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Thermal Manipulation of Magma Boundaries: Advancing Controls on Fluid Flow via the Krafla Magma Testbed (KMT)

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ABSTRACT

Magmatism generates the strongest heat fluxes in the Earth's crust and powers geothermal resources. There is, however, a long-standing paradox as to the occurrence and extent of fracturing in magma and superhot rocks deemed ductile, which would prevent efficient fluid exchange. Drilling of IDDP-1 (at Krafla caldera, Iceland) was associated with complete fluid loss as reservoir rock temperature rose from ca. 400 to 850 °C in a 30-meter thick rock-magma boundary. Tripping out and back of the drill string allowed upwelling of magma, hot enough to flow in and clog the bottom of the borehole. This behavior provides evidence for the importance of thermal controls to open or maintain pathways for extracting fluids along magma boundaries.

What is envisioned here is cooling-induced contraction fracturing of a hot rock/ magma, rather than fracturing by increasing fluid pressure above the least principal stress in cold rocks, as practiced in EGS. Laboratory measurements provide accurate constraints for the modeling of thermal stimulation, showing that the magnitude of cooling and the thermo-mechanical properties of rocks and magmas are primary controls on the width of fractures and the magnitude of reservoir permeability. The resultant width of thermal fractures in this high-temperature boundary region is greater than those developed in conventional geothermal reservoirs, and thus access to this resource may require a new generation of proppants and/ or thermal stimulation practices. Laboratory tests on hot rocks show that damage accumulation from successive rupture cycles may contribute to the ingression and/ or displacement of fragments along fracture intersections, acting as natural proppants. Tests on magma show that fracture healing may be prevented by lowering the temperature and injecting proppants with slow diffusive properties. Whilst material properties suggest that cooling is the solution to maintaining fluid pathways, field surveys and mineralogical investigations of the eroded vestiges of hydrothermal systems show the important role of secondary mineral precipitation in clogging fluid pathways as a result of pressure and temperature changes (mostly cooling). Thus field constraints, drilling information and laboratory descriptions must be combined to determine the optimum temperature window in which fluid exchange may be maximized for prolonged timescales. This framework will be developed in this study and tested in the Krafla Magma Testbed (KMT), with the aim to shift the current paradox to a paradigm whereby temperature regulation of magma body boundaries will enable use of this new magma energy resource.

1. INTRODUCTION

Fluid circulation in the Earth's crust is central to a spectrum of geological processes, influencing the development of both geohazards and georesources. In our attempts to increase our understanding of hazards (e.g., earthquakes, landslides, volcanic eruptions) and optimize extraction of resources (e.g., hydrothermal fluids, oil, gas, etc), comprehensive knowledge of fluid flow in Earth materials becomes imperative. Fluid flow in geothermal reservoirs is controlled by the permeability of rocks, which is dominantly regulated by the network of faults and fractures (especially if the rocks have low permeability). To improve the access and extraction of fluids from rocks, different practices are commonly employed, including fluid pressure gradient fluctuations, fracking and thermal stimulation (Siratovich et al., 2015), to name but a few common practices. Here we address the case of thermal stimulation and the resultant fracture networks available for fluid flow in (and near) magmatic environments to discuss the enriched value of this new energy frontier.

Geothermal energy currently accounts for <1% of energy consumption worldwide; yet, studies suggest the potential is 10-100x the current generation (World Energy Council, 2010). Supercritical fluids along magma boundaries are more energetic (than conventional fluids) and may be a solution to this prospect. Searches for such high-enthalpy fluids have been conducted, as targeted by the Iceland Deep Drilling Project (IDDP). Drilling of the first IDDP borehole (IDDP-1) halted before reaching the supercritical zone as it serendipitously intersected a magmatic body at ~2100m (Elders et al., 2014). Importantly, drilling along the magmatic aureole was accompanied by no fluid return, revealing that circulation increases drastically towards the magma margin (Scott et al., 2015), countering the commonly held view that fracture pathways for fluid circulation may not be sustained at depth (Mortensen et al., 2014) owing to ductile flow (cf, Violay et al., 2012). Taking advantage of this unique knowledge of magma location, properties and behavior, and of the architecture of the reservoir rock from decades of monitoring by volcanologists and geothermal exploitation by Landsvikjun, we created the Krafla Magma Testbed (KMT; kmt.is), which aims to establish the first magma observatory – an international, open access, scientific platform to advance magma research via drilling activities over the course of the next 2-3 decades. This frontier undertaking will enable direct, in situ sampling, instrumentation and manipulation of magma in the Earth's crust, in order to monitor its evolution through time and inform the future models of the Earth (Eichelberger et al., 2020; Ludden et al., 2020; Papale et al., 2020). Providing that a careful engineering strategy is devised (Hólmgeirsson et al., 2018) and a cautious risk assessment is performed (Ilic et al., 2020), the prospect of accessing magma has become possible due to the extensive knowledge that we have gained of this system and the volcanic material properties. Herein, we review the physical and mechanical properties of volcanic materials, and explore the thermo-mechanical properties that prevail in superhot systems surrounding magma reservoirs, to outline the considerations that must be made to ensure the success of KMT.

2. PHYSICAL PROPERTIES OF VOLCANIC ROCKS

2.1 The porosity and permeability and volcanic rocks

Volcanic provinces comprise complex successions of physically, chemically and structurally heterogeneous rocks, which may exhibit variable states of physical coherence, alteration, and saturation with fluid (Lavallée et al., 2018; White and Houghton, 2006). Micro-structurally, volcanic rocks commonly exhibit heterogeneous textures with varying crystallinity and fabrics, surrounded by interstitial glass (metastable phase abundant in magmatic scenarios where magma has been forced to cool relatively quickly). The porosity, and thus storage capacity, of volcanic rocks can vary widely (0-97% porosity) and, as a result, the permeability of volcanic materials can range by >10 orders of magnitudes (Figure 1a; Eggertsson et al., 2018; Heap et al., 2017; Lamur et al., 2017; Mueller et al., 2005). The common abundance of micro-fractures is essential in allowing fluid flow in dense volcanic rocks (Mueller et al., 2005); yet, the tendency of volcanic materials to fracture or fragment during magma transport or eruptions as well as the tendency of the Earth's shallow crust to develop pervasive macroscopic fracture networks, imposes primary controls on fluid flow in volcanic provinces (Eggertsson et al., 2018; Lamur et al., 2017).

2.2 Permeable pathways in volcanic rocks: open fractures, tuffisites and mineralization veins

Recent experimental studies in which macro-fractures were generated in laboratory specimens using the Brazilian indirect tensile testing method have quantified the impact of fracture width on fluid flow (Heap and Kennedy, 2016; Lamur et al., 2017); a constraint which, once upscaled (Farmer, 2002; Heap and Kennedy, 2016), allows consideration of rock micro- as well as macro-structures in fluid flow simulations. Lamur et al. (2017) found that the presence of a macro-fracture primarily enhances the permeability of volcanic rocks with porosity less than ca. 15 vol.% (Figure 1b); in more porous rocks, the presence of fracture imparts moderate additional pathways for fluid flow. Interestingly, a complementary experimental study investigating the impact of multiple orthogonal fractures in laboratory specimens of volcanic rocks showed that the presence of two fractures may not necessarily increase permeability by a similar amount as the first fracture generated (at the laboratory sample scale); yet, the results showed that comminution and offset of rough fractures are more common in such fracture networks, which can then act as natural proppants that prevent fracture closure when pore pressure decreases (or effective pressure increases), thus ensuring the prolonged connectivity of fluid pathways (Figure 1c).

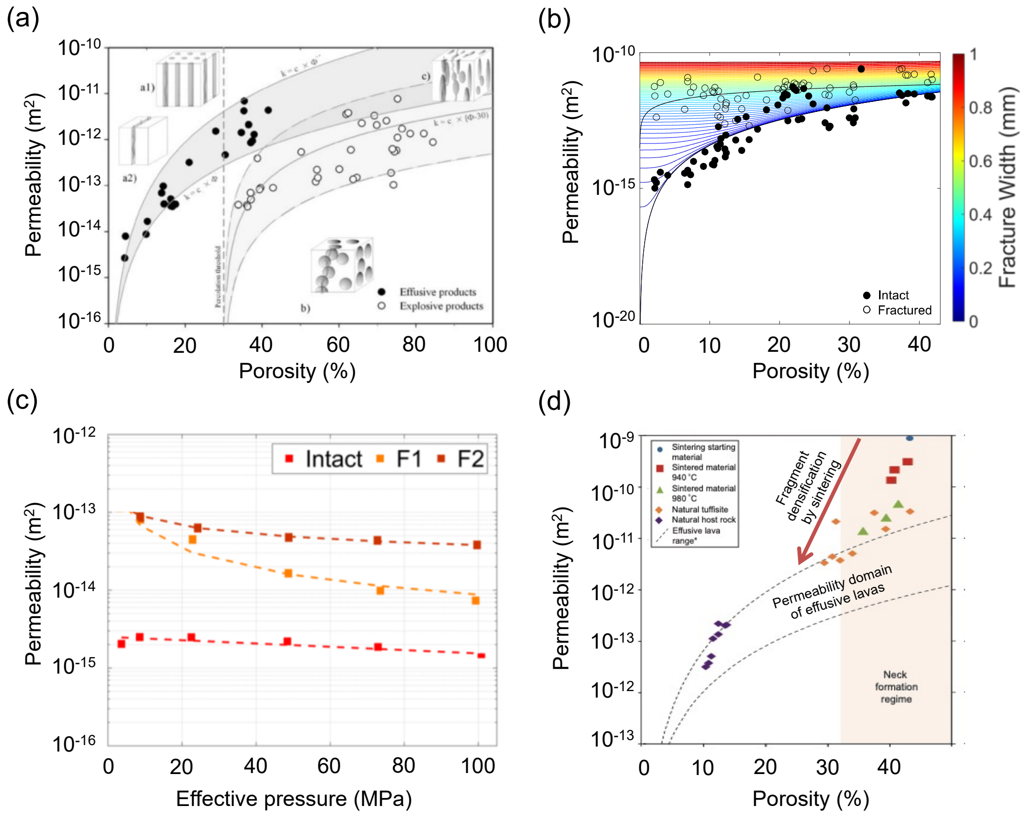

Figure 1: Permeability of volcanic materials. a) Permeability of effusive versus explosive volcanic rocks (modified from Mueller et al., 2015). b) Permeability of intact volcanic rocks versus volcanic rocks hosting one fracture of variable width (modified from Lamur et al., 2018). c) Permeability of intact rocks versus rocks containing one (F1) or two (F2) orthogonal fractures (from Eggertsson et al., 2018). d) permeability of tuffisite veins in andesites versus that of fragmental volcanic rocks sintered at 940 and 980 °C (modified from Kendrick et al., 2016) showing offset of granular-filled fractures from the porosity-permeability trend for effusive lavas in (a).

Fractures may infill with fragmental particles deposited from energetic gas fluxes, leading to the formation of tuffisite veins (e.g.,Castro et al., 2012; Owen et al., 2019; Tuffen and Dingwell, 2005). [Note that this is somewhat akin to cataclasites in fault structure yet often lacking the shear indicators.] As the particles may indurate or sinter (in hot volcanic regions), tuffisite formation has the ability to either prop open fractures (e.g. Kendrick et al. 2016), maintaining their permeability (Figure 1d) or to shut permeable networks entirely (Farquharson et al., 2015; Kolzenburg et al., 2012); the timescale of this densification depends on temperature and the mineralogical attributes of the phases present in the system (Ryan et al., 2018; Wadsworth et al., 2017). In other instances, fluids flushing through fracture networks may precipitate secondary hydrothermal minerals (e.g., calcite, chlorite, quartz, epidote) owing to changes in pressure, temperature and chemistry of the system (Fisk et al., 1998; Hedenquist et al., 1998; Richards, 2011; Williams-Jones and Heinrich, 2005). Quartz and calcite veins are common byproducts of such a process, which may form in a single or multiple precipitation events (Ramsay, 1980). Detailed mineralogical studies have shown that precipitation may occur from the edge of the veins inward as well as outward from crystal nuclei present in fractures (Jebrak, 1997; McNamara et al., 2016; Oliver and Bons, 2001). Whereas in the former case, mineral growth may keep the permeability of a fracture to a low nominal value (Bons et al., 2012), crystal growth in the second case may act as proppants, keeping fractures open for efficient fluid transfer. Yet, if hydrothermal mineralization concludes and seals the fracture, permeable flow ceases (Dobson et al., 2003). Importantly, the highly sensitive nature of hydrothermal mineralization to pressure, temperature and chemistry of the systems implies that mineral breakdown may take place, leading to rejuvenation of the permeable network (Heap et al., 2017). In active hydrothermal systems, exploited for geothermal energy, this may lead to construction and destruction of fluid flow barriers, promoting shifts in the architecture of hydrothermal convection cells on variable timescales. Therefore, understanding the formation of hydrothermal veins and palaeo-permeability of fossil hydrothermal settings is critical to establishing the longevity of fractured geothermal reservoir rocks (Gomila et al., 2016; Sibson, 1996).

3. THERMO-MECHANICAL PROPERTIES OF VOLCANIC ROCKS

3.1 The mechanical properties of volcanic rocks

The heterogeneous make-up of porous volcanic rocks – in particular their different ratios of pores to micro-fractures – makes their mechanical behavior highly variable (Heap et al., 2015a). The strength of volcanic rocks has been shown to decrease with porosity, irrespective of volcanic type and chemistry; yet, for a given porosity, volcanic rocks can exhibit a wide range of compressive strengths (Figure 2a), depending on their mineralogy, crystallinity and fabrics. Similarly, the tensile strength of volcanic rocks decreases with porosity, irrespective of whether tensile stresses are generated by internal pore pressure accumulation (Alatorre-Ibargüengoitia et al., 2010; Spieler et al., 2004) or external forcing (Figure 2b; Harnett et al., 2019; Hornby et al., 2019). Investigations into the impact of strain rate have shown that the strength (Figure 2c) and Young's modulus (Figure 2d) of volcanic rocks increase as a function of strain rate (Coats et al., 2018; Lavallée et al., 2018). However, if volcanic rocks contain zeolites, studies have found that the compressive strength may decrease with the presence of water owing to stress corrosion (Heap et al., 2018b). Micro-mechanical modeling of rock compressive strength has shown that volcanic rocks do not abide to either the pore-emanated crack model (Sammis and Ashby, 1986) or the wing-crack model (Ashby and Sammis, 1990), but a yet-to-be constrained combination of the too, owing to the coexistence of convoluted vesicle geometries and micro-fracture networks (Coats et al., 2018; Heap et al., 2014); empirically, experimental studies have found that permeability may be a good descriptor of volcanic rock strength (Schaefer et al., 2015), thus providing a proxy during geothermal exploration to link fluid flow to strength based on drilling data.

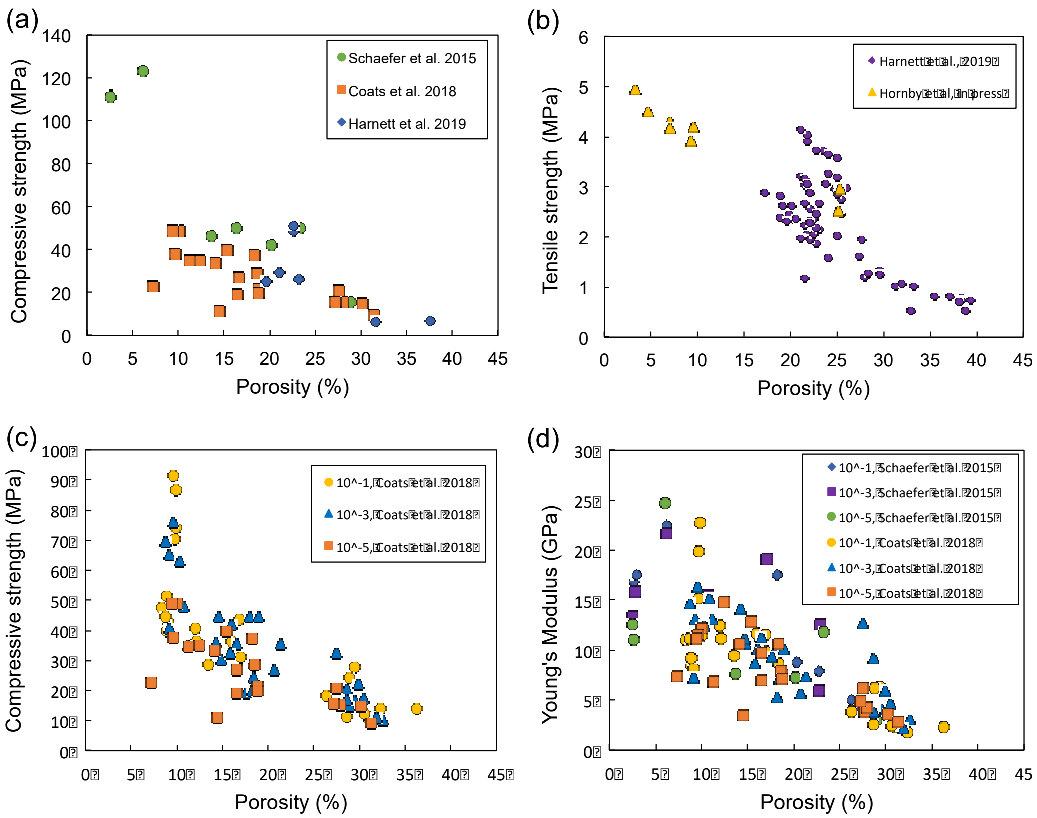

Figure 2: Mechanical properties of volcanic rocks. a) Compressive strength tests of dacites (Coats et al., 2018), andesites (Harnett et al., 2019) and basalts (Schaefer et al., 2015) at a strain rate of 10^{-5} s^{-1}. b) Fragmentation criterion (cf, tensile strength) determined as pore overpressure to cause porous material rupture during rapid decompression in shock tube apparatus (modified from Alatorre-Ibargüengoitia et al., 2010). c) Compressive strength of dacite lavas as a function of strain rate and porosity (modified from Coats et al., 2018). d) Young's modulus of dacite lavas (Coats et al., 2018) and basalt (Schaefer et al., 2015) as a function of strain rate and porosity.

The heterogeneous nature of volcanic rocks makes their deformation regime highly sensitive to, and dependent upon, local stress conditions. In particular, the switch from brittle (referring to localized dilatant behavior) to ductile (referring to pervasive compactional behavior) deformation is strongly influenced by the porosity of rocks, with high porosity rocks exhibiting low strengths and transitioning from brittle to ductile failure mode at low effective pressure (Figure 3). Porous rocks which may, *a priory* behave in a ductile manner, can transition to a brittle mode of deformation if they compact sufficiently during prolonged straining, as for instance, due to volcanic, magmatic or geothermal drilling activities (see blue arrow on Figure 3c). As experimental studies have shown that the strength of volcanic rocks increases with strain rate (Figure 2c; Coats et al., 2018; Lavallée et al., 2018; Schaefer et al., 2015), we also expect that the depth of the brittle-ductile transition in volcanic rocks can shift depending on local strain rates experienced due to fluctuations in tectonic forces and geothermal exploration activity. This may be accentuated if pore pressure, which counteracts local external stresses (and thus decreases the effective mean stress), fluctuates and shifts conditions towards the brittle field, causing embrittlement of the materials (see orange arrow on Figure 3; Farquharson et al., 2016).

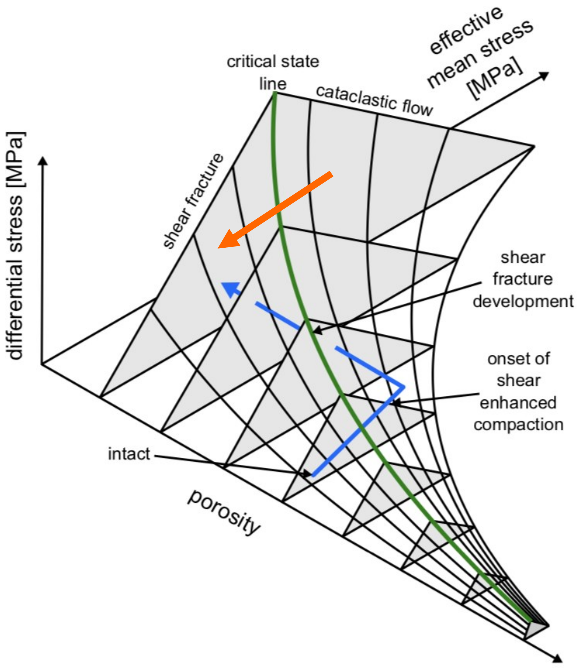

Figure 3: Schematic of differential stress, effective mean stress and initial connected porosity. The critical state line (the green solid line) divides the transition between compactant (cataclastic flow) and dilatant (shear fracture) regimes as a function of porosity (Modified from Heap et al., 2015). The orange arrow shows a scenarios in which pore pressure increase causes a decrease in mean effective pressure, which prompt a shift in macroscopic deformation mode from ductile to brittle. The blue arrow shows an example in which sustained compaction of porous rocks may promote rock densification and a resultant shift in rheology from ductile to brittle.

3.2 Thermal expansivity, thermal stressing and thermal micro-cracking

Temperature changes play an important role in the formation and evolution of volcanic materials. All volcanic rocks form through cooling, which induces crystallization and vitrification; both of which cause a volume decrease, taking place over extensive periods of time. We have few constraints on the magnitude and rate of this process, yet cooling contraction of the ~20 m thick lava flows erupted in 1991 and 2000 at Hekla (Iceland) are still contracting at ~2 mm/year and they are modeled to be still on the order of a few hundred degrees Celcius today (Wittmann et al., 2017). The densification associated with these physical processes and importantly the mismatch in contraction of different mineral phases (commonly with anisotropic thermal expansivity), imparts strong local stresses that cause micro-fracturing of rocks (Meredith and Atkinson, 1983; Meredith et al., 2001). Yet in the lifespan of volcanic provinces, which may experience numerous magmatic events and in the case of geothermal exploitation, rocks are subjected to multiple cooling and heating cycles which may strongly modify their uptake of fracture damage and thus, degree of coherence.

The thermal expansivity of volcanic rocks has been the focus of numerous studies and has been found to be dependent on the mineralogy and porosity of the materials. Richter and Simmons (1974) demonstrated that the coefficient of thermal expansion is approximately that calculated from the constituent minerals' single crystal values. Cooper and Simmons (1977) assessed the dependence of the coefficient of thermal expansion on rock porosity, suggesting that thermal expansivity is reduced with the presence of micro-fractures which provide void space for mineral expansion. Siratovich et al. (2015) tested this hypothesis on variably porous and fractured andesites from the Rotokawa geothermal field, finding little change and a lack of a relationship with total porosity. The study found no changes (or hysteresis) in thermal expansion coefficient between heating and cooling cycles, yet observed that the coefficient of thermal expansion increases ~2 fold between ambient conditions and the local reservoir temperature of ~350 °C. Importantly, the significant thermal expansivity of volcanic rocks is argued to enhance the generation of fractures during natural temperature fluctuations or thermal stimulation practices.

Thermal stressing of volcanic rocks has been extensively studied (Browning et al., 2016; Heap et al., 2018a; Heap et al., 2015b; Heap et al., 2012; Heap et al., 2014; Kendrick et al., 2013). Laboratory specimens have been subjected to cycles of heating and cooling whilst monitoring acoustic emissions, associated with resultant micro-cracking. The data show that laboratory specimens are not all equally affected by thermal stressing as we note only a mild generation of acoustic emissions resulting from micro-cracking in some materials (Heap et al., 2014). If the original material contains micro-cracks, then few cracks are generated during thermal stressing and the strength and permeability of the rocks remains largely unchanged (Eggertsson et al., 2018); yet, poorly fractured rocks are more affected by thermal stressing as we note that the resultant permeability can continue to increase over several heating/cooling cycles (Eggertsson et al., 2018). As a result, micro-fractured volcanic rocks only exhibit a mild thermal Kaiser effect as they need to be subjected to higher temperature than previously experienced in order to undergo thermal cracking (Heap et al., 2014).

3.3 Thermal-sensitivity of mineralogical assemblages

Volcanic rocks, easily altered due to the common presence of metastable glass, easily hydrated and crystallized when dwelling at low to high temperatures over long timescales (Castro et al., 2008; Castro et al., 2009; Cerling et al., 1985). So, volcanic rocks commonly contain thermally liable minerals such as xeolites, clays and carbonates (Heap et al., 2012; Heap et al., 2013). Laboratory studies have shown that these minerals may efficiently breakdown when subjected to moderately high temperatures, leaving void space upon heating. If these thermally liable minerals are abundant, their breakdown may have drastic consequences, such as in hyaloclastite, commonly enriched in smectite in geothermal fields such as Krafla (Weaver et al., 2019). Importantly, mineral breakdown creates pore space, building pathways for fluid flow (Heap et al., 2012; Weaver et al., 2019). In addition, if this enhanced permeability enables outgassing (Balashov and Yardley, 1998), empty voids generated by the breakdown of minerals can cause a decrease in the Young's modulus and strength of these rocks, leading to premature failure without changes in the stress field (Heap et al., 2012).

3.4 Mechanical properties of volcanic rocks at elevated temperature

Previous studies have found that the compressive strength of volcanic rocks keep a similar inverse relationship with porosity, however the strength increases with temperature (Figure 4; Coats et al., 2018; Heap et al., 2018a; Lavallée et al., 2018; Schaefer et al., 2015), provided that the rock-forming minerals do not breakdown at high temperature (Heap et al., 2012). Similarly the Young's modulus (Lamur et al., 2018) and the tensile strength (Hornby et al., 2019) of volcanic rocks increases with temperature. These relationships hold so long as the material remains within the brittle regime, i.e. temperatures are kept below the solidus or the glass transition temperature (Tg ≈ 700-800 °C, depending on chemistry, or deformed so rapidly that their deformation rate exceeds the relaxation timescale of the melt (see section 4). In particular, the brittle-ductile transition of glassy volcanic rocks at high temperature can shift to lower pressure conditions (than at ambient temperature), owing to increased diffusional processes that enhance viscous relaxation above the glass transition, and which may favor compaction over dilatant rupture (Violay et al., 2012); yet, such a transition is rate (or time) dependent (Coats et al., 2018; Dingwell and Webb, 1989, 1990; Lavallée et al., 2018; Lavallée et al., 2007; Lavallée et al., 2008). These studies highlight that caution must be exerted when modeling large-scale behavior using mechanical data obtained via conventional testing at ambient conditions; at high temperature, any modeling efforts must ensure careful consideration of thermo-kinetic constraints.

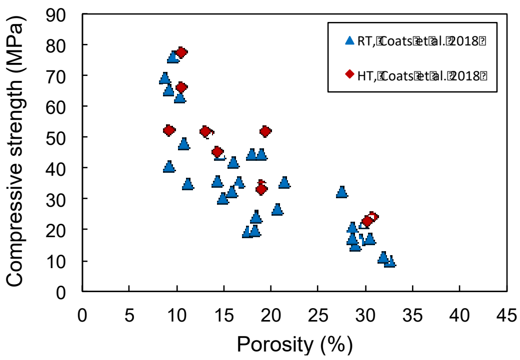

Figure 4: Compressive strength of porous andesite at 20 °C (RT) versus 900 °C (HT) deformed at 10^{-3} s^{-1} (to remain in the brittle deformation regime). The data shows a mild strengthening of volcanic rocks with temperature (modified from Coats et al., 2018).

4. THERMAL FRACTURING OF MAGMA

4.1 Silicate melt rheology and thermal fracturing

Magma comprises silicate melt with variable amounts of crystals and gas bubbles. Silicate melts are viscoelastic bodies abiding to the concept of structural relaxation (Dingwell and Webb, 1989, 1990). At high temperatures (and low viscosities) or upon long timescales, a melt structure may relax an applied stress and flow; however at lower temperature (and thus high viscosity) or upon short observation (or deformation) timescales, the melt struggles to relax the applied stress and behaves elastically, as a glass. If stresses are sufficient the melt structure may rupture. The thermo-kinetic nature of silicate melts' viscoelastic properties means that constraints on the rheology of silicate melts and glasses require consideration of both temperature and timescales (or reciprocally, strain rates). Thus in situations in which magma would be cooled (naturally or anthropogenically via the incursion of drilling fluids), the melt phase may be driven through the glass transition at a temperature concordant with the cooling rate experienced (Gottsmann et al., 1999; Gottsmann et al., 2002); that is, with faster cooling rate, the glass transition would be intersected at higher temperature with the result that the volume of the melt structure turned to glass would be larger than if cooling had been experienced more slowly. In nature, silicate melts have been observed to fracture due to rapid cooling, as preserved in perlites (Denton et al., 2012) as well as in the entablature of columnar jointed rock formations (Forbes et al., 2014). If the temperature increases above the glass transition, or the strain rate slows, the glass may relax (to a melt) and fractures may heal at a rate proportional with the melt viscosity (Lamur et al., 2019). Thus understanding the thermal history (i.e., cooling rates and timescales) is central to estimating the development and longevity of fractures in silicate melts, present in magmatic environments.

4.2 Thermal fracturing of crystal-bearing magma

As the cooling of magmatic bodies generally tends to be slow, magmas commonly contain abundant crystals, forming a mush near their margins. These near solid environment are more prone to rupture at lower stresses and strain rates than their fully molten counterparts (Gottsmann et al., 2009). Contraction associated with crystallization and subsequent cooling can cause cracking and it is rather common for plutons to exhibit columnar joints across their bodies. Recent thermo-mechanical experiments were developed to constrain the degree of undercooling required to develop columnar joints in micro-gabbroic rocks from Seljavellir, Iceland (Lamur et al., 2018). The study found that the magma must cool below the solidus before accumulating any thermal stress (for the cooling rates tested; i.e., ≤10 K.min^{-1}), and needs to reach some 90-130 °C below the solidus temperature before rupturing (Figure 5). The importance of constraining the temperature of columnar jointing is that from this point forward, further cooling causes the widening of fractures which enhances permeability, allowing ingression of fluids; thus cooling no longer proceeds via thermal conduction alone, but is accentuated by fluid advection, until chemical alteration modifies the rock and fluid pathways (e.g., Schiffman et al., 2014).

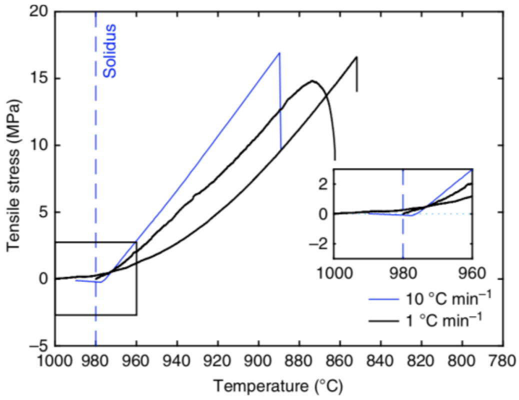

Figure 5: Accumulation of thermal stresses during cooling of micro-gabbro with locked-in dimensions as experienced by a segment of lava between the centre of two adjadent column, contracting until a joint develops. The data shows that cooling below the solidus (blue dashed line) leads to contraction which near-linearly accumulates tensile stress until rupture; some 90-130 °C below the solidus (Modified from Lamur et al., 2018).

4.3 Assessing the evolution of thermal cracks in cooling crystal-bearing magmas and rocks

Here we used the relationship developed experimentally for columnar jointing (Figure 5) to shed light on the widening of fractures and development of permeability of similarly jointed rocks. Considering a dense rock with a permeability of 10^{-20} m^2 and fracture spacing of ~50 cm, along with measured temperature-dependent coefficient of thermal expansion and Young's modulus, we model the widening of fractures in gabbroic near-magma environments to find that thermal contraction could generate fractures up to ca. 6 mm wide if cooled from magmatic temperature to drilling fluid temperatures of ca. 80 °C – a magnitude much greater than cooling of rocks in the cool, shallow parts of hydrothermal systems (e.g., cooling from 300 °C; Figure 6a). Accordingly, the greater propensity of near-magmatic rocks to open wide fractures would promote the generation of fluid pathways some 2-3 orders of magnitude more efficient at channeling fluid circulation (Figure 6b). Thus we advocate that near magmatic environments are ideal target for enhanced geothermal energy production, provided that they are accessed safely.

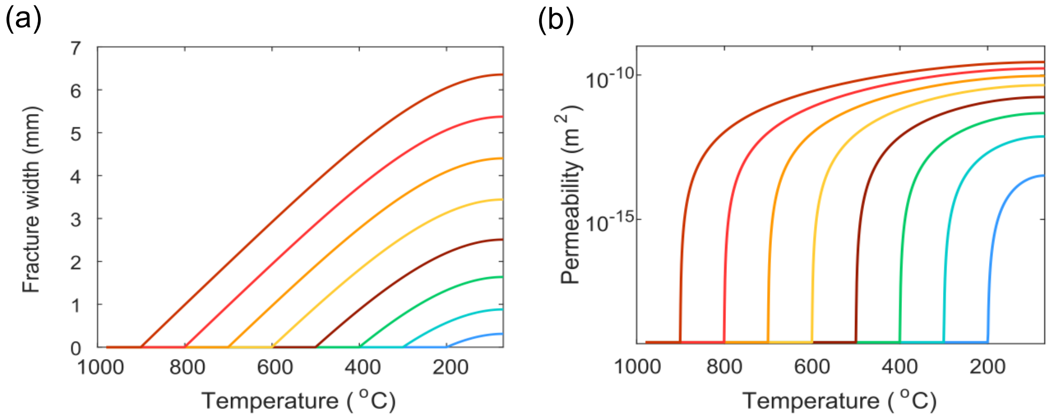

Figure 6: Evolution of a) fracture width and b) permeability following thermal fracturing of volcanic rocks. For a range of starting temperature scenarios (different colored lines), a) fracture width increases near-linearly during cooling until <200 ºC, with a material starting at 1000 ºC generating fractures wider than cooling rocks from a conventional well at a temperature of a few 100 ºC. Simultaneously, b) permeability development associated with fracture widening in each scenarios (color-coded as in (a)) shows a rapid rise as the fracture initiates. Modeled using the columnar jointing model of Lamur et al., (2018)

5. CONCLUSIONS

Here we review the physical and mechanical properties of volcanic rocks, highlighting the importance of physico-chemical attributes (mineralogy, porosity, permeability) in controlling materials' response to mechanical and thermal stressing. The review highlights that volcanic rocks behave contrastingly at high temperature, and that the important consideration of timescale is necessary in such systems. Cooling of hot volcanic rocks along magmatic aureoles (e.g., 800~1200 °C) leads to substantial contraction, which can generate larger cracks than the cooling of initially cooler rocks (e.g., 300-400 °C) commonly targeted in conventional geothermal fields. Yet, the longevity of such fractures may be thwarted by the secondary mineralization over prolonged periods of precipitation. Thus careful consideration of the system chemistry along with the thermo-kinetic properties of the volcanic materials and exploration of mechanisms to maintain fracture permeability are essential to ensure sustained fluid flow to maximize energy production. We must be mindful that the prospect of accessing magma energy relies on careful safety considerations of geological risks and engineering practices along with extensive simulations of system response scenarios – a strategy to be established by the Krafla Magma Testbed.

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