The rapid resetting of the Ca isotopic signatures of calcite at ambient temperature during its congruent dissolution, precipitation, and at equilibrium

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Abstract- This study provides direct experimental evidence of the resetting of the calcium (Ca) isotope signatures of calcite in the presence of an aqueous fluid during its congruent dissolution, precipitation, and at equilibrium at ambient temperatures over week-long timescales. Batch reactor experiments were performed at 25 ºC in aqueous NaCl solutions; air or CO_2-gas mixtures were bubbled through this fluid to fix pH. During congruent calcite dissolution, the fluid became enriched in isotopically heavy Ca, and the Ca isotope composition continued to become heavier after the fluid attained bulk chemical equilibrium with the mineral; the $\delta^{44/42}$Ca composition of the fluid was up to 0.8 ‰ higher than the dissolving calcite at the end of the dissolution experiments. Calcite precipitation was provoked by increasing the reactor fluid pH after chemical equilibrium had been attained via dissolution. Rayleigh isotope fractionation effects were observed immediately after the pH was increased and rapid calcite precipitation occurred. However, isotopic exchange continued after the system chemically equilibrated, eradicating this Rayleigh signal. Taken together, these observations 1) confirm dynamic mineral-fluid equilibrium (i.e. dissolution and precipitation occur at equal, non-zero rates at equilibrium), and 2) indicate that isotopic compositions of calcite can readily equilibrate even when this mineral is in bulk chemical equilibrium with its coexisting fluid. This latter observation suggests the preservation of paleo-environmental isotopic signatures in calcite may require a combination of the isolation

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of fluid-mineral system from external chemical input and/or the existence of a yet to be defined calcite dissolution/precipitation inhibition mechanism.

1. INTRODUCTION

The Ca isotope compositions of natural calcium carbonates are widely used to illuminate a large number of natural processes including the global Ca cycle (Zhu and Macdougall, 1998; Heuser *et al.,* 2005; Tipper *et al.,* 2006; Fantle and Tipper, 2014; Sawaki *et al.,* 2014; Farkas et al., 2016; Silva-Tamayo *et al.,* 2018), continental weathering rates (Tipper *et al.,* 2008; Blattler *et al.,* 2011; Hindshaw *et al.,* 2011; 2013; Kasemann *et al.,* 2014), soil formation (Page *et al.,* 2008; Cenki-Tok *et al.,* 2009; Holmden and Belanger; 2010), past and present environmental conditions (Kasemann *et al.,* 2005; Owen *et al.,* 2016), and the mechanism of biomineral formation (Pruss e*t al.,* 2018). Moreover, the Ca isotope signatures of natural waters are being used to trace their origin (Druhan *et al.,* 2013; Yan *et al.,* 2016; Lyons *et al.,* 2017; Li *et al.,* 2018). These potential applications have motivated an increasing number of studies focused on the measurement and behavior of calcium isotopic compositions in low-temperature systems (e.g. DePaolo, 2004; Tang e*t al.,* 2008, 2012; Fantle and Tipper, 2014; AlKhatib and Eisenhauer, 2017; Wang *et al.,* 2017). A substantial number of experimental studies have concluded that isotopically light Ca is preferentially incorporated into calcium carbonate minerals (Lemarchand *et al.,* 2004; Marriott *et al.,* 2004; Gussone *et al.,* 2005, 2011; Reynard *et al.,* 2011; Tang *et al.,* 2008, 2012).

The interpretation of stable isotope compositions of natural minerals and waters commonly relies on two assumptions. The first is that once a mineral is precipitated it retains its original isotopic signal and thus preserves information about its formation conditions. This assumption serves as the basis for paleo-environmental and paleo temperature

reconstructions (e.g. Marshall, 1992; Koch, 1998; Nagler *et al.,* 2000; Leng and Marshall, 2004; Fairchild et al., 2006). The second assumption is that stable isotope signatures are transferred conservatively to the fluid phase during congruent mineral dissolution[1] (e.g. Jacobson and Holmden, 2008; Ryu *et al.,* 2011; Turchyn and DePaolo, 2011). This latter assumption has been adopted to trace flow paths and the origin of fluids in natural systems (e.g. Grazis and Feng, 2004; Wiederhold, 2015).

A number of studies, however, have presented evidence that the isotopic compositions of carbonate minerals can be altered while dissolving stoichiometrically or when the mineral is in bulk chemical equilibrium with the fluid phase over the course of hours to days. For example, Pearce e*t al.* (2012) observed that Mg isotopes fractionated substantially during stoichiometric magnesite dissolution at 150 and 200 °C, and that the fluid Mg isotopic composition continued to evolve after bulk equilibrium was attained between the mineral and fluid. Mavromatis *et al.* (2016) reported that the Ba isotope signature of precipitated witherite continued to evolve in closed system experiments after the attainment of bulk mineral-aqueous fluid equilibrium. Mavromatis *et al*. (2017a) observed the continued Sr isotope exchange in strontianite in closed system reactors after the attainment of bulk mineral-aqueous fluid equilibrium in both dissolution and precipitation experiments. Similar observations have been made for Mg isotope exchange in hydromagnesite (Oelkers *et al.,* 2018), Mg and Ca isotope exchange in dolomite (Perez-Fernandez *et al.,* 2017), Mg isotope exchange in amorphous calcium carbonate (ACC) and Mg-rich calcite (Mavromatis et al., 2017b), Ba isotope exchange in barite (Curti *et al.,* 2010), and Si isotope exchange in quartz (Liu et al., 2016). Mavromatis *et al.* (2012) and Shirokova *et al.* (2013) observed the

[1] The term congruent dissolution in this study refers to stoichiometric dissolution without the formation of secondary minerals nor evidence for re-crystallisation.

continuous re-equilibration of Mg isotopes between the hydrous Mg carbonate minerals hydromagnesite/dypingite and its co-existing aqueous fluid at ambient temperatures during experiments that lasted no more than 4 weeks. In a follow-up study, Mavromatis *et al.* (2015) reported the continuous re-equilibration of C isotopes between the hydromagnesite/dypingite and its co-existing aqueous fluid at the same conditions. The resetting of mineral isotopic signatures in natural sedimentary rocks has been referred to as 'diagenetic effects'; a number of studies have concluded that these effects influence the calcium isotopic signatures of natural calcium carbonate minerals (Fantle and DePaolo, 2007; Teichert *et al.,* 2009; Fantle *et al.,* 2010; Druhan *et al.,* 2013; Fantle and Higgins, 2004; Harouaka et al., 2014; Jost *et al.,* 2014; Steefel *et al.,* 2014; Fantle, 2015; Chanda et al., 2019).

Isotope exchange towards mineral-fluid *isotopic equilibrium* during congruent dissolution and at *bulk chemical equilibrium* is consistent with the concept of dynamic equilibrium (van't Hoff, 1884). This concept is often invoked for the interpretation of mineral dissolution and precipitation rates (e.g. Aagaard and Helgeson, 1982; Oelkers *et al.,* 1994; Schott and Oelkers, 1995; Oelkers, 2001; Schott et al., 2009, 2012). Critical to the concept of dynamic equilibrium as applied to mineral-fluid interaction is that chemical reactions proceed by two-way mass transfer, the combination of forward dissolution and reverse precipitation. Note that the reverse precipitation is of the identical material that is removed from the mineral during forward dissolution. In undersaturated fluids, the rate of the forward dissolution is faster than that of the reverse precipitation and the net overall process is mineral dissolution, whereas in supersaturated solutions the rate of the reverse precipitation is faster than that of forward dissolution and the overall net reaction is mineral precipitation. At equilibrium, the rate of forward dissolution is equal to that of the reverse precipitation

such that there is no net bulk chemical transfer. As such, even if the original removal of an element from a mineral via forward dissolution is isotopically conservative[2], isotopic fractionation can occur as a consequence of the coupled reverse precipitation, both during bulk dissolution and at equilibrium (see Steefel et al., 2014). To assess the degree to which such processes can effect the isotopic compositions of calcite, this study followed the temporal evolution of calcium isotope compositions of the fluid phase during the congruent dissolution, precipitation, and at chemical equilibrium during ambient temperature closed-system calcite-water experiments. Results of shorter than month-long experiments demonstrate that 1) calcium isotopes are not conservatively transferred the fluid during congruent calcite dissolution, 2) the calcium isotopic composition of calcite and its co-existing fluid phase continue to evolve after the fluid has attained bulk chemical equilibrium, and 3) kinetically mediated isotopic signatures obtained during rapid calcite precipitation can be subsequently re-equilibrated by further fluid-calcite interaction. These observations require that a substantial portion of the calcium within the calcite is transferred in and out of the fluid phase in experiments performed at conditions at or near to calcite/aqueous fluid bulk chemical equilibrium over the course of only a few weeks. The purpose of this article is to report the results of this experimental study and use these to assess the potential for the preservation of the original calcium isotope compositions of calcite in natural systems.

2. COMPUTATIONAL, EXPERIMENTAL, AND ANALYTICAL METHODS

The standard state adopted in this study for thermodynamic calculations is that of unit activity for pure minerals and H_2O at any temperature and pressure. For aqueous

[2] The term isotopically conservative in this manuscript refers to the transfer of a metal to an aqueous solution without zero isotopic fractionation.

species other than H_2O, the standard state is unit activity of the species in a hypothetical 1 molal solution referenced to infinite dilution at any temperature and pressure. Calcite dissolution can be described using:

$$\text{Calcite} = Ca^{2+} + CO_3^{2-} \quad (1)$$

Taking account of the standard state, the law of mass action for this reaction is given by:

$$K_{calcite} = a_{Ca^{2+}} a_{CO_3^{2-}} \quad (2)$$

where $K_{calcite}$ stands for the equilibrium constant of reaction (1), and a_i represents the activity of the subscripted aqueous species. The saturation state of an aqueous fluid with respect to calcite can be quantified using the saturation index (*SI*) defined by

$$SI = log\left(\frac{IAP}{K_{calcote}}\right)$$

where IAP signifies for the aqueous ion activity product for reaction (1). Note that *SI* is negative when the fluid is undersaturated with respect to calcite, positive when supersaturated, and zero at fluid-calcite equilibrium. All thermodynamic calculations in this study were performed using the PHREEQC computer code, together with its minteq.v4 database (Parkhurst and Appelo, 1999). Note that consideration of the analytical uncertainties described below and those associated with the equilibrium constants present in the minteq.v4 database, uncertainties associated with the calcite *SI* values calculated in this study are on the order of ±0.1 (c.f. Voigt et al., 2018).

Two calcite-aqueous fluid batch reactor experiments were performed in 1000 ml polypropylene reactors that were placed in a thermostated bath operating at 25 ^{o}C. Batch reactor systems were chosen for this study to allow the fluid-water system to attain close to bulk chemical equilibrium conditions, such that the isotopic evolution of the calcite and its

coexisting fluid phase could be observed in the absence of net chemical transfer between the mineral and the fluid. Each batch reactor was equipped with a floating stirring bar that rotated at ~250rpm. Experiment A was initiated by placing 2.35 g of calcite seed crystals together with 965.4 g of an aqueous 0.01 mol/kg NaCl solution into the reactor. Air was bubbled continuously through the reactor to fix pH. After 25 hours, the air was replaced by pure CO_2 gas, lowering the fluid pH to 6.2 and provoking calcite dissolution; this part of experiment A is referred to as the 'dissolution leg'. After an additional 193 hours, the pure CO_2 gas was replaced with a 1% CO_2/N_2 gas mixture, raising the pH to 7.5 and provoking calcite precipitation; this part of experiment A will be referred to as the 'precipitation leg'. Experiment B was initiated by placing 2.03 g of calcite crystals, together with 1001.1 g of a 0.008 mol/kg NaCl solution into a reactor. Pure CO_2 gas was bubbled in the reactor fluid, leading to a fluid pH of 6.3 and provoking calcite dissolution. Fluid samples were collected regularly from each reactor using a 0.45 μm cellulose nitrate syringe filter. Prior to sampling, the stirring bar was stopped to allow the suspended calcite to settle, minimizing changes in the mass of solid present in the reactor due to sampling; no evidence of calcite passing through the filter syringes during the sampling was observed. Note also that prior to their introduction in the reactors, all gasses were saturated with a 0.01 mol/kg NaCl solution in a separate fluid cell to avoid fluid evaporation in the reactors.

The solids used in these experiments were pure synthetic Merck reagent grade calcite. The calcite used in the two experiments was taken from different containers of Merck calcite, so differed slightly in their initial Ca isotope compositions (see below). Pure synthetic calcite was chosen for these experiments to avoid any potential artifacts (such as non-congruent dissolution) arising from the presence of impurities in the solids. Note that, for example, the presence of Mg in biogenic calcite can provoke the dissolution of the

original solid and the precipitation of a more stable Mg-free calcite during isotope exchange experiments (Chanda et al., 2019). The solids in our study were not cleaned prior to use to avoid altering the surfaces prior to the experiments. The synthetic calcite consisted of 4 to 15 µm rhombohedral shaped crystals as shown in Fig. 1. X-ray Diffraction (XRD) and Energy Dispersive X-ray Spectroscopic (EDS) analyses confirmed that these crystals were pure calcite. Potential isotopic heterogeneities in these solids were not investigated. The specific surface area of the original calcite, as determined by multi-point krypton adsorption according to the BET method (Brunauer *et al.,* 1938) using a Quantachrome Instruments Autosorb 1, was 0.25 ± 10% m^2/g.

Aqueous fluids were regularly sampled from the reactor and analyzed for total alkalinity, pH, and calcium concentration. The alkalinity of each fluid sample was obtained by HCl titration using an automatic Schott TitroLine alpha TA10^{plus} titrator with an uncertainty of ±2% and a detection limit of 5×10^{-5} eq kg^{-1}. Fluid phase pH measurements were performed at 25 °C immediately after sampling using a standard glass electrode, previously calibrated with 4.01, 6.86, and 9.18 NIST pH buffers; the precision of these measurements is ±0.02 pH units. The Ca concentration of each sample was measured by flame Atomic Absorption Spectroscopy (AAS) with an analytical uncertainty of ~3% and a detection limit of 6×10^{-7} M. The estimated uncertainties in these analyses are based on repeat analyses of selected samples performed regularly during the analyses.

The calcium isotope compositions of the fluid samples, as well as the bulk solids before and after the experiments were determined. Calcium isotope compositions are reported as $\delta^{44/42}$Ca normalized to the NIST Ca standard SRM915a (Schmidt *et al.,* 2001) consistent with

192 $$\delta^{44/42}Ca = \{[(^{44}Ca/^{42}Ca)_{sample} - (^{44}Ca/^{42}Ca)_{standard})]/(^{44}Ca/^{42}Ca)_{standard}\} \times 1000 \ ,$$

193 where $(^{44}Ca/^{42}Ca)_{sample}$ refers to the indicated isotopic molar ratio of the subscripted phase.
194 Calcium was purified for isotopic analysis by ion exchange chromatography. Prior to
195 separation, fluid samples were evaporated to dryness in a Savillex beaker, redissolved in
196 concentrated aqueous HNO_3, evaporated to dryness again and dissolved in aqueous 2M HCl
197 ready for loading on columns. For the solid calcite samples, ~10 mg of the solid was
198 dissolved in concentrated aqueous HCl before being evaporated to dryness, then dissolved
199 in concentrated aqueous HNO_3, evaporated to dryness again then dissolved in aqueous 2M
200 HCl. An aliquot of this fluid was subsequently loaded onto the ion exchange resin, with 20–
201 30 µg of Ca processed for each sample. Purification chemistry was performed in the isotope
202 laboratories in the Earth Science Department at Oxford University. The method was
203 previously described (Chu *et al.,* 2006; Reynard *et al.,* 2010; Blattler *et al.,* 2011), but, briefly,
204 samples were purified through a 2-step ion-exchange method to isolate Ca^{2+}, where the first
205 column used AG50W X12 (200-400 mesh) cation exchange resin, and the second column a Sr
206 spec resin, to separate the isobaric interferences from strontium. Total Ca yields were
207 greater than 99%, as determined by Ca content analysis of splits collected before and after
208 the main collection bracket. The total procedural blank for Ca isotope analysis is ~0.5–0.7 ng,
209 which is insignificant compared to the mass of sample used. Purified fluids were diluted to a
210 concentration of 10 µg/ml, and measured on a Nu Instruments multi-collector inductively
211 coupled plasma mass spectrometer (MC-ICP-MS), using a sample-standard bracketing
212 system relative to SRM-915a. With an uptake rate of 100 µl/min, and using a Nu Instruments
213 DSN desolvating nebulizer, a sensitivity of ~40 pA on ^{44}Ca was achieved. Mass 43.5 was
214 continuously monitored to assess potential doubly charged Sr; the contribution of Sr to the
215 Ca isotope ratio was consistently less than 0.01‰. Instrument precision was assessed by

running in-house Ca standards, and accuracy and external precision was assessed by repeated analyses of seawater ($\delta^{44/42}$Ca = 0.96 ± 0.11‰ (n=8, chemistry=7), in keeping with previously cited long-term reproducibility (Reynard *et al.,* 2011). The uncertainties in measured isotopic analyses are given as two standard deviations of repeated analyses.

3. Results

The temporal evolution of the chemical and isotopic composition of the fluid, as well as the mass of fluid remaining in the reactor after each sample in experiments A and B are listed in Tables 1 and 2; the temporal evolution of Ca concentration and isotopic composition during these experiments are shown in Figs. 2 and 3. The $\delta^{44/42}$Ca of the original calcite grains in experiment A was -0.25±0.08 ‰. The dissolution of calcite during the first 25 hours of the experiment performed in an aqueous 0.01 mol/kg NaCl solution leads to a fluid phase Ca concentration of 0.22×10^{-3} mol/kg and a $\delta^{44/42}$Ca of 0.01±0.03‰; this latter value is 0.26‰ greater than that of the dissolving calcite. The fluid phase is calculated to be slightly supersaturated with respect to calcite at this time, which may be the result of either analytical uncertainty on the pH, alkalinity, and aqueous Ca measurements, and/or due to the presence of minor high-energy surfaces or ultra fine particles, unobserved by SEM, in the original calcite powder. The reactor fluid was then bubbled with pure CO_2, which decreased its pH to 6.2 and provoked further calcite dissolution. The fluid phase Ca concentration increased to ~4.5×10^{-3} mol/kg after ~96 hours of elapsed time then remained close to constant. The saturation state of the reactive fluid was no greater than 0.09 from this time until the end of this experiment, consistent within uncertainty of calcite equilibrium. The isotopic composition of the fluid phase, however, evolved continuously during this

239 dissolution leg of the experiment. The $\delta^{44/42}$Ca of the fluid phase was 0.34±0.06‰ 23 hours
240 after the pH change; this increased to 0.57±0.01‰ before the fluid pH was increased after
241 218 hours of elapsed time. These values are 0.45‰ and 0.72‰ greater than the initial
242 dissolving calcite. Changing the bubbling gas in the reactor to a CO_2/N_2 mixture after 218
243 hours increased the fluid pH to 7.5 leading to calcite precipitation. Calcite precipitation
244 lowered the fluid phase Ca concentration to 1.8×10^{-3} mol/kg; this concentration was
245 constant and within analytical uncertainly of calcite-fluid equilibrium for the final 140 hours
246 of the experiment. Despite the fact that the fluid phase was in chemical equilibrium with
247 calcite over this period, the fluid-phase Ca isotopic composition evolved substantially. The
248 fluid phase $\delta^{44/42}$Ca first increased from 0.57±0.01 to 1.15±0.04‰ after the onset of
249 precipitation, then decreased to ~0.8‰ at the end of the experiment. Mass balance
250 calculations indicate that after calcite dissolution during the dissolution leg, 18% of the Ca in
251 the system was in the fluid phase. This was lowered to 9% during the precipitation leg.

252 The evolution of the fluid phase composition of experiment B is shown in Fig. 3. The
253 $\delta^{44/42}$Ca of the original calcite grains in experiment B, which were obtained from a distinct
254 container of Merck reagent grade calcite as that of experiment A, was 0.12±0.07‰. The
255 fluid in this experiment was constantly bubbled with pure CO_2 such that the fluid pH
256 remained at a near constant 6.3. Calcite dissolved rapidly, the fluid phase attained a Ca
257 concentration of 9.5×10^{-3} mol/kg, consistent, within uncertainty, with calcite equilibrium,
258 after less than 28 hours and remained near this concentration through the end of the
259 experiment. The isotopic composition of the fluid, however, increased with time throughout
260 the experiment, increasing from 0.22±0.06‰ to ~0.6‰ after the aqueous Ca concentration
261 attained a close to constant value at near to calcite-fluid chemical equilibrium conditions.
262 These fluid Ca isotopic compositions are substantially heavier than that measured in the

solid calcite before and after the experiment. After calcite-fluid equilibrium was attained, roughly 45% of the Ca in the system was present in the fluid phase.

Photomicrographs of the calcite following experiments A and B are shown in Fig 1. In both cases only calcite is present in the post-experiment solids. Similarly post-experiment analysis of the solids recovered from these experiments by XRD revealed the presence of only calcite. The calcite recovered from experiment B shows the effect of dissolution; dissolution appears to be driven by the removal of material from the corners of calcite crystals rather than due to the formation of etch pits. This behavior is consistent with dissolution occurring at near to equilibrium conditions where there is insufficient driving force to create etch pits. Note also that as 45% of the calcite dissolved during experiment B is it possible some of the smaller grains dissolved completely. The calcite recovered from experiment A has far fewer steps than that recovered from experiment B, likely due to the precipitation of calcite onto pre-dissolved grains.

4. Discussion

4.1 Isotope release during congruent calcite dissolution and at bulk equilibrium

Results indicate that Ca isotopes fractionated during congruent calcite dissolution. Evidence that calcite dissolved congruently during the dissolution experiments include 1) the results of thermodynamic calculations demonstrating that the fluid phase was either undersaturated or in equilibrium within uncertainty with respect to calcite, and undersaturated with respect to other potentially precipitating phases during the dissolution leg of experiment A, and throughout experiment B, and 2) the SEM images after the experiments indicate the presence of only calcite in the solid phase. Moreover, the temporal

evolution of reactive fluid Ca concentrations calculated using calcite dissolution and precipitation rates reported by Chou *et al.* (1989) together with calcite solubility constants generated using PHREEQC match closely those measured in the reactors (See Figs 2a and 3a). As a result of calcite dissolution in experiment A, the measured $\delta^{44/42}$Ca of the fluid phase evolved to 0.57±0.01‰ compared to the δ^{44}Ca of -0.25±0.08‰ of the original calcite. The change in the Ca isotopic composition of the calcite due to dissolution in experiment A was not determined directly; although the Ca isotopic composition of the final solids recovered from this experiment was measured, these solids experienced both dissolution and subsequent precipitation. Nevertheless, because the experiment initially consisted of calcite and a Ca-free fluid phase, the change in the calcite bulk Ca isotope composition during this experiment can be computed from the measured isotopic composition of the fluid phase via mass conservation taking account of

$$\delta^{44/42}\mathrm{Ca}_{initial\ solid} m_{\mathrm{Ca},initial\ solid} = \delta^{44/42}\mathrm{Ca}_{solid} m_{\mathrm{Ca},solid} + \delta^{44/42}\mathrm{Ca}_{fluid} m_{\mathrm{Ca},fluid} \quad (3)$$

where $m_{\mathrm{Ca},i}$ refers to the mass of calcium in the *i*th phase, and the subscripts $initial\ solid$, $solid$, and $fluid$ designate the original solid, and the solid and fluid phase present in the reactor at the time of interest. This calculation was performed iteratively using an excel spreadsheet using the data summarized in Table 1. At each time interval, delineated by the sampling times, the Ca isotope composition of calcite was determined taking account of the mass and composition of the fluid remaining in the reactor and that removed by sampling over time. This calculation indicates that following just the 'dissolution leg' of experiment A, the bulk calcite obtains a $\delta^{44/42}$Ca of -0.42‰, a decrease of 0.17‰ during 9 days of calcite-fluid interaction at ambient temperature. The decrease between the bulk $\delta^{44/42}$Ca of the calcite before and after the 8 days of dissolution during experiment B was measured directly

and determined to be 0.19‰. The degree to which the calcite was isotopically homogeneous was not determined either before or after the experiments. It is also possible that Ca isotopes are heterogeneously distributed in the solids at the end of the experiments due to sluggish solid-state transport rates. Note that although isotopic heterogeneities in the solid phase could have accounted for the difference in isotopic compositions between the bulk original calcite and the reactive fluid during the first ~24 hours of the dissolution leg of experiment A and experiment B, the calcium isotope compositions of the reactive fluids continued to increase after this fluid attained chemical equilibrium with respect to calcite.

The exact mechanism by which isotopes fractionate during congruent calcite dissolution and at equilibrium is somewhat uncertain. Either some mechanism must allow isotopically heavy Ca to be preferentially released from the calcite structure during forward dissolution or lighter calcium needs to be returned from the fluid to the calcite during reverse precipitation. The first mechanisms seems rather unlikely, as earlier work on mineral dissolution suggest that in both carbonates and silicate minerals, lighter isotopes are preferentially released to the fluid phase at the onset of dissolution. This is because lighter isotopes generally form weaker bonds that require less energy to break (see Oelkers et al., 2015, 2018; Maher et al., 2016). Moreover, the transport of calcium, either through the fluid or solid phase would favor the transfer to the fluid of light rather than heavy Ca. The second mechanism, however, seems to be more likely as isotopic fractionation favors the incorporation of light Ca into the calcite structure during its rapid precipitation (Lemarchand *et al.,* 2004; Gussone *et al.,* 2005; Reynard *et al.,* 2011; Tang *et al.,* 2008, 2012). As such it seems likely that the observed fractionation of Ca isotopes into the fluid phase during congruent dissolution and at equilibrium stems from the two-way transfer of Ca between calcite and the fluid phase. Such a process could proceed by the conservative release of Ca

from the calcite surface coupled to fractionation during its reincorporation into the solid calcite. At bulk chemical equilibrium, this overall process would both maintain a constant fluid Ca concentration and evolve the fluid-mineral system towards isotopic equilibrium. This possibility is favored by the observation that the isotopic composition of the fluid phase evolves after the fluid has attained bulk chemical equilibrium with the dissolving calcite (see Figs 2 and 3). This possibility is also consistent with the concept of dynamic equilibrium (e.g. Aagaard and Helgeson, 1982; Oelkers *et al.,* 1994; Schott and Oelkers, 1995; Oelkers, 2001; Schott et al., 2009, 2012; Steefel et al., 2014; Lui et al. 2016). Note that a number of studies have attempted to recover bulk mineral dissolution rates at close to equilibrium conditions by measuring the temporal evolution of the isotopic composition of minerals and/or their co-existing fluids (c.f. Gruber *et al.,* 2013; Zhu et al., 2014; Subhas et al., 2015, 2017; Liu et al., 2016; Naviaux et al., 2019). It follows from the results presented in this study that such efforts need to take explicit account of the role of mineral-fluid isotope exchange during congruent dissolution and at bulk equilibrium to accurately retrieve near to equilibrium bulk dissolution rates.

The observations of isotopic equilibration during the congruent dissolution and at equilibrium in the present study differ fundamentally from the resetting of isotopic compositions of biogenic carbonates as reported by Chanda et al. (2019). This former study considered the temporal evolution of Ca isotope compositions in carbonate minerals that contained substantial Mg in its structure placed in an initially Mg-free aqueous solution. This resulted in the incongruent dissolution of the original biogenic carbonate, where the original Mg bearing carbonate was replaced by a more stable Mg-poor calcite. The observed Ca isotope composition evolution in the solids was thus driven by the thermodynamic instability of the solids placed in the reactor. In contrast, the calcite in the present study dissolved

congruently and approached a stable bulk chemical equilibrium. Isotopic evolution in the experiments performed in the present study was thus driven by the isotopic disequilibrium in the water-mineral system.

Although the net mass of total calcium transferred from the calcite to the fluid phase via calcite dissolution is readily determined from the aqueous Ca concentrations (c.f. Fig 2b), it is challenging to estimate the fraction of the Ca in the original calcite transferred to the fluid and that reincorporated into the solid during each bulk reactor dissolution experiment. A crude estimate can be made by assuming that the Ca isotopic composition of the solid at the end of the dissolution experiments consisted of a fraction, *f*, of the Ca from the original bulk calcite and a fraction 1-*f* of Ca in isotopic equilibrium with the final fluid with a fractionation factor equal to $\mathbf{\Delta}^{44/42}\mathrm{Ca}_{\text{calcite-fluid}}$. Note such estimates do not take account of changing fluid compositions during water-mineral interaction, so tend to underestimate the overall mass of calcium exchanged. Taking account these assumptions and mass balance requires that

$$\delta^{\frac{44}{42}}\mathrm{Ca}_{final\ solid} = f\delta^{44/42}\mathrm{Ca}_{initial\ solid} + (1-f)\delta^{44/42}\mathrm{Ca}_{final\ fluid} - \mathbf{\Delta}^{44/42}\mathrm{Ca}_{calcite-fluid} \quad (4)$$

The results of this calculation obtained from Eqn. (4) are shown in Fig 4 for the case of the dissolution leg of experiment A. In accord with these calculations, more than 90% of the Ca present in the original calcite powder would have equilibrated isotopically with the fluid phase during the dissolution leg of this experiment to attain the fluid phase $\delta^{44/42}$Ca observed at the end of the dissolution experiment for all $\mathbf{\Delta}^{44/42}\mathrm{Ca}_{\text{calcite-fluid}} < -0.9$ and more than 30% for all $\mathbf{\Delta}^{44/42}\mathrm{Ca}_{\text{calcite-fluid}} < -1.6$. The observed fluid phase $\delta^{44/42}$Ca could not be attained with any $\mathbf{\Delta}^{44/42}\mathrm{Ca}_{\text{calcite-fluid}} < -0.82$. This comparison suggests that a majority of the

calcium present in the initial calcite grains were transferred into and out of the fluid phase during the 8 to 9 day dissolution experiments performed in this study.

4.2 Ca Isotope evolution during precipitation

The variation of reactive fluid phase $\delta^{44/42}$Ca as a function of time during the precipitation leg of experiment A is similar to that predicted by Steefel et al. (2014) for the precipitation of calcite in a closed system reactor. An initial stage of Ca isotope fractionation is observed due to the kinetically controlled preferred incorporation of light Ca into the precipitated calcite during the rapid precipitation of the mineral. This initial stage is followed by a longer stage of isotopic re-equilibration. Within the Steefel et al. (2014) formalism, this isotopic re-equilibration is a direct consequence of dynamic equilibrium; the combined effects of reverse precipitation coupled to forward dissolution, which are equal at bulk chemical equilibrium, tends to drive the mineral-fluid system towards isotopic equilibrium after the calcite has precipitated.

Further insight into the evolution of the fluid phase calcium isotopic composition during the calcite precipitation experiment can be gained with the aid of Fig. 5, which compares the evolution of reactive fluid $\delta^{44/42}$Ca as a function of *F*, the fraction of Ca precipitated from the fluid during the experiment. The curves in this figure were calculated assuming Rayleigh fractionation and equilibrium fraction (see Johnson *et al.,* 2004; Pearce *et al.,* 2012) by assuming a calcite-fluid equilibrium fractionation factor of 0.4. Rayleigh fractionation is consistent with the one-way transfer of material from the fluid phase into the solid, whereas equilibrium fractionation requires the two-way transfer of material to and from the precipitated solid. It can be seen that measured reactive fluid δ^{44}Ca is initially consistent with Rayleigh fractionation, but that $\delta^{44/42}$Ca evolves to become consistent with

equilibrium fractionation later in the experiment, while the calcite-reactive fluid system is at chemical equilibrium within analytical uncertainty. This behavior is itself consistent with kinetic theory; precipitation at far from equilibrium conditions is dominated by the transfer of material from the fluid to the solid, whereas two way transfer dominates near and at equilibrium in accord with the concept of dynamic equilibrium (e.g. Schott and Oelkers, 1995; Oelkers, 2001; Schott *et al.,* 2009; see also Pearce *et al.,* 2012).

An open question remains as the origin of the distinct behavior of calcium isotopic transfer during dissolution and during precipitation in these experiments. The Ca isotope fractionation during mineral precipitation was consistent with an equilibrium Ca fractionation factor of ~0.4, whereas that observed during calcite dissolution apparently exceeds 0.8. One possible explanation for these contrasting equilibrium fractionation factors is the distinct pH of the two experiments. The dissolution experiment presented in this study was performed at pH 6.2, whereas the precipitation experiment was performed at pH 7.5. Calcite dissolution rates are approximately 1.5 orders of magnitude faster at pH 6.2 compared to pH 7.5 (e.g. Chou et al., 1989; Cubillis et al., 2005). According to Tang *et al.* (2008) this difference in reaction rates can more than double the Ca fractionation factor between calcite and its co-existing aqueous fluid. The difference in pH of the fluid phase however may also provoke a change in the Ca isotope fractionation due a change in aqueous Ca speciation. The isotopic fractionation among aqueous species stems from the change in coordination observed between aqueous Ca^{2+} and Ca-bearing complexes (see Moynier and Fujii, 2017) and has been experimentally shown to be an important fractionation mechanism for a number of divalent cations such as Mg and Zn (Schott et al., 2016; Mavromatis et al., 2019).

An alternative explanation for different observed fractionation factors in the different experimental conditions is that the closed system experiments may not have completely attained fluid-mineral isotopic equilibrium. Such uncertainties could be overcome through the use of the three-isotope method to determine unambiguously both isotopic exchange rates at bulk mineral fluid equilibrium as well as the equilibrium isotopic fractionation factor in aqueous fluid-mineral systems (e.g. Beard *et al.*, 2010; Li *et al.,* 2011, 2014; Frierdich *et al.,* 2014; Reddy *et al.,* 2015; Zheng *et al.*, 2016; Stamm *et al.*, 2018).

4.3 Preservation of isotopic signatures in natural systems

The results reported in this study appear to contravene the commonly held assumption that calcite can preserve its isotopic signatures over geological timeframes if fluid-mineral isotopic disequilibrium exists, even if the fluid-mineral system is at bulk chemical equilibrium. This observation is not unique to either calcite or the carbonate minerals. Similar observations have been reported for the carbonates dolomite (Perez-Fernandez et al., 2017), hydromagnesite (Oelkers et al., 2018) and other hydrous Mg-carbonates (Mavromatis et al., 2012, 2015), strontianite (Mavromatis et al., 2017a), and witherite (Mavromatis et al., 2016), as well as non-carbonate minerals including goethite (Beard *et al.,* 2010; Ready *et al.,* 2015), epsonite (Li *et al.,* 2011), brucite (Li *et al.,* 2014), quartz (Lui et al., 2016), and amorphous SiO_2 (Stamm *et al.,* 2018). It follows that isotopic preservation of the original Ca signature of calcite would require either the isolation of the calcite-fluid system from external Ca input, by a process that decreases dramatically the rock permeability or the slowing of the coupled mineral dissolution/precipitation reactions. There is some indication that such processes are common in some natural systems. Turchyn and

DePaolo (2011) noted that much anecdotal evidence suggests that carbonate fossils are best preserved in clay or organic-rich sedimentary horizons. The presence of clay minerals can dramatically lower host rock permeability (e.g. Klimentos and McCann, 1990; Revel and Cathels, 1999; Luijendijk and Gleeson, 2015) and thereby aid in the isolation of the fluid system from external inputs. As a relatively small fraction of the Ca in most natural rocks is present in the fluid phase, changes in fluid-mineral Ca isotopic fractionation due to temperature changes would negligibly alter the isotopic signature of the calcite present in an isolated system. Similarly, the presence of aqueous organic compounds may aid in the preservation of isotopic compositions of calcite. Although they likely do not alter significantly the dissolution rates of calcite (Oelkers *et al.,* 2011), the presence of aqueous organic compounds can dramatically inhibit calcite precipitation (e.g. Meldrum and Hyde, 2001; de Leeuw and Cooper, 2004; Lakshtnov *et al.,* 2011; Nielsen et al., 2012; Mavromatis et al., 2017c). As such it appears that Ca isotopic equilibration in the presence of aqueous organic compounds may be slowed by a decrease in the rate of attachment of Ca to the calcite surface. Moreover the presence of aqueous Mg has also been observed to slow calcite precipitation rates (e.g. Fernandez-Diaz et al., 1996; Morse et al., 2007).

The results presented above suggest that isotopic exchange in calcite may be rapid and stems from a coupled forward dissolution/reverse precipitation process. It seems likely, therefore that the preservation of isotopic signals would be favored in minerals that have relatively slow dissolution rates, or those that do not precipitate at ambient temperatures. For the case of preserving calcium isotopic signals, apatite dissolves ~5 orders of magnitude slower at ambient conditions and seawater pH (c.f. Plummer *et al.* 1979; Chou *et al.,* 1989; Valsami-Jones *et al.,* 1998; Cubillis *et al.,* 2005; Chaïrat *et al.,* 2007). It therefore seems likely that apatite may be a far better guardian of original Ca isotopic signatures than calcite. For

the case of carbonate minerals, dolomite or magnesite do not appear to precipitate via abiotic processes at ambient temperatures (e.g. Saldi *et al.,* 2009, 2012; Gautier et al., 2014). As such these minerals may better preserve original C and Mg and/or Ca isotopic signatures than calcite.

5. Conclusions

The results of this study demonstrate isotopic compositions of calcite and its co-existing fluid phase can be reset rapidly by congruent dissolution, precipitation, and at equilibrium. As such the preservation of original Ca isotopic signatures in calcite may require that the calcite is isotopically isolated from its surroundings, or a not yet to be identified preservation mechanism is at play. Moreover, if the release of metals from minerals is isotopically non-conservative as seems to be the case for calcite, it may not possible to use the stable isotopic compositions of fluids to determine the source of metals, for example of environmental pollutants. Such conclusions call for further investigation of the rates and mechanisms of near to bulk equilibrium isotopic exchange to assess how best to interpret the isotopic signals recorded in minerals and natural fluids.

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Figure Captions

Figure 1. SEM images of the calcite powder used in the experiments a) the initial calcite prior to experiment A, b) the calcite grains recovered after the experiment A, and c) calcite grains recovered after experiment B.

Figure 2: Temporal chemical and isotopic evolution of the fluid phase during experiment A: a) Measured fluid phase Ca concentration, the error bars in this figure correspond to an estimated 4% uncertainty in these analyses; b) fraction of the total Ca preset in the system within calcite, the error bars in this plot represent an estimated 2% uncertainty in these values; c) fluid Ca isotopic composition; the error bars in c are taken as two standard deviations of repeated analyses. Note that the pH of the fluid phase was changed after 25 and after 125 hours, by changing the identity of the gas bubbling through the reactor. The timing of these pH changes is indicated by the vertical lines. The thin dashed curves in a were calculated using calcite dissolution/precipitation rate equations of Chou et al. (1989) whereas those in b and c are for the aid of the reader. The thick dashed line in c corresponds to the Ca isotopic composition of the original dissolving solid its associated 2 standard deviation uncertainty – see text and Table1.

Figure 3: Temporal chemical and isotopic evolution of the fluid phase during experiment B: a) Measured fluid phase Ca concentration, the estimated 4% uncertainty in these analyses is within the size of their symbol; b) fluid Ca isotopic composition; the error bars in b are taken as two standard deviations of repeated analyses. The dashed curve in a were calculated using calcite dissolution/precipitation rate equations of Chou et al. (1989) whereas that in b is for the aid of the reader. The thick grey line in c corresponds to the Ca isotopic composition of the original dissolving solid and its associated 2 standard deviation uncertainty – see text and Table1.

Figure 4. Calculated fraction of calcite present in the inital calcite (*f*) required to pass through the fluid phase to obtain the observed fluid phase Ca isotopic composition at the end of the dissolution leg of Experiment A, as a function of the equilibrium calcite-fluid Ca fractionation factor.

Figure 5: The Ca isotopic composition of the fluid phase provoked by calcite precipitation during the final 240 hours of experiment A plotted as a function of the fraction of the reactive fluid Ca incorporated into the solid. The solid curve represents the fluid composition of this fluid calculated using a Rayleigh distillation model whereas the linear curve corresponds to the fluid composition calculated with an equilibrium fractionation model. In each case the solid curves were calculated by adopting the calcium-fluid fractionation factor, $\Delta^{44/42}Ca_{fluid\text{-}solid}$ = -0.4.

Table 1. Measured chemical and isotopic compostion of the fluid and solid phases during experiment A. Note that the pH of the fluid phase was changed after 25 and 218 hours of elapsed time by changing the composition of the gas bubbled through the reactor – see text.

Sample	Elapsed time (hours)	Reactive fluid mass (g)[a]	*pH*	c_{Ca} (mol/kg x10^3)	Alkalinity (eq/kg x10^3)	Percent aqueous Ca[b]	$\delta^{44/42}$Ca[c]	Saturation index[d]
A-1	25	948.2	8.39	0.22	5	0.9	0.01±0.03	0.24
A-2	48	933.3	6.23	3.85	29	15.6	0.28±0.11	-0.02
A-3	73	918.3	6.23	4.99	33	20.1	0.34±0.06	0.10
A-4	96	902.4	6.19	4.14	32	16.8	0.36±0.07	-0.02
A-5	145	886.6	6.19	4.21	32	17.0	0.55±0.04	-0.02
A-6	170	870.7	6.20	4.80	33	19.3	0.27±0.04	0.05
A-7	196	854.4	6.18	4.99	36	20.0	0.51±0.12	0.07
A-8	218	837.6	6.18	4.81	35	19.3	0.57±0.01	0.05
A-9	242	810.0	7.43	1.65	3.8	8.0	0.72±0.07	0.08
A-10	265	780.7	7.49	1.75	2.7	8.4	1.06±0.06	0.02
A-11	289	751.6	7.56	1.75	2.7	8.4	1.15±0.04	0.09
A-12	337	722.4	7.46	1.94	2.8	9.0	0.75±0.09	0.05
A-13	363	692.4	7.48	1.81	2.8	8.6	0.77±0.01	0.04
A-14	385	660.9	7.49	1.87	2.4	8.7	0.92±0.04	0.00
Initial Calcite							-0.25±0.08	
Final Calcite							-0.41±0.01	

a) Reactive fluid remaining in the reactor after the fluid sample was collected as determined by weighing the fluid samples. b) Percent of calcium in the batch reactor present in the aqueous phase as calculated from mass balance constraints. c) All $\delta^{44/42}$Ca listed in this table were measured; listed +/- uncertainties refer to two standard deviations of repeated analyses. d) Saturation index of the sampled aqueous fluid with respect to calcite calculated using the PHREEQC computer code together with its minteq.v4 database. It is estimated that the uncertainty in calculated saturation indexes are approximately ±0.1.

Table 2. The measured chemical and isotopic compostion of the fluid and solid phases during experiment B.

Sample	Elapsed time (hours)	Reactive fluid mass (g)[a]	*pH*	c_{Ca} (mol/kg x10^3)	Alkalinity (eq/kg x10^3)	Percent Aqueous Ca[b]	$\delta^{44/42}$Ca[c]	Saturation index[d]
B-1	3	984.7	6.39	8.77	8.32	43.2	0.15±0.08	-0.01
B-2	5.5	965.7	6.21	9.31	11.49	45.9	0.22±0.06	-0.05
B-3	28	947.6	6.28	9.27	11.13	45.7	0.45±0.05	0.01
B-4	45	927.5	6.23	9.40	11.26	46.3	0.42±0.02	-0.03
B-5	70	909.3	6.25	9.42	11.24	46.4	0.50±0.04	-0.01
B-6	101	891.4	6.23	9.44	10.74	46.5	0.52±0.10	-0.03
B-7	121	875.7	6.22	9.46	11.02	46.6	0.64±0.10	-0.05
B-8	149	858.0	6.25	9.64	11.04	47.3	0.59±0.04	-0.01
B-9	172	838.5	6.25	9.66	12.03	47.4	0.52±0.03	0.02
B-10	193	814.2	6.26	9.09	10.64	45.1	0.64±0.04	-0.04
Initial Calcite							0.12±0.07	
Final Calcite							-0.07±0.02	

a) Reactive fluid remaining in the reactor after the fluid sample was collected as determined by weighing the fluid samples. b) Percent of calcium in the batch reactor present in the aqueous phase as calculated from mass balance constraints. c) All $\delta^{44/42}$Ca listed in this table were measured; listed +/- uncertainties refer to two standard deviations of repeated analyses. d) Saturation index of the sampled aqueous fluid with respect to calcite calculated using the PHREEQC computer code together with its minteq.v4 database. It is estimated that the uncertainty in calculated saturation indexes are approximately ±0.1.

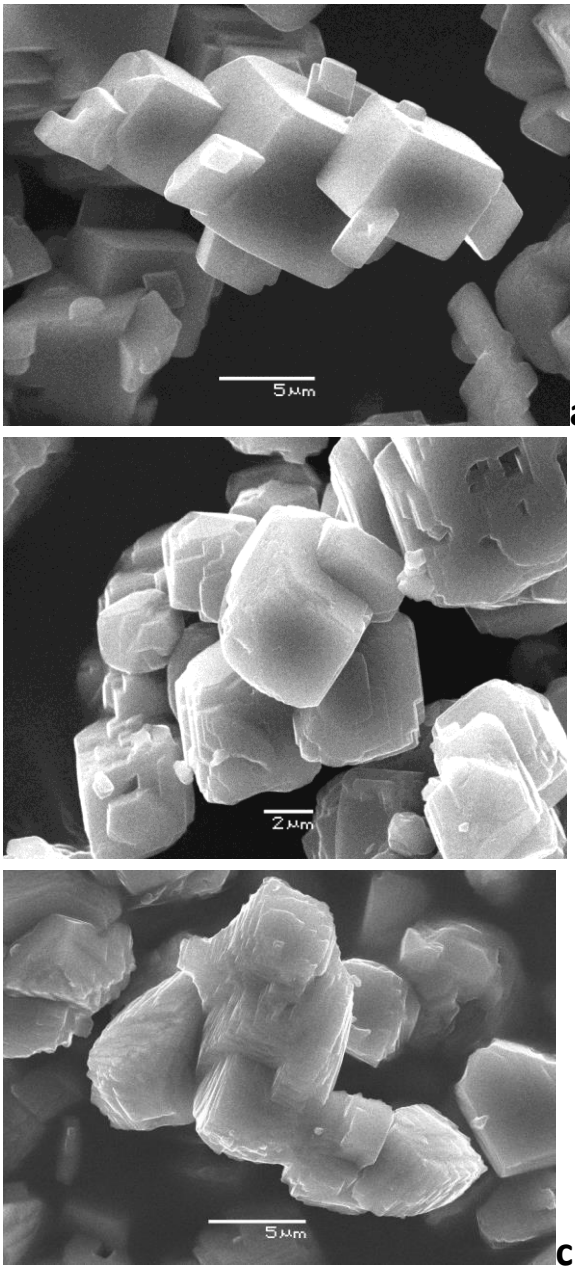

Figure 1.

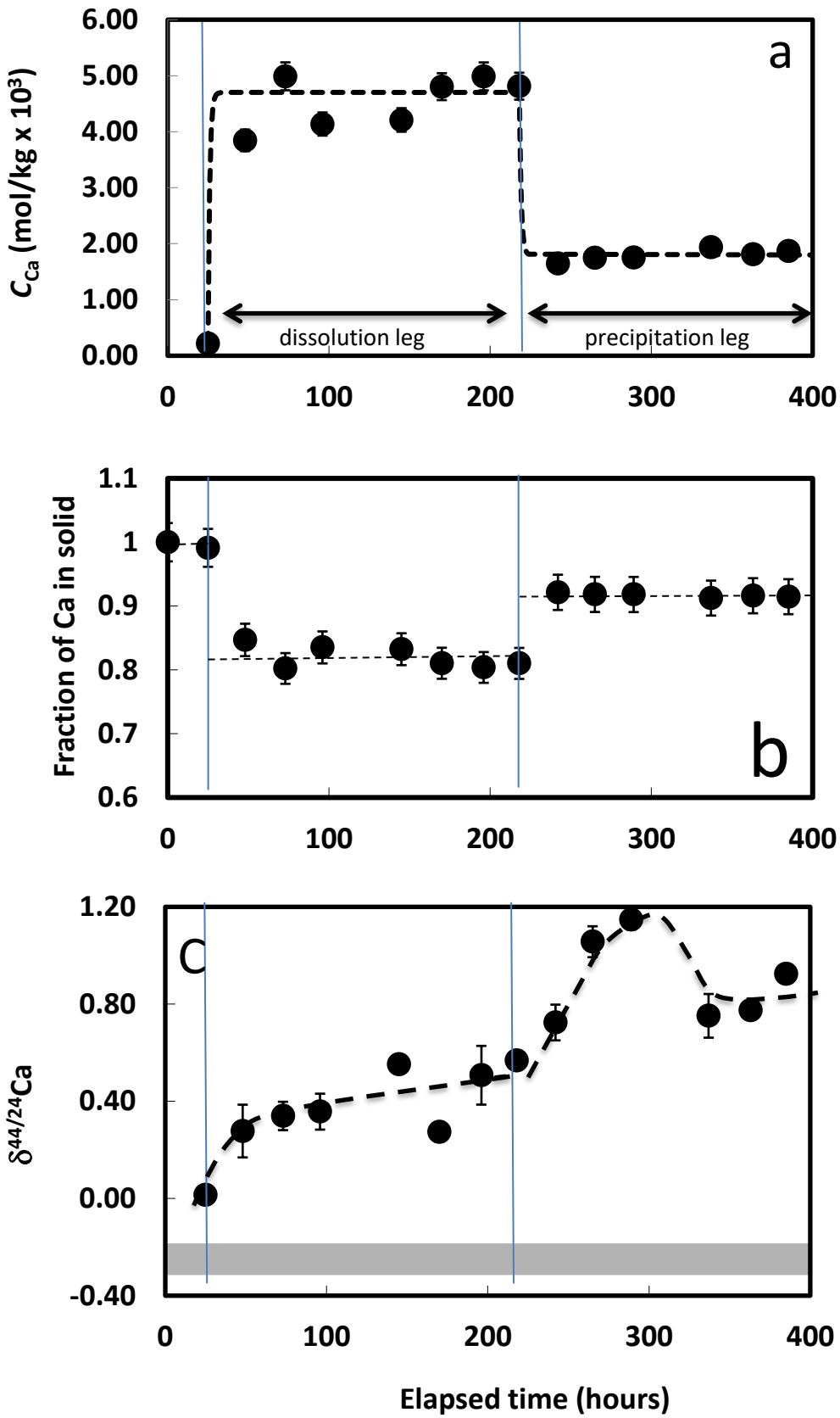
a
C_{Ca} (mol/kg x 10^3)
6.00
5.00
4.00
3.00
2.00
1.00
0.00
dissolution leg
precipitation leg
0
100
200
300
400
b
Fraction of Ca in solid
1.1
1
0.9
0.8
0.7
0.6
c
$\delta^{44/24}$Ca
1.20
0.80
0.40
0.00
-0.40
Elapsed time (hours)

FIGURE 2:

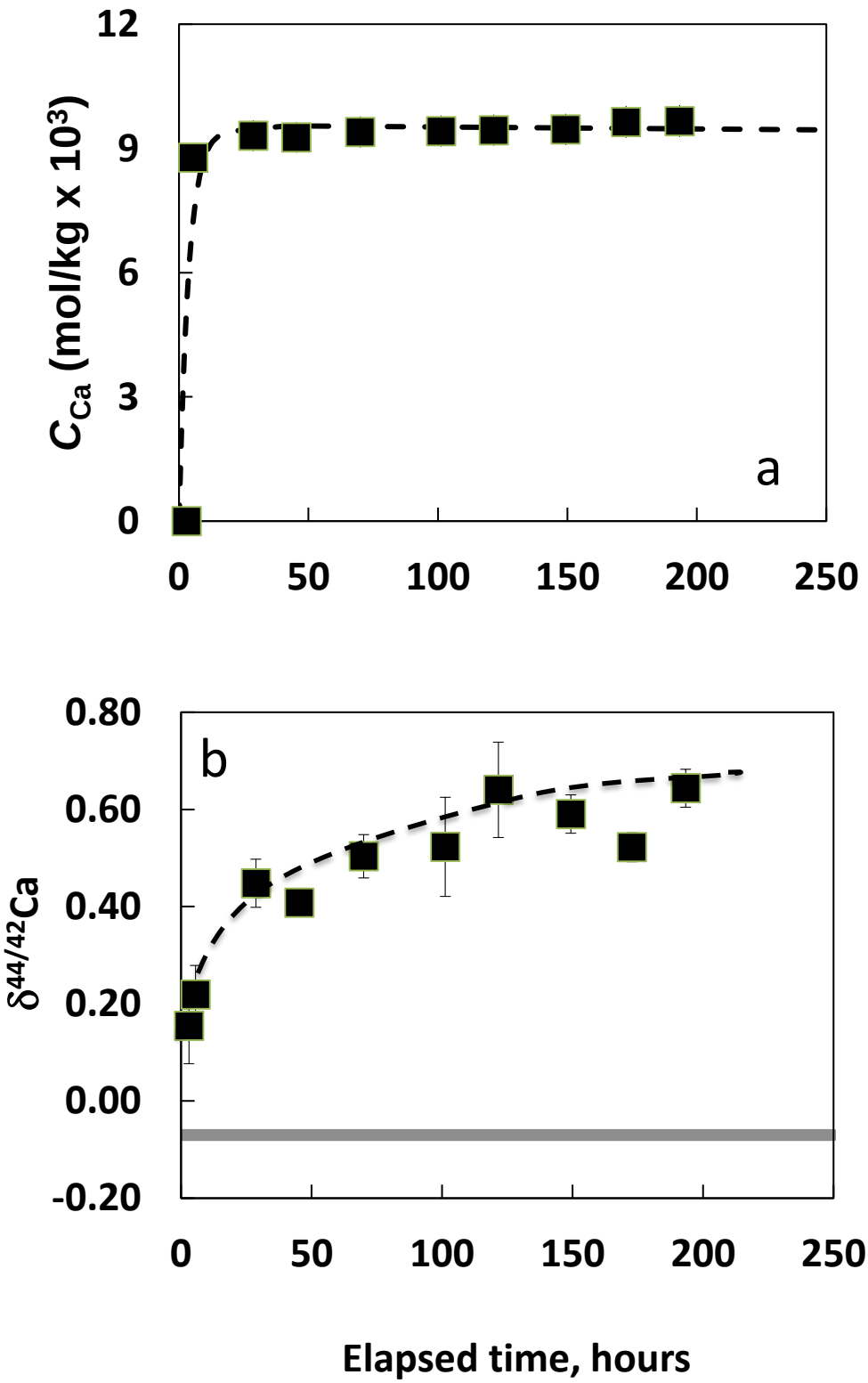
12
9
6
3
0
C_{Ca} (mol/kg x 10^3)
a
0
50
100
150
200
250
0.80
0.60
0.40
0.20
0.00
-0.20
$\delta^{44/42}$Ca
b
0
50
100
150
200
250
Elapsed time, hours

Figure 3.

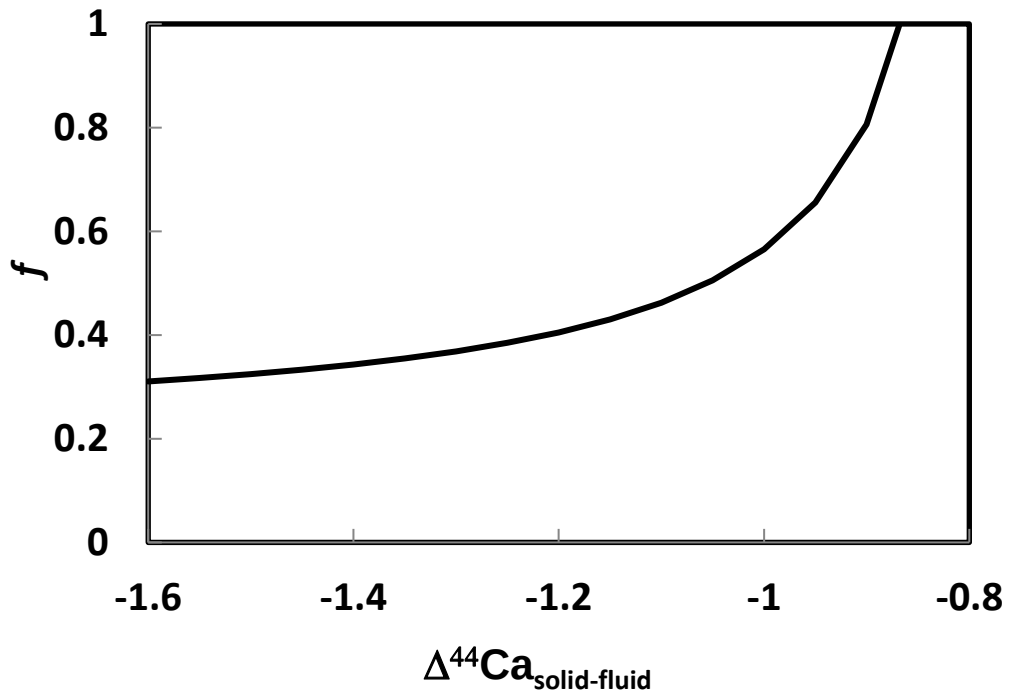

Figure 4.

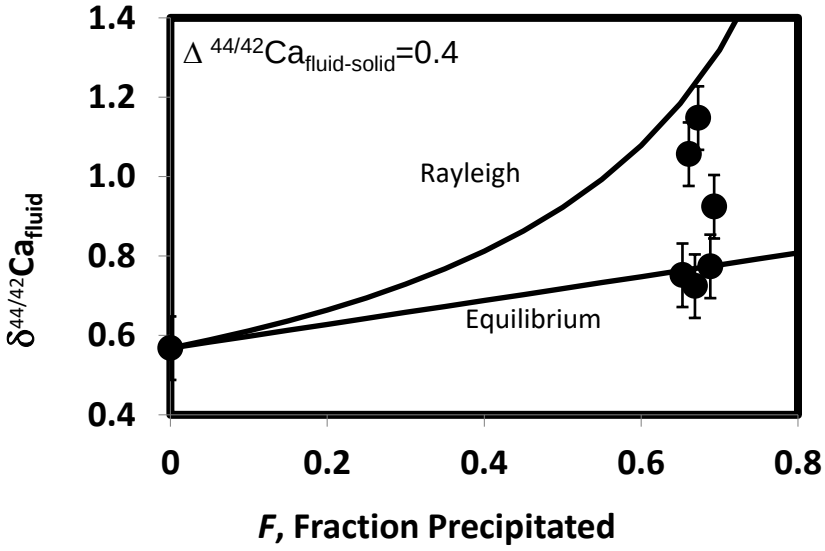
1.4
1.2
1.0
0.8
0.6
0.4
0
0.2
0.4
0.6
0.8
Δ 44/42Ca fluid-solid=0.4
Rayleigh
Equilibrium
δ44/42Ca fluid
F, Fraction Precipitated

Figure 5