

Supporting Information for “Experimental evidence of reaction-driven fracturing in ultramafic rock induced by CO₂ mineralization”

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S1 Methods; Sample Preparation and Experiment Setup

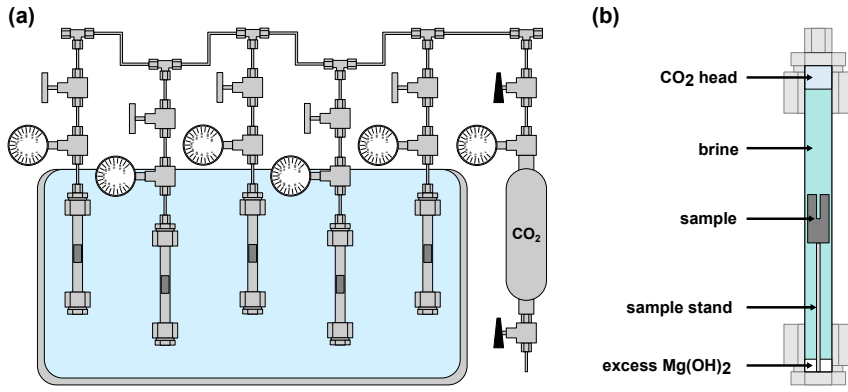


Figure 1. (a) Schematic of reactor stainless-steel tubes submerged in the temperature-controlled water bath. Each reactor tube is 15 cm tall and 10 mm in diameter. (b) Schematic of a single reactor tube with a sample submerged in a supersaturated magnesium hydroxide brine. Head of CO₂ was pressurized to 3.5 MPa at the top of the reactor and undissolved magnesium hydroxide deposited on the bottom of the reactor.

Table 1. List of samples and corresponding conditions used in this study.

Sample	Aqueous Solution	Temperature	Residence Time
A	CO ₂ + DI H ₂ O + Mg(OH) ₂	90° C	2 days
B	CO ₂ + DI H ₂ O + Mg(OH) ₂	90° C	4 days
C	CO ₂ + DI H ₂ O + Mg(OH) ₂	90° C	11 days
D	CO ₂ + DI H ₂ O + Mg(OH) ₂	90° C	18 days
E	CO ₂ + DI H ₂ O + Mg(OH) ₂	90° C	32 days
I	CO ₂ + DI H ₂ O	90° C	11 days
J	CO ₂ + DI H ₂ O + Mg(OH) ₂	20° C	11 days

S2 Methods: Digital Volume Correlation of X-ray Scans

SPAM employs a correlation-based approach to compute local voxel (3D pixels) displacement fields (Stamati et al., 2020). This allows for characterization of mechanical deformation and morphological changes within the reacted samples. Computations using SPAM utilized a half-window spacing of 6 voxels and a node spacing of 8 voxels, achieving approximately 80% convergence in displacement calculations. The original CT volumes are approximately $1000 \times 1000 \times 1500$ voxels.

S3 Methods: Raman Spectroscopy and SEM-EDS

Raman spectroscopy was performed using a ThermoScientific DXR3 Raman Microscope equipped with a 532 nm laser and 1800 and 2400 groove gratings, covering a spectral range of 50–3500 cm^{-1} . The laser was carefully focused on individual crystals to minimize signal interference from adjacent phases (Vulchi et al., 2024). Raman spectra were compared with reference mineral spectra from the RRUFF Raman Database.

Scanning Electron Microscopy (SEM) and Energy Dispersive Spectroscopy (EDS) were subsequently used to confirm Raman identifications and analyze features too small for reliable Raman analysis. Samples were examined at the e-MAGE Laboratory of the Bureau of Economic Geology using a Zeiss Sigma field emissions FE-SEM equipped with an Oxford X-Max 50 Silicon 50 mm^2 drift EDS detector (SDD) operated at 15 kV accelerating voltage and a 8 mm working distance. Samples were coated with carbon prior to imaging.

S4 Results: Intensity Change and Displacement Fields of Samples A, B, I and J

The following figures show the calculated change in gray value (intensity) and displacement fields for: samples A and B (early reaction times: 2 and 4 days), and for control samples I and J (no added magnesium hydroxide and low temperature, respectively).

S5 Results: Raman Spectroscopy

Figure 6 summarized measured Raman Spectroscopy results. Additional details are provided in the main article.

S6 Discussion: Bridging Geochemical Reactions and Fracture Mechanics

Reaction-driven cracking caused by carbonate mineral precipitation has been proposed as a key mechanism for creating new permeability pathways and enabling fluid–rock reactions to progress toward completion, even in low-permeability rocks (Kelemen et al., 2013, 2020). Although magnesite is the most thermodynamically stable magnesium carbonate phase, its precipitation is kinetically inhibited under ambient conditions and commonly requires temperatures $> 80^\circ\text{C}$ (Saldi et al., 2009). At lower temperatures, hydrated magnesium carbonates such as hydromagnesite and nesquehonite preferentially precipitate. However, these hydrous phases are mechanically weaker, raising questions about their ability to generate stresses sufficient to induce fracturing within the rock.

To estimate the pressure induced by mineral precipitation and assess which precipitating phase is most capable of driving cracking under the unconfined conditions of our experiments, we made use of a kinetic-fracture mechanics model and an analysis of the stresses generated by the precipitation of different magnesium carbonate phases within our samples. We adopt a two-dimensional wedge geometry (etch-pit plus edge crack) be-

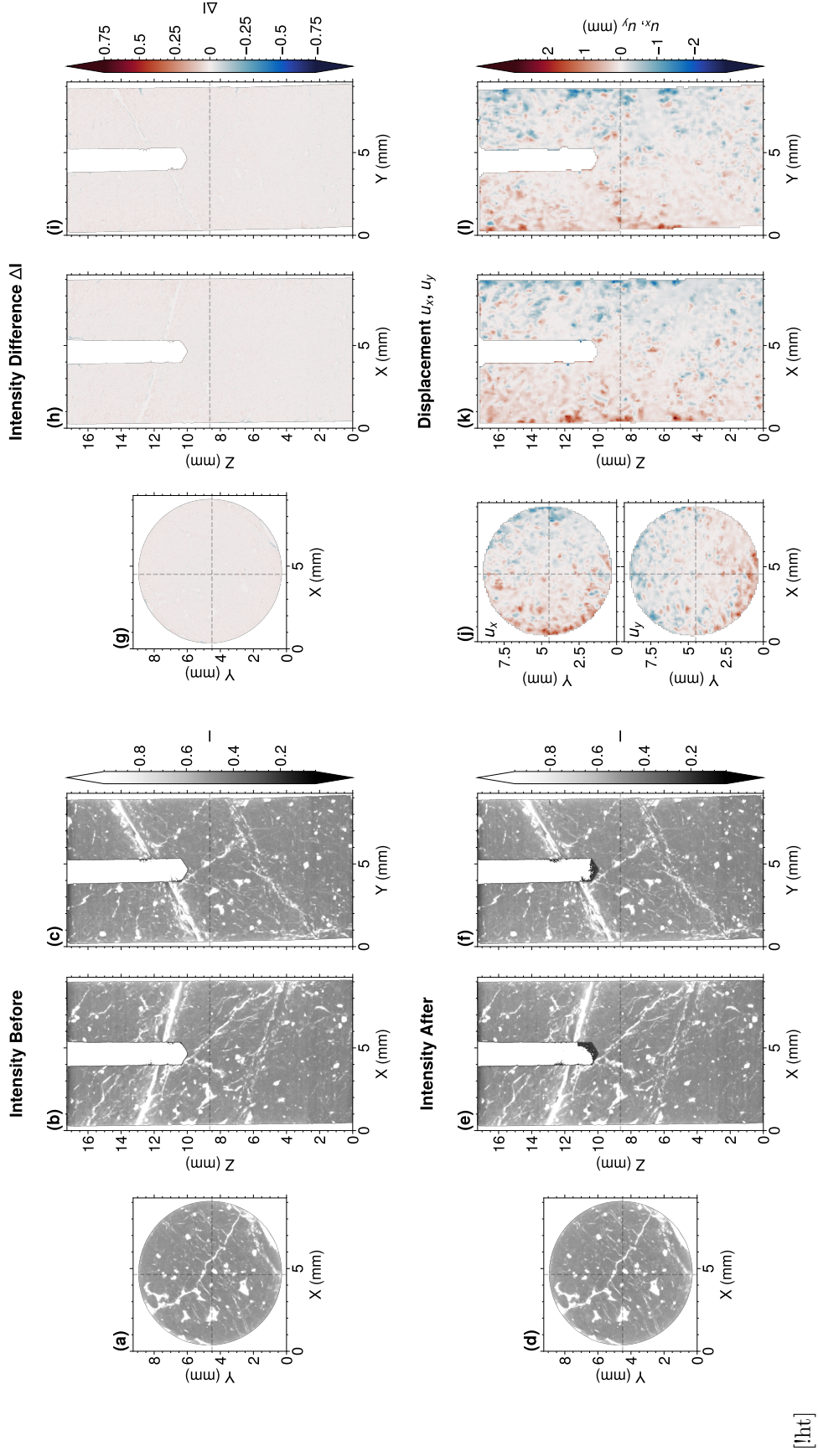
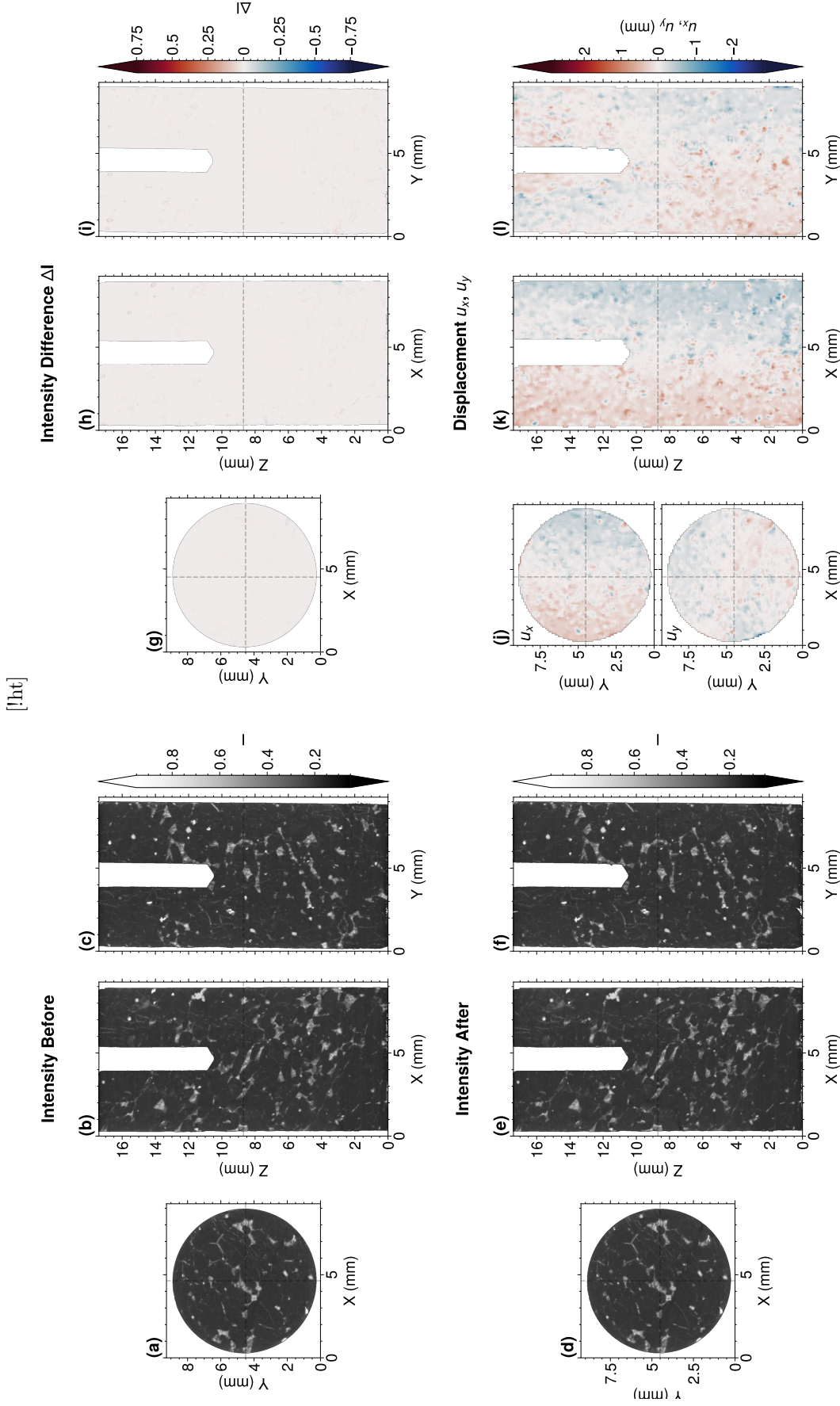


Figure 2. Planar slices from X-ray CT volumes before (a-c) and after (d-f) reaction for Sample A (2 days). (g-i) Slices from intensity difference (ΔI) for sample C. Blue corresponds decreased intensity (lost material) and red corresponds to increased density. (j-l) Slices from displacement fields, $u_x + u_y$, computed from digital volume correlation. There are no noticeable changes of intensity (dissolution, precipitation, or displacement). The displacement field is random and limited to small values.



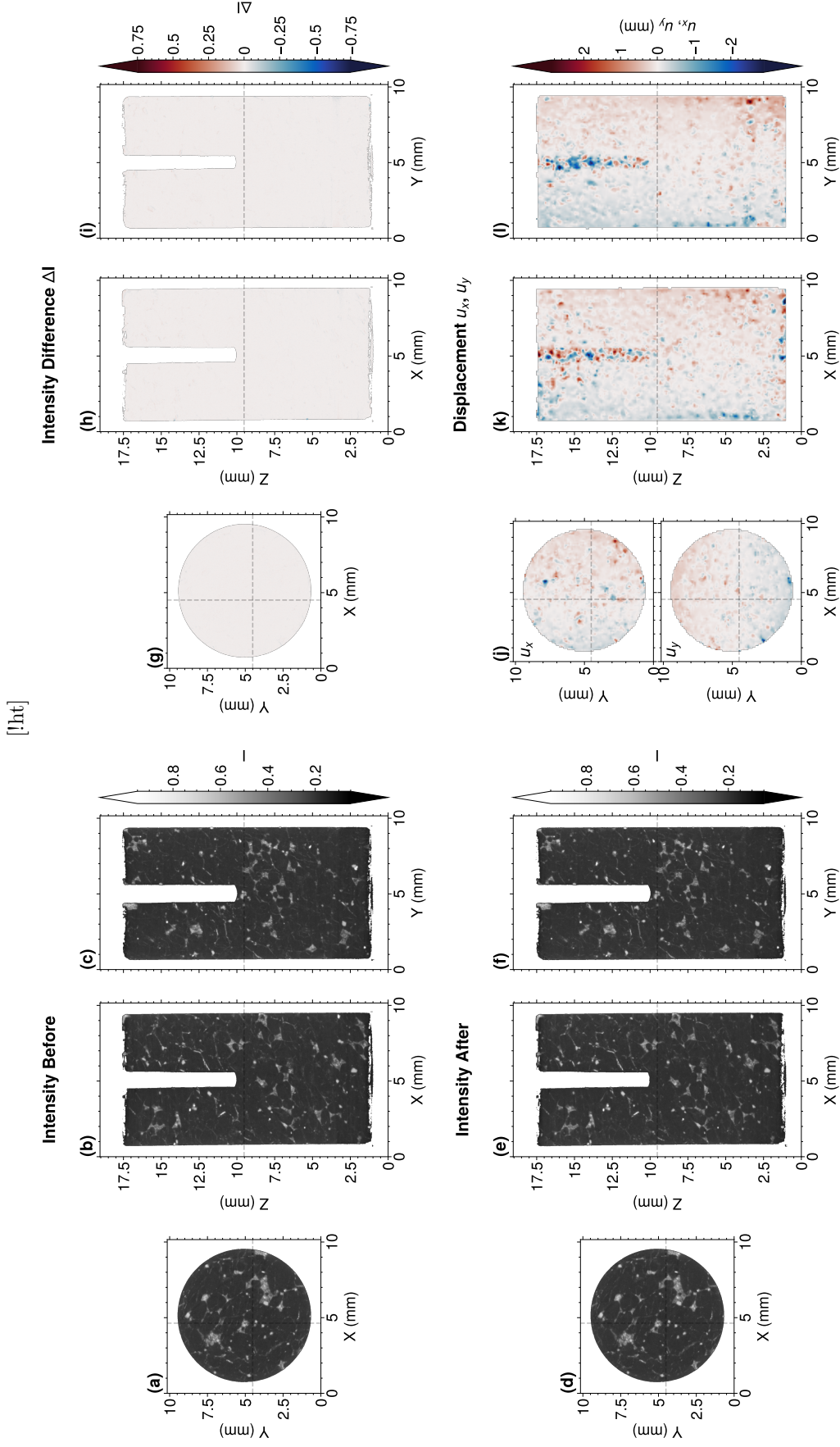


Figure 4. Planar slices from X-ray CT volumes before (a-c) and after (d-f) reaction for sample I (11 days, no added $\text{Mg}(\text{OH})_2$). (g-i) Slices from intensity difference (ΔI) for sample C. Blue corresponds decreased intensity (lost material) and red corresponds to increased density. (j-l) Slices from displacement fields, $u_x + u_y$, computed from digital volume correlation. There are no noticeable changes of intensity (dissolution, precipitation, or displacement). The displacement field is random and limited to small values. Ignore the values at the center borehole.

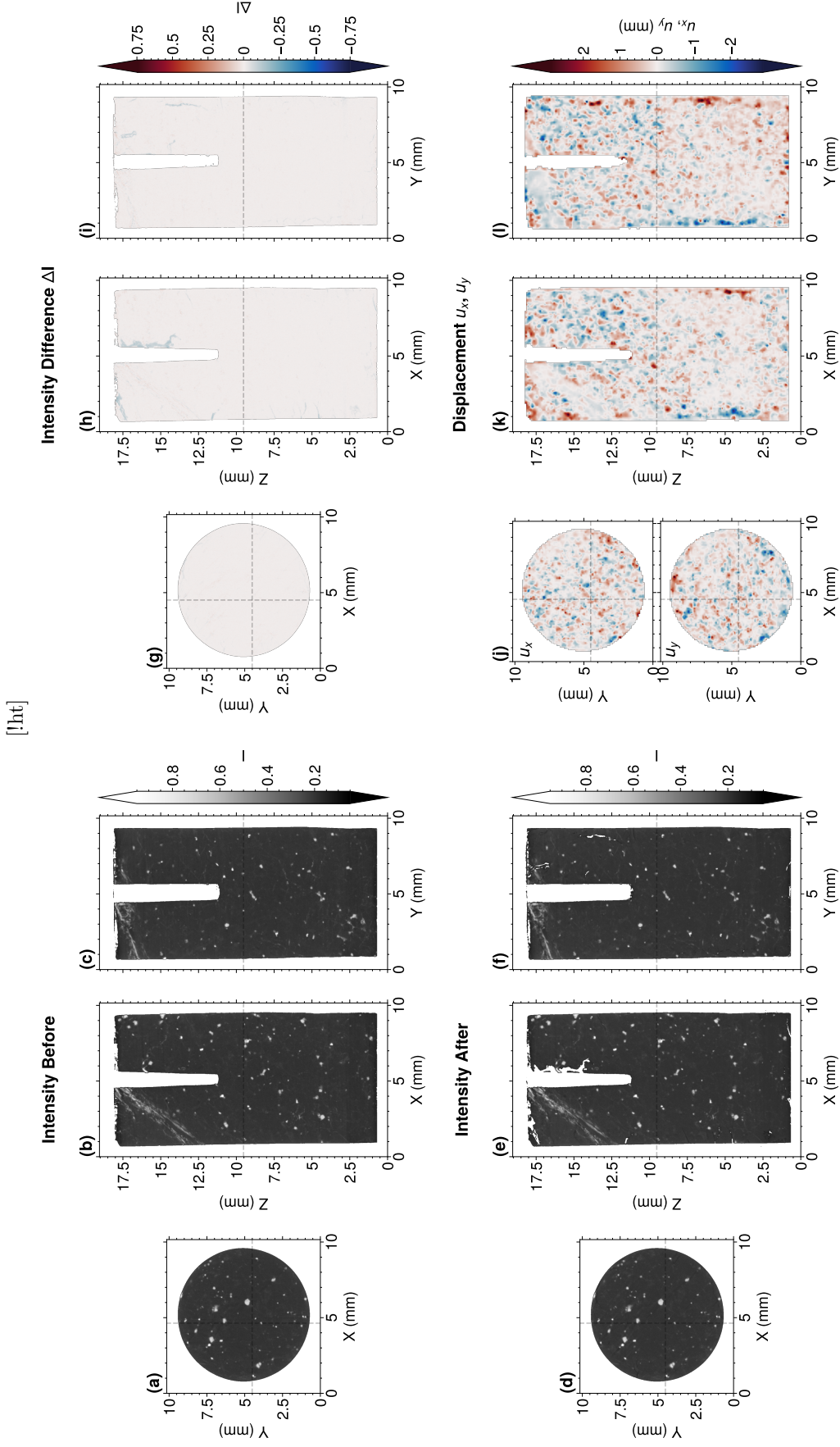


Figure 5. Planar slices from X-ray CT volumes before (a-c) and after (d-f) reaction for sample J (11 days, low temperature). (g-i) Slices from intensity difference (ΔI) for sample C. Blue corresponds decreased intensity (lost material) and red corresponds to increased density. (j-l) Slices from displacement fields, $u_x + u_y$, computed from digital volume correlation. While some features of dissolution are observed near the sample top, the displacement field is random. Displacements are larger than for the case without added $\text{Mg}(\text{OH})_2$.

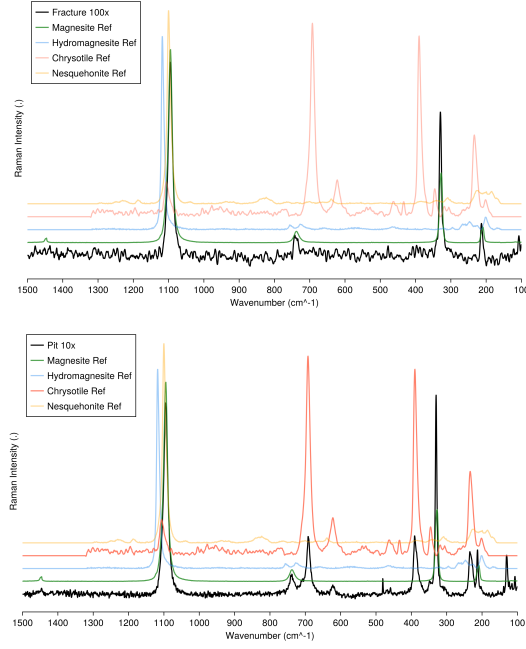


Figure 6. Reference locations for Raman spectroscopy of sample C on a newly formed fracture surface (a) and adjacent to a reactivated vein (b).

cause it (i) upper-bounds the mechanical effect of growth stresses, (ii) approximates the observed reaction front advancing from pit/vein walls in SEM images, and (iii) matches the modeling framework used for reaction-driven cracking in previous work (Malvoisin et al., 2017; Kelemen & Hirth, 2012). In this conceptual model, precipitation into an etch pit creates a wedge size a that presses on the crack faces with length c , promoting propagation from the pit base; the model explicitly balances chemical and mechanical work at the fracture tip (where the distance to the fracture tip is $c - a$), Figure 7a. Following the chemo-mechanical formulation in the literature (Correns, 1949; Steiger, 2005; Kelemen et al., 2011; Malvoisin et al., 2017), the exerted crystallization pressure/stress is approximated by,

$$\sigma_c = \frac{RT}{V_m} \ln \left(\frac{K}{Q} \right), \quad (1)$$

where R is the gas constant, T is the temperature, V_m is the molar volume of the precipitating phase, K is the solubility product, and Q is the reaction quotient. K/Q is known as the supersaturation ratio.

Operationally, the crystallization pressure is controlled by the logarithm of the product of activity ratios of the precipitating ions relative to the host $\sigma_c \propto \ln a_{prec}/a_{host}$ (Correns, 1949; Steiger, 2005). Full formulation for specific species is available in the next subsection. Here, the “serpentine host” refers to the lizardite/chrysotile matrix ($\rho \approx 2.5$ g/cm³, $E \approx 50$ GPa, $\nu = 0.25$) enclosing brucite-rich veins ($\text{Mg}(\text{OH})_2$, $V_m = 24.6$ cm³/mol, $\rho \approx 2.36$ g/cm³), where replacement occurs via the reaction $\text{Mg}(\text{OH})_2 + \text{CO}_2 \rightarrow \text{MgCO}_3 + \text{H}_2\text{O}$ for magnesite, and $\text{Mg}(\text{OH})_2 + \text{CO}_2 + 4\text{H}_2\text{O} \rightarrow \text{MgCO}_3 \cdot 5\text{H}_2\text{O}$ for nesquehonite. To illustrate the importance of precipitating species we estimate the crystallization pressure of (i) magnesite ($V_m = 28.1$ cm³/mol) and (ii) nesquehonite ($V_m = 117.4$ cm³/mol) in a serpentine host. Figure 7b shows σ_c as a function of temperature for magnesite (mgs)

and nesquehonite (nes), where notably, nesquehonite quickly decreases below zero at room temperature (negative Gibbs free energy). An estimate of the minimum pit size c_{min} required to initiate crack growth is given by (Malvoisin et al., 2017),

$$c_{min} = \frac{\pi K_{IC}^2}{4 [\gamma(1) \sigma_c - \mu \sigma'_\infty]^2}. \quad (2)$$

where $\gamma(1)$ is a geometric factor for uniform crack-face traction (Sawaki et al., 1987), $K_{IC} = 1.012 \text{ MPa}\cdot\text{m}^{1/2}$ is mode-I fracture toughness, (measured from semi-circular bending tests on similar serpentinized dunites; see Supplemental Information, Section S5), and σ_c is the crystallization pressure (Eq. 1), with far-field stress $\sigma'_\infty = 0 \text{ MPa}$ due to unconfined conditions and $\mu \approx 1.76$. Figure 7c shows c_{min} as a function of temperature for these parameters. Overall, these approximations highlight that magnesite precipitation in a serpentine host can exert pressures $\sigma_c > 100 \text{ MPa}$ (comparable in order of magnitude to values calculated for serpentine precipitation in olivine $\sim 250 \text{ MPa}$ at $100\text{--}200^\circ\text{C}$ (Kelemen & Hirth, 2012; Malvoisin et al., 2017)) which is sufficient to propagate fractures in micron-scale cracks/veins, thus supporting our reaction-driven fracturing observations from DVC and SEM-EDS. Notice that according to Eq. 2, a confining stress of $\sim 50\%$ of the crystallization pressure is required to balance the stresses that drive fracture propagation. This is about 250 MPa of confining stress for magnesite at 100°C . A mid-shallow confining of 10 MPa reduces the fracture propagation stress (magnitude within square brackets in denominator of Eq. 2) in less than 5% for magnesite at 100° . Hence, a 500 m burial would not stop reaction-driven cracking by magnesite. On the other hand, nesquehonite's larger V_m creates positive pressures at low temperatures but becomes negative above $\sim 35^\circ\text{C}$. Comparatively, the minimum crack size needed to propagate a fracture is much larger than for magnesite, and can be easily inhibited by confining stress.

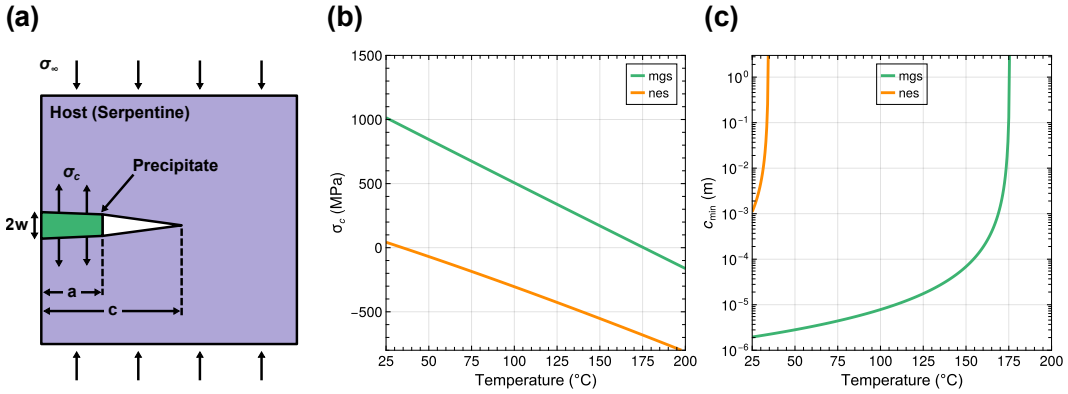


Figure 7. (a) Conceptual plane-strain model of fracture propagation facilitated from carbonate precipitation inside a pre-existing crack (natural or via dissolution) in serpentine rock, model after (Malvoisin et al., 2017). (b) Crystallization pressure, σ_c , as a function of temperature for a fluid pressure of 3.5 MPa , as computed from Eq. 1 for magnesite (mgs) and nesquehonite (nes) precipitating in a cracked serpentine host. (c) Minimum crack size c_{min} whose length is sufficient for fracture propagation, i.e. fracture intensity surpasses toughness, calculated from Eq. 2. At 90°C , fractures as small as $2c \approx 10 \mu\text{m}$ can propagate when magnesite precipitates in unconfined serpentine rock.

S6.1 Estimation of Crystallization Pressure Induced by Precipitation

We estimate the maximum normal stress that confined carbonate growth can exert on fracture walls (the crystallization pressure, σ_c) by linking geochemical saturation state to mechanical work. The approach follows classical treatments of reaction affinity and crystallization pressure: precipitation is driven by the chemical potential drop associated with supersaturation, but growth in a confined geometry must perform mechanical work against a compressive normal stress (Correns, 1949; Scherer, 2004; Steiger, 2005). In practice, we compute σ_c from the equilibrium constant K and the activity product Q using standard states and non-ideal activity models for aqueous ions and CO_2 (Anderson & Crerar, 1993). The procedure is similar to the one shown by Kelemen and Hirth (2012).

Reaction-driving force and supersaturation

We write reactions using stoichiometric coefficients ν_i that are positive for products and negative for reactants. For each species i the chemical potential is expressed as

$$\mu_i(T, P, \mathbf{a}) = \mu_i^\circ(T, P) + RT \ln a_i, \quad (3)$$

where μ_i° is the standard-state chemical potential and a_i is activity (Anderson & Crerar, 1993). Summing over all species gives the reaction chemical potential (reaction Gibbs free energy per mole of reaction),

$$\Delta\mu_r(T, P, \mathbf{a}) \equiv \sum_i \nu_i \mu_i = \Delta\mu_r^\circ(T, P) + RT \ln Q, \quad (4)$$

where $\Delta\mu_r^\circ = \sum_i \nu_i \mu_i^\circ$ and

$$\ln Q \equiv \sum_i \nu_i \ln a_i. \quad (5)$$

The equilibrium constant is defined by

$$\ln K(T, P) \equiv -\frac{\Delta\mu_r^\circ(T, P)}{RT}. \quad (6)$$

Combining these relations yields an expression for the reaction driving force:

$$\Delta\mu_r(T, P, \mathbf{a}) = RT \ln \left(\frac{Q}{K} \right). \quad (7)$$

Equation 7 is the bridge between geochemical saturation state and mechanical loading. For precipitation reactions written with the solid as a product, pure solids have $a_s \equiv 1$ and therefore do not appear in Q (while still contributing to K through $\Delta\mu_r^\circ$) (Anderson & Crerar, 1993). In this precipitation convention, supersaturation corresponds to $Q < K$, which implies $\Delta\mu_r < 0$ and favors precipitation.

S6.1.1 Connecting saturation state to crystallization pressure

We treat confined growth as doing work against a crack-normal compressive stress σ_c (positive in compression). In classical crystallization-pressure arguments, the stress required to halt net precipitation is found by balancing the chemical potential driving force with a mechanical work term proportional to the volume of solid formed (Correns,

1949; Weyl, 1959; Steiger, 2005). For a reaction producing ν_s moles of solid s per mole of reaction, we impose the metastable equilibrium condition stress scale, we treat

$$\Delta\mu_r(T, P, \mathbf{a}) + \nu_s V_{m,s}(T, P) \sigma_c = 0, \quad (8)$$

where $V_{m,s}$ is the molar volume of the precipitate. Substituting Eq. 7 yields

$$\sigma_c(T, P, \mathbf{a}) = \frac{RT}{\nu_s V_{m,s} \ln \left(\frac{K}{Q} \right)} \quad (9)$$

Eq. 9 satisfies the expected limits: (i) at equilibrium ($Q = K$), $\sigma_c = 0$; (ii) for supersaturation ($Q < K$), $\sigma_c > 0$ (compressive), consistent with the standard interpretation of crystallization pressure (Correns, 1949; Steiger, 2005).

Equation 9 defines a thermodynamic upper bound on stress generation by confined growth. Whether this bound is attained depends on transport, nucleation kinetics, evolving pore geometry, and mechanical compliance of the host (Steiger, 2005; Espinosa-Marzal & Scherer, 2010).

Species activities

Computing Q requires activities for dissolved and gaseous species appearing in the reaction stoichiometry. We use a standard-state framework common in aqueous geochemistry, in which $\mu_i^\circ(T, P)$ defines the reference thermodynamics and a_i accounts for composition and non-ideality (Anderson & Crerar, 1993; Bethke, 2007). A key bookkeeping rule is that pressure/non-ideality corrections are included *either* in μ_i° *or* in a_i , but not both (Bethke, 2007).

S6.1.1.1 Solids and pure liquids Pure solids and pure liquids have unit activity ($a \equiv 1$) and do not appear in Q , but they do contribute to K through $\Delta\mu_r^\circ$ (Anderson & Crerar, 1993).

S6.1.1.2 Aqueous solutes For aqueous ions and complexes,

$$a_i = \gamma_i \left(\frac{m_i}{m^\circ} \right), \quad (10)$$

where m_i is molality, $m^\circ = 1 \text{ mol kg}^{-1}$, and γ_i is the activity coefficient (Stumm & Morgan, 2013; Bethke, 2007). Ionic strength is

$$I = \frac{1}{2} \sum_i m_i z_i^2. \quad (11)$$

We model activity coefficients using an extended Debye–Hückel (B-dot) form,

$$\log_{10} \gamma_i = -\frac{A_\gamma(T, P) z_i^2 \sqrt{I}}{1 + B_\gamma(T, P) a_i \sqrt{I}} + b_\gamma(T, P) I, \quad (12)$$

which is widely used for dilute to moderately saline solutions (Debye & Hückel, 1923; Helgeson, 1969; Truesdell & Jones, 1974; Bethke, 2007).

S6.1.1.3 Gas species (CO₂) For gases, we write activity in terms of fugacity,

$$a_g = \frac{f_g}{P^\circ} = \frac{\phi_g y_g P}{P^\circ}, \quad (13)$$

where ϕ_g is the fugacity coefficient, y_g is mole fraction, and P is total pressure (Stumm & Morgan, 2013). We compute ϕ_g using the Peng–Robinson equation of state (Peng & Robinson, 1976). With this convention, pressure and nonideality of the gas phase are carried by a_g , so we do not separately apply an additional pressure integration for gas standard chemical potentials beyond P° (Bethke, 2007).

Computation of K

The equilibrium constant is determined from the standard-state reaction chemical potential:

$$\Delta\mu_r^\circ(T, P) = \sum_i \nu_i \mu_i^\circ(T, P), \quad \ln K(T, P) = -\frac{\Delta\mu_r^\circ(T, P)}{RT}. \quad (14)$$

Standard-state chemical potentials are evaluated using established thermodynamic models and databases. For aqueous solutes we use a revised Helgeson–Kirkham–Flowers (HKF) formulation (or equivalent SUPCRT-type standard state) to capture the T, P dependence through water density and dielectric properties (Helgeson, 1981; Johnson et al., 1992). Solid phases are treated using tabulated thermodynamic data with appropriate compressibility/EOS corrections over the experimental T, P range (Holland & Powell, 1998). Gas species are referenced to P° with fugacity corrections handled via a_g (Section S6.1.1) (Peng & Robinson, 1976).

Carbonate speciation and calculation of $a_{\text{CO}_3^{2-}}$

When carbonate activity is not directly prescribed, we compute the distribution of dissolved inorganic carbon (DIC) among CO_2^* , HCO_3^- , and CO_3^{2-} using the dissociation equilibria

$$K_1 = \frac{a_{\text{H}^+} a_{\text{HCO}_3^-}}{a_{\text{CO}_2^*}}, \quad K_2 = \frac{a_{\text{H}^+} a_{\text{CO}_3^{2-}}}{a_{\text{HCO}_3^-}}, \quad (15)$$

where CO_2^* denotes the conventional lumping of $\text{CO}_2(\text{aq})$ and H_2CO_3 (Stumm & Morgan, 2013). Defining

$$D = 1 + \frac{K_1}{a_{\text{H}^+}} + \frac{K_1 K_2}{a_{\text{H}^+}^2}, \quad (16)$$

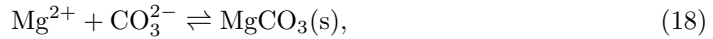
the activity fractions are

$$\alpha_0 = \frac{1}{D}, \quad \alpha_1 = \frac{(K_1/a_{\text{H}^+})}{D}, \quad \alpha_2 = \frac{(K_1 K_2/a_{\text{H}^+}^2)}{D}, \quad (17)$$

corresponding to CO_2^* , HCO_3^- , and CO_3^{2-} , respectively (Stumm & Morgan, 2013). These relations are combined with mass balance (DIC) and charge balance constraints to solve for a_{H^+} (pH) and carbonate species activities (Bethke, 2007).

Worked examples for Q (precipitation convention)

S6.1.1.4 Magnesite precipitation For



the activity product is

$$Q = \frac{1}{a_{\text{Mg}^{2+}} a_{\text{CO}_3^{2-}}}, \quad \ln Q = -\ln a_{\text{Mg}^{2+}} - \ln a_{\text{CO}_3^{2-}}. \quad (19)$$

This is equivalent to the usual ion-activity product (IAP) form for carbonate minerals, expressed consistently in activities (Stumm & Morgan, 2013; Bethke, 2007).

S6.1.1.5 Gas CO_2 contribution If a reaction includes $\text{CO}_2(\text{g})$, its contribution to $\ln Q$ is $\nu_{\text{CO}_2} \ln a_{\text{CO}_2}$ with

$$a_{\text{CO}_2} = \frac{\phi_{\text{CO}_2} y_{\text{CO}_2} P}{P^\circ}, \quad (20)$$

where ϕ_{CO_2} is computed from a cubic EOS (here PR) (Peng & Robinson, 1976).

S6.2 Fracture Toughness

We used the semicircular bending method to measure fracture toughness (Lee et al., 2015). The fracture toughness is obtained from the following equation:

$$K_{IC} = \frac{P_{\max} \sqrt{\pi a}}{2rt} Y_1 \text{ and} \quad (21)$$

$$Y_1 = 5.6 - 22.2 \left(\frac{a}{r}\right) + 166.9 \left(\frac{a}{r}\right)^2 - 576.2 \left(\frac{a}{r}\right)^3 + 928.8 \left(\frac{a}{r}\right)^4 - 505.9 \left(\frac{a}{r}\right)^5, \quad (22)$$

where $P_{\max}$ is the peak load, a is the notch length, r is the sample radius, t is the sample thickness, and Y_1 is a dimensionless coefficient (Basham, 1989; Lee et al., 2015). Figure 8 shows the results.

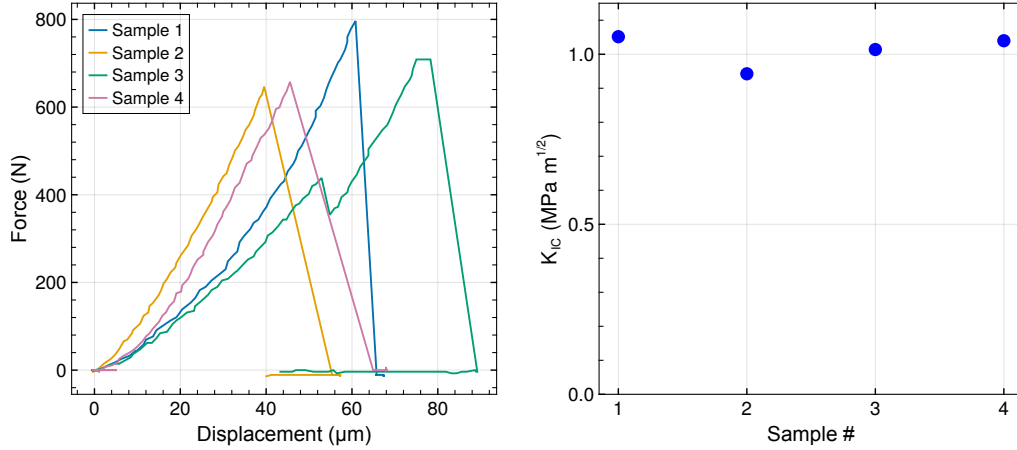


Figure 8. Semi-circular bending test for 4 samples taken from the same larger core used in this study. (a) Force as a function of displacement for each sample. (b) Fracture toughness, calculated from Equation 21 for each sample.

S7 Broader Impacts and Insights

Basalt (mafic) and peridotite (ultramafic) are promising targets for in-situ carbon mineralization because they can permanently and securely sequester large quantities of CO_2 underground (Kelemen et al., 2018; Nisbet et al., 2024). Basaltic lava flows and flood basalts, which have already been tested at pilot scale (e.g., Carbfix and Wallula projects), possess interlayer zones and fracture networks that can facilitate CO_2 injection and fluid circulation (Raza et al., 2022). However, much of the basalt volume, particularly within flow interiors, exhibits injectivity challenges (Aubele et al., 1988). Peridotites contain higher concentrations of divalent cations such as Mg^{2+} , Ca^{2+} and Fe^{2+} , making them more susceptible to carbonation, yet they are less widespread and commonly exhibit even lower permeability than basalts (Kelemen et al., 2020; Raza et al., 2022).

In both lithologies, efficient mineralization of the rock mass depends on the presence or generation of interconnected fracture networks that enable the transport and reaction with CO₂-bearing fluids. Achieving this goal requires understanding the coupled thermal, hydrological, mechanical, and chemical processes that control long-term carbonation. These processes involve complex feedbacks among fracture propagation, fluid flow, mineral dissolution, precipitation, and fracture sealing, including passivation of reactive mineral surfaces (Béarat et al., 2006), carbonate precipitation that clogs pores and fractures (Jöns et al., 2017), subcritical crack growth (Atkinson, 1984), and reaction-driven cracking (Kelemen & Hirth, 2012).

Among these mechanisms, reaction-driven cracking is likely critical for sustaining permeability and allowing carbonation reactions to progress through otherwise low-permeability rock masses. Our experiments show that reaction-driven cracking is particularly effective when driven by magnesite precipitation rather than by softer hydrous magnesium carbonates. We observed extensive precipitation-induced cracking and progressive disaggregation of the samples, with some cores (samples D and E) ultimately breaking apart completely. Importantly, these reactions occurred at temperatures below 100° C, conditions representative of depths of approximately 2–4 km in the crust, which are readily accessible by drilling for engineered CO₂ injection and carbon mineralization. The experimental conditions also capture an important aspect of the spatial evolution of the reaction around a CO₂ injection wellbore. Near the wellbore, CO₂-charged water is expected to dissolve brucite, progressively enriching the fluid in dissolved Mg²⁺ as it moves through the rock mass. Farther from the injection point, therefore, the reacting fluid may be substantially more Mg-rich than the injected fluid. Our Mg(OH)₂-saturated batch experiments approximate these far-field aqueous conditions, where brucite dissolution has already supplied Mg²⁺ to the fluid, and demonstrate that carbonate precipitation and associated reaction-driven cracking can remain effective under these evolved fluid compositions. This suggests that precipitation-driven fracturing may extend beyond the immediate vicinity of the injection wellbore and develop progressively along the fluid flow path.

The fractured and mechanically weakened rock produced during our experiments has important implications for mining and enhanced mineral extraction, particularly in ultramafic systems. Ultramafic rocks commonly host nickel sulfides and other critical minerals that require energy-intensive crushing, grinding, and flotation to separate valuable phases from the silicate gangue. Our results suggest that carbonation-driven alteration could act as a form of reactive geomechanical pre-conditioning, in which injection of CO₂-rich fluids before mining induces carbonate precipitation, reaction-driven cracking, and progressive rock weakening. Because carbonate minerals are less dense and mechanically weaker than the primary mafic and ultramafic silicates they replace, carbonation promotes volumetric expansion, fracture propagation, and partial disaggregation of the rock mass. As shown in our experiments, this process not only increases fracture pore space and reactive surface area, but may also reduce the mechanical energy required for blasting, comminution, and beneficiation, while permanently storing CO₂ within the waste rock. In both open-pit and underground operations, such pre-conditioning could lower operational costs and reduce the environmental footprint of mining by decreasing energy consumption during ore processing (Ukar, 2026).

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