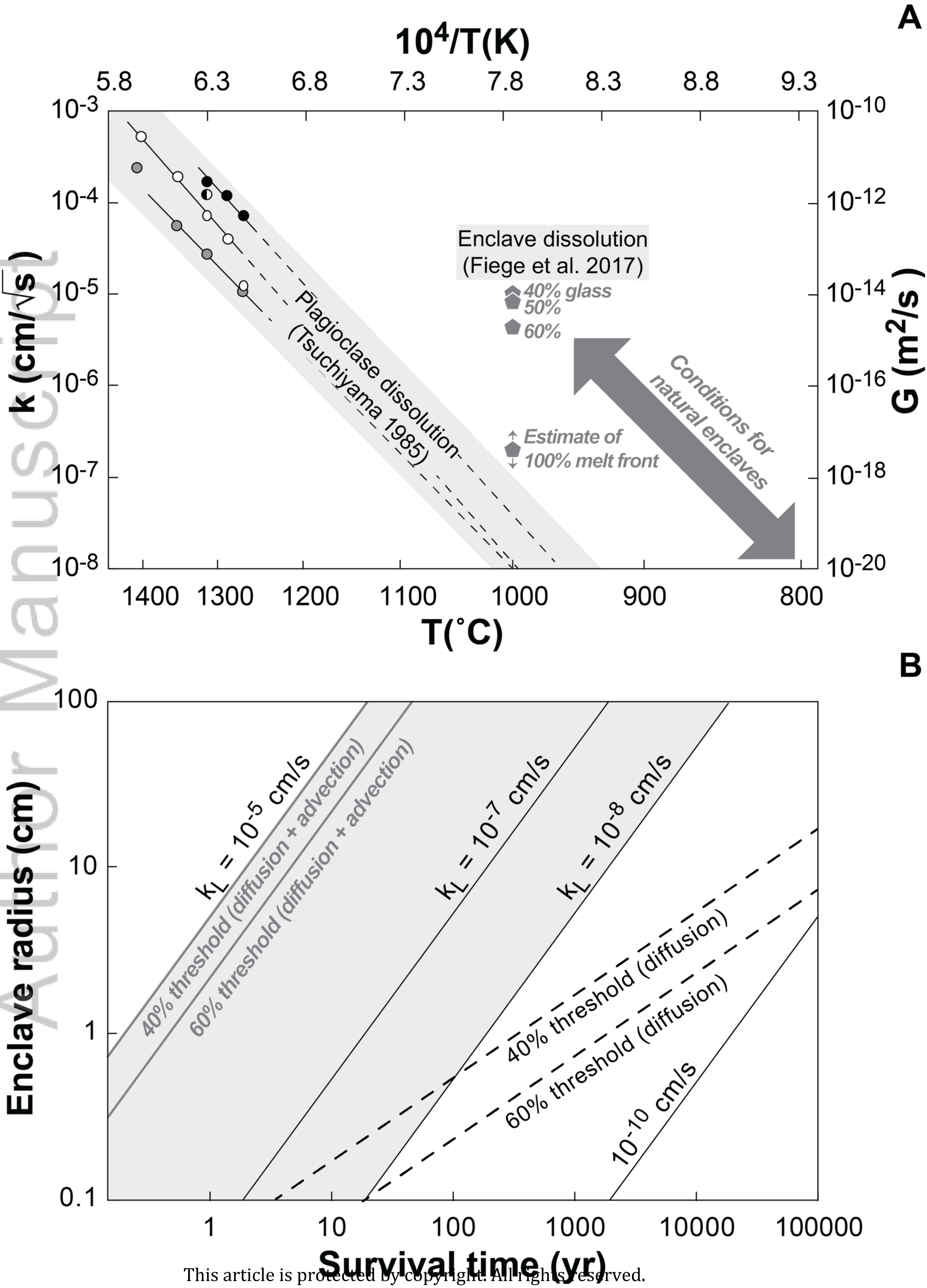
A
B
10^4/T(K)
5.8
6.3
6.8
7.3
7.8
8.3
8.8
9.3
k (cm/√s)
G (m²/s)
T(˚C)
Enclave dissolution
(Fiege et al. 2017)
40% glass
50%
60%
Plagioclase dissolution
(Tsuchiyama 1985)
Conditions for
natural enclaves
Estimate of
100% melt front
Enclave radius (cm)
Survival time (yr)
40% threshold (diffusion + advection)
60% threshold (diffusion + advection)
40% threshold (diffusion)
60% threshold (diffusion)
k_L = 10^-5 cm/s
k_L = 10^-7 cm/s
k_L = 10^-8 cm/s
10^-10 cm/s