Experimental partitioning of Ca isotopes and Sr into anhydrite: Consequences for the cycling of Ca and Sr in subseafloor mid-ocean ridge hydrothermal systems


Abstract

The elemental and isotopic mass balance of Ca and Sr between seawater and the oceanic crust at mid-ocean ridge (MOR) hydrothermal systems integrates various physiochemical processes in the subseafloor, such as dissolution of primary silicate minerals, formation of secondary minerals, and phase separation in the subseafloor. In particular, the precipitation and recrystallization of anhydrite are recognized as important processes controlling the Ca and Sr elemental and isotope composition of high temperature vent fluids and coexisting ocean crust, and yet, little experimental data exist to constrain the mechanism and magnitude of these critical geochemical effects. Thus, this study experimentally examines Sr/Ca partitioning, Ca isotope fractionation, and the rate of exchange between anhydrite and dissolved constituents. These experimental constraints are then compared with Sr/Ca and Ca isotope compositions of anhydrite and vent fluids sampled from the TAG hydrothermal system. Accordingly, anhydrite precipitation and recrystallization experiments were performed at 175, 250, and 350 °C and 500 bar at chemical conditions characteristic of active MOR hydrothermal systems. Experimental data suggest that upon entrainment and recharge of seawater into MOR hydrothermal systems anhydrite will rapidly precipitate with a Ca isotopic composition that is depleted in the heavy isotope compared to the hydrothermal fluid. The magnitude of the Ca isotope fractionation, Δ⁴⁴/⁴⁰Ca(Anh-Fluid), is temperature dependent, -0.45, -0.22, and -0.02‰, for 175, 250, and 350 °C, respectively, but likely indicative of kinetic effects. Utilization of a ⁴³Ca spike in solution was implemented to quantify the time-dependent extent of isotope exchange during anhydrite recrystallization at chemical equilibrium. These data indicate that the rate of exchange is a function of temperature, where 12, 46, and 45% exchange occurred within 1322, 867, 366 h at 175, 250, and 350 °C, respectively. The partitioning of Sr/Ca between anhydrite and constituent dissolved species during precipitation depends greatly on the saturation state of the hydrothermal fluid with respect to anhydrite at each experimental temperature, K_D(Anh-Fluid) = 1.24-0.55 at 175–350 °C, broadly similar to results of earlier experimental observations by Shikazono and Holland (1983). Equilibrium K_D(Anh-Fluid) values were estimated by taking explicit account of time dependent magnitude of exchange, yielding values of 0.43, 0.36, 0.29 at 175, 250, and 350 °C, respectively. Coupling these experimental constraints with the temperature gradient inferred for high temperature MOR hydrothermal systems suggests that the Ca isotope and Sr elemental composition of anhydrite formed near the seafloor will retain the composition derived upon initial formation conditions, which is indicative of disequilibrium. In contrast, at greater depths and at higher temperatures, anhydrite will reflect...
close to equilibrium Sr/Ca partitioning and Ca isotope fractionation conditions. The experimental and natural data presented in this study can be used to further understand the effect of anhydrite precipitation during hydrothermal circulation in the oceanic crust and on the chemical and isotopic composition of seawater on geologic timescales.

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**Keywords:** Anhydrite; Ca isotopes; Sr/Ca partitioning; Mid-ocean ridge hydrothermal systems

1. INTRODUCTION

The formation of anhydrite (CaSO$_4$) in active high-temperature mid-ocean ridge (MOR) hydrothermal systems occurs as a consequence of conductive heating of Ca$^{2+}$ and SO$_4^{2-}$-bearing seawater that is recharged into the oceanic crust and during mixing of high temperature hydrothermal fluid with seawater at vent sites on the seafloor (Fig. 1). This reaction is thought to be widespread in modern oceanic mid-ocean ridge hydrothermal systems due to the retrograde solubility of anhydrite at pressure, temperature, and chemical conditions indicative of MOR hydrothermal systems and the high Ca$^{2+}$ and SO$_4^{2-}$ contents of modern seawater (Bischoff and Seyfried, 1978; Seyfried and Ding, 1993; Newton and Manning, 2004). However, in spite of the ubiquitous occurrence of anhydrite in high temperature MOR hydrothermal systems, very little anhydrite is preserved in older oceanic crust and the importance of subsurface anhydrite as a long-term sink of seawater SO$_4^{2-}$ and/or Ca$^{2+}$ is controversial (Sleep, 1991; Alt, 1995; Teagle et al., 1998b; Alt et al., 2003). In addition to MOR hydrothermal systems, anhydrite also persistently forms in oceanic back-arc basin hydrothermal systems and in continental hosted porphyry hydrothermal deposits as a consequence of heating and mixing of fluids that are concentrated in Ca$^{2+}$ and SO$_4^{2-}$. Geologic environment notwithstanding, anhydrite formation has important geo-physical and geochemical implications for fluid flow in the crust, elemental cycling between rock and fluid reservoirs, and hydrothermal processes such as redox buffering and mineral deposition associated with the formation of economic ore deposits (Lowell et al., 2003; Ono et al., 2007; Tolstoy et al., 2008; Craddock et al., 2010; Peters et al., 2010; Blundy et al., 2015; McDermott et al., 2015; Nadeau, 2015).

Geochemical tracers that have been applied to the study of anhydrite formation in subseaflour hydrothermal systems include Sr content and $^{87}$Sr/$^{86}$Sr ratios, $^{18}$O/$^{16}$O ratios, and multiple S isotope ratios ($^{34}$S/$^{32}$S and $^{32}$S/$^{34}$S) (Chiba et al., 1998; Teagle et al., 1998a; Mills and Tivey, 1999; Ono et al., 2007; Peters et al., 2010). These applications benefit from experimental studies bearing on the redox and temperature dependent stability of anhydrite at elevated temperature and pressures (Ohmoto and Rye, 1979; Chiba et al., 1981; Ohmoto and Lasaga, 1982). In combination, these data have shown that seawater pervasively enters the subsaeafloor and mixes with high temperature hydrothermal fluids, inducing the formation of extensive sulfide deposits at and below the seafloor. In addition, these data have helped to constrain the degree of water-rock interaction in the subseaflloor, while further documenting microbial effects that influence mass transfer of elements between seawater and the oceanic crust. However, complementary elemental and isotopic systems, such as Ca, have not been studied extensively and may provide important additional information that can further elucidate the temporal and spatial evolution of marine hydrothermal systems.

Previous experimental efforts precipitated anhydrite upon heating Ca and SO$_4$ bearing fluid, although equilibrium phase relations were not determined unambiguously (Shikazono and Holland, 1983). The experimental data reported here not only address anhydrite solubility, but also quantify the magnitude and sign of fractionation of Ca isotopes during anhydrite precipitation. These data benefit from careful analysis of the effect of temperature and time on the rate of exchange of Ca isotopes between anhydrite and dissolved Ca$^{2+}$ species, with implications for interpretation of the physiochemical characteristics of active MOR and oceanic/continental volcanic arc hydrothermal systems, especially since anhydrite precipitation has been invoked as an important process in the Ca isotope budget of subseaflloor hydrothermal systems but comparable experimental data are lacking (Amini et al., 2008).

With the continued advancement of multi-collector ICP-MS (MC-ICP-MS) technology and increasingly more sophisticated techniques and procedures for isotopic analysis and experimentation, the Ca isotope composition of hydrothermal fluids and coexisting minerals from MOR systems are now being determined to better understand subseaflloor hydrothermal processes. However, there are few experimental data with respect to the Ca isotope system to broaden the interpretation of fluid and mineral alteration processes in natural hydrothermal systems. Furthermore, Ca isotopes can be used to trace rates and mechanisms of mass transfer in experimental studies, with important implications for testing equilibrium Sr/Ca partitioning, $K_D$, between anhydrite and fluid, despite incomplete exchange between distinct chemical reservoirs.

This study examines two important geochemical tracers, Ca isotope fractionation and partitioning of elemental Sr, during precipitation and recrystallization of anhydrite at 175, 250, and 350 °C and 500 bars at controlled chemical conditions in hydrothermal experiments. Experimental constraints are then compared with Ca and Sr elemental and isotopic compositions of hydrothermal fluid – anhydrite pairs recovered from the MOR hydrothermal systems, Logatchev and TAG, as presented by Amini et al. (2008) and this study, respectively. In addition, the experimental and natural data presented in this study are also compared with a limited number of measured end-member hydrothermal fluids from a variety of MOR hydrothermal systems.

(Schmitt et al., 2003). The Ca isotope data derived from TAG are also compared with previously measured stable O isotope, radiogenic Sr isotope, and fluid inclusion homogenization temperature data of anhydrite recovered from drill core from the TAG mound (Chiba et al., 1998; Humphris, 1998; Teagle et al., 1998a; Humphris and Bach, 2005). The experimental constraints applied to the natural anhydrite and hydrothermal fluid data provide a better understanding of Sr/Ca partitioning and Ca isotope fractionation systematics at MOR and volcanic arc hydrothermal systems.

2. METHODS

2.1. Experimental hydrothermal batch reactions

This study employs a novel experimental method utilizing flexible gold cell reactor technology coupled with the use of Ca isotopes as a tracer of reaction progress (Fig. 2) (Seyfried et al., 1987; Berndt et al., 1988; Beck et al., 1992). The experimental system, housed at the University of Minnesota, permits time-series sampling of reactor solution and injection of chemically anomalous solutions into...
the gold cell reactor at experimental conditions. Two different solutions were introduced into the experiment at desired temperature and pressure conditions: (1) a \( \text{SO}_4^{2-} / \text{CO}_3^- \) bearing solution and (2) an enriched \( ^{43}\text{Ca} \) spike solution. The former is used to induce the precipitation of anhydrite from a \( \text{CaCl}_2 \)-bearing solution, whereas the addition of the \( ^{43}\text{Ca} \) spike solution is used to quantify the extent and rate of isotope exchange between precipitated anhydrite and dissolved \( \text{Ca}^{2+} \) upon precipitation (Fig. 2).

Experiments are initiated at ambient conditions by addition of approximately 50 grams of synthetic Na–Ca–Sr–Cl solution of near seawater ionic strength composed of approximately 20 mmol/kg \( \text{CaCl}_2 \) – 135 \( \mu \text{mol/kg} \) \( \text{SrCl}_2 \) – 550 mmol/kg \( \text{NaCl} \) into the gold cell reactor, which is subsequently sealed (Table 1 and Fig. 2). The gold cell reactor assembly is then transferred and sealed into a steel autoclave pressure vessel, pressurized, and brought to the desired experimental temperature in a power-proportioning controlled furnace monitored by a Type K thermocouple in direct contact with the \( \text{H}_2\text{O} \) confining medium. The fluid in the gold cell reactor is sampled at each condition prior to anhydrite precipitation, at 175, 250, and 350 \( ^\circ \text{C} \) and 500 bars, to constrain the bulk chemical and Ca isotope composition of the experimental system (i.e. \( \Sigma \text{Ca} \) and \( \Sigma \text{Sr} \)). After the initial sampling, anhydrite precipitation commenced upon the metered delivery of approximately 5–7 g of 250 mmol/kg \( \text{Na}_2\text{SO}_4 \) solution (pH = 6.5). The added \( \text{Na}_2\text{SO}_4 \) solution was preceded and followed by injection of approximately 1 g of 550 mmol/kg \( \text{NaCl} \) solution to prevent anhydrite precipitation in the sampling tubing and to ensure quantitative delivery of the \( \text{SO}_4^{2-} \)-bearing solution into the gold cell reactor, respectively. The amount of \( \text{Na}_2\text{SO}_4 \) solution injected into the gold cell reactor was designed such that approximately 30–40% of the total \( \text{Ca} \) remained in the fluid phase after the anhydrite precipitation event (Tables 1 and 2). The first solution sample after the onset of the anhydrite precipitation was typically taken within the first hour to track chemical and isotopic changes associated with the rapid precipitation of anhydrite. Sampling continued prior to the injection of the \( ^{43}\text{Ca} \) spike to further characterize Sr/Ca partitioning and fractionation of Ca isotopes during the initial period of recrystallization. After allowing for isotopic and elemental exchange between anhydrite and coexisting dissolved components to approach chemical saturation (Table 1 and Fig. 3), an anomalous solution enriched in \( ^{43}\text{Ca} \) was added to the contents of gold cell reactor by use of the high pressure ISCO™ pump, following similar procedures as previously described above for the \( \text{Na}_2\text{SO}_4 \) solution. Approximately 2–3 g of the \( ^{43}\text{Ca} \) isotope solution, with a \( ^{43}\text{Ca}/^{40}\text{Ca} \) ratio and \( \text{Ca}^{2+} \) concentration of 0.35 and 0.54 mmol/kg, respectively, were introduced to the experimental system, resulting in a mixed Ca isotope composition of the gold cell solution ranging between approximately 180 and 380‰, in \( \delta \)-notation relative to the \( ^{43}\text{Ca}/^{40}\text{Ca} \) ratio of modern seawater (Tables 1 and 2). The
Table 1

Time-series solution samples and product anhydrite derived from the anhydrite precipitation and recrystallization experiments at 175, 250, and 350 °C and 500 bars.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Time (h)</th>
<th>pH</th>
<th>Sln. mass (g)</th>
<th>[Ca2+] (mM)</th>
<th>[Sr2+] (µM)</th>
<th>[SO42−] (mM)</th>
<th>[Na+] (mM)</th>
<th>[Cl−] (mM)</th>
<th>δ18O (Sr/Ca)</th>
<th>δ34Ca (%)</th>
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<td>Anh-3</td>
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The relative standard deviation (2σ) for the concentration of the individual dissolved components by ICP-OES and ion chromatography is ±2% for Na+, ±1% for Ca and Sr, and ±1% for Cl− and SO42−. The solution mass reported is representative of the mass before each sampling event.

a The Sr/Ca partition coefficient, K_D, between anhydrite and dissolved constituents is calculated by mass balance of the time-series samples solution chemistry.

b The bold time-series solution samples, 1-5, 2-6, and 3-6, represent the initial sample taken after the 4Ca spike injection, where the “time-zero” δ34Ca isotope composition is calculated by linear extrapolation of the two following samples post-injection.

c The abbreviate, “n.a.,” stands for “not analyzed” with respect to the Ca isotope composition of the solution sample. The Ca isotope samples are shown with the 2σ relative standard deviation. These specific samples were not analyzed as a result of low Ca/Na ratios due to mixing effects immediately after the Na2SO4 solution injection into the gold cell reactor solution.

d The directly measured K_D between product anhydrite and the final solution sample.
Table 2

$^{43}$Ca isotope exchange and rate systematics derived from the anhydrite-fluid exchange experiments.

<table>
<thead>
<tr>
<th>Sample</th>
<th>$^{43}$Ca (system)</th>
<th>$^{43}$Ca (Fluid)</th>
<th>$^{43}$Ca</th>
<th>2σ</th>
<th>$X_{\text{Ca(Fluid)}}$</th>
<th>$\delta^{43}$Ca</th>
<th>$\delta^{43}$Ca (system)</th>
<th>$\delta^{43}$Ca (Fluid)</th>
<th>$r$</th>
<th>$F$ (%)</th>
<th>$h_{(t(F)}}$</th>
<th>$K_{\text{D(Equil.)}} \left( \text{Sr/Ca} \right)$</th>
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- The first spike sample, normalized at time zero, was taken approximately a half hour after the injection of the $^{43}$Ca solution spike into the experimental solution. The concentration of dissolved Ca of the $^{43}$Ca isotope spike solution is 0.54 mmol/kg and has a $^{43}$Ca/$^{40}$Ca ratio of 0.35 whereas the natural $^{43}$Ca/$^{40}$Ca ratio is ~0.001.
- The measured composition of product anhydrite recovered upon termination of the experiment. The mass of $^{43}$Ca solution spike added to the gold cell reactor solution was approximately 1.5–2.5 g for each experiment. Before and after the spike injection, 1 g of 550 mmol/kg NaCl solution was added to effectively flush the Ti-capillary tubing free of experimental solution and $^{43}$Ca spike, respectively.
- The directly measured composition of the system (anhydrite + dissolved Ca) at the termination of the experiment.
- $F(\%)$ is the fractional degree of Ca isotope exchange. The degree of exchange, $F$, for the final sample is determined by measuring the complete isotopic mass balance of the system, i.e. product anhydrite and the final solution sample.
- The rate of Ca isotope exchange between the anhydrite and fluid, $k(F)$, is determined by the best fit of a 1st order exchange model, where the units are in s$^{-1}$.
- Equilibrium estimate of the Sr/Ca partitioning between anhydrite and fluid, $K_{D(Equil.)}$, by use of the rate constant data from each experiment coupled with time-series changes in $K_D$ (see text).
time-zero, \( t = 0 \), \( \delta^{44}\text{Ca} \) composition for the exchange rate analysis was estimated by determination of the y-intercept by linear regression of the first three samples immediately after the injection of the \( ^{43}\text{Ca} \) spike solution into the reactor. Subsequently, time-series solution samples were taken to gauge the extent of exchange between anhydrite and dissolved \( \text{Ca}^{2+} \) that occurred upon recrystallization throughout the remainder of each experiment.

### 2.2. Major/minor element analyses

The time-series solution samples were retrieved in a syringe attached to a high pressure metering valve, where approximately 2–3 g of solution were allocated between bottles and diluted appropriately with de-ionized H\(_2\)O for anions and cations and for \( \text{Ca} \) isotope analysis. Prior to recovery of the solution sample used for chemical and isotopic analysis, a rinse sample (1 g) was taken to effectively flush the sampling tubing of residual fluid not indicative of the solution chemistry of the reactor solution at each time step. After completion of each experiment, the gold cell reactor assembly was quenched, within 30 min, to ambient temperature and pressure conditions and product anhydrite was immediately recovered, filtered, washed in de-ionized water, and allocated into bottles for cation/anion and \( \text{Ca} \) isotope analysis. The chemical components of sample solutions and product anhydrite were analyzed by ICP-OES and IC to distinguish from published values determined by both MC-ICP-MS follow previously published methods (Balcaen et al., 2005). The accuracy and precision of our analyses is determined by repeated chromatography and mass spectrometry of modern seawater (\( ^{187}\text{Sr}^{16}\text{Sr} = 0.70918 \), Faure and Mensing (2005)). Measured \( ^{87}\text{Sr}^{16}\text{Sr} \) ratios of 8 seawater standards are 0.709188 ± 27 ppm (2\( \sigma \)). All \( \text{Sr} \) isotope data were normalized to the standard, NBS SRM987 \( ^{87}\text{Sr}^{16}\text{Sr} = 0.710248 \), analyzed during the same analytical session.

Long-term external reproducibility for each isotopic system is determined based on the difference between two known standards taken through the full chemical procedure (column chromatography and mass spectrometry) with each batch of samples. For \( \text{Ca} \) isotopes we report external reproducibility using the measured value of SRM915b relative to modern seawater, both of which are processed and analyzed together with a set of 20–30 samples on the same IC run. Our measured \( ^{44}\text{Ca} \) value for SRM915b relative to modern seawater is \(-1.18 \pm 0.16\%e \) (2\( \sigma \), n = 125), indistinguishable from published values determined by both MC-ICP-MS and TIMS (Fantle and Tipper, 2014; Feng et al., 2017). All \( \text{Ca} \) isotope samples are reported relative to modern seawater \( ^{44}\text{Ca}_{\text{seawater}} = 0\%e \). \( ^{43}\text{Ca}_{\text{seawater}} = +1.92\%e \) on the SRM915a scale and +0.98\%e on the bulk silicate Earth (BSE) scale (Fantle and Tipper, 2014). Reported errors for each sample depend on the number of times the sample has been separated and analyzed. For a single separation and analysis, we report the long-term external reproducibility of SRM915b or Cambridge-1 \( (2\sigma = \pm 0.16\%e, \pm 0.09\%e \) respectively). For multiple separations and analysis (N > 1) we also report the standard error of the mean (SE).

### 2.3. \( \text{Ca} \) and \( \text{Sr} \) isotope analyses

The \( \text{Ca} \) and \( \text{Sr} \) isotope composition of naturally and experimentally derived solution and anhydrite samples were processed and analyzed by a Neptune Plus MC-ICP-MS at Princeton University. The anhydrite was initially digested in warm 0.5 N H\(_2\)O\(_3\) and evaporated to dryness and reconstituted in 2–3 mL of 0.5 N H\(_2\)O\(_3\). Samples for \( \text{Ca} \) and \( \text{Sr} \) isotope analyses were processed using an automated high-pressure ion chromatography (IC) system (Dionex ICS-5000+) following previously published methods (Blattler et al., 2015; Husson et al., 2016). Dissolved samples were analyzed for \( \delta^{44}\text{Ca} \) and \( \delta^{44}\text{Ca} \) values on a Thermo Scientific Neptune Plus MC-ICP-MS (Blattler and Higgins, 2014; Fantle and Higgins, 2014; Blattler et al., 2015; Husson et al., 2015; Gothmann et al., 2016). Measurements are carried out in medium resolution for \( \text{Ca} \) to avoid \( \text{ArH}^+ \) interferences. All \( \text{Ca} \) isotope data are reported in delta notation relative to a known standard for \( \text{Ca} \) isotopes, seawater, and measured \( \delta^{44}\text{Ca} \) values are converted to \( \delta^{44}\text{Ca} \) values assuming mass-dependent fractionation with a slope of 2.05 and no excess radiogenic \( ^{40}\text{Ca} \) (Fantle and Tipper, 2014):

\[
\delta^{\text{Ca}} = \left[ \frac{R^{44}}{R^{40}} - 1 \right] (\%e) \quad \text{where } x = 43 \text{ or } 44
\]

Although the \( \text{Sr}^{2+} \) is separated from \( \text{Ca} \) during ion chromatography, we correct for occasional small \( \text{Sr}^{2+} \) isobaric interferences in the \( \text{Ca} \) measurements using measurements at m/z = 43.5 (doubly-charged \( ^{87}\text{Sr}^{2+} \)). All \( \text{Ca} \) isotope values are compared in three-isotope space \( (\delta^{44}\text{Ca} \) vs. \( \delta^{43}\text{Ca} \)) to verify that the \( \text{Ca} \) isotope variability falls along the expected mass-dependent line (with the exception of samples containing the \( ^{43}\text{Ca} \) spike).

\( \text{Sr} \) isotope analyses of dissolved anhydrite were carried out at Princeton University on a Neptune MC-ICP-MS. \( \text{Sr} \) was separated from other matrix cations using a Dionex ICS-5000 + high-pressure IC system as described above. \( \text{Sr} \) isotope analyses by MC-ICP-MS follow previously published methods (Balcaen et al., 2005). The accuracy and precision of our analyses is determined by repeated chromatography and mass spectrometry of modern seawater (\( ^{187}\text{Sr}^{16}\text{Sr} = 0.70918 \), Faure and Mensing (2005)). Measured \( ^{87}\text{Sr}^{16}\text{Sr} \) ratios of 8 seawater standards are 0.709188 ± 27 ppm (2\( \sigma \)). All \( \text{Sr} \) isotope data were normalized to the standard, NBS SRM987 \( ^{87}\text{Sr}^{16}\text{Sr} = 0.710248 \), analyzed during the same analytical session.

### 2.4. \( \text{Sr}/\text{Ca} \) partitioning in batch experiments

The \( \text{Ca} \) isotope data of anhydrite and dissolved \( \text{Ca} \), together with time-series changes of dissolved \( \text{Sr}^{2+} \) can be used to determine if the observed \( \text{Sr}/\text{Ca} \) partition constant, \( K_{\text{SR}} \), is indicative of equilibrium, or if a steady-state has been achieved between dissolved \( \text{Ca} \) and \( \text{Sr} \) species and the bulk composition of coexisting anhydrite. At each experimental temperature, \( K_{\text{SR}} \) is calculated as followed:

\[
K_{\text{SR}}(\text{Anh}/\text{Fluid}) = \left[ \frac{[\text{Sr}]}{[\text{Ca}]}_{\text{Anh}} \right] \left/ \left[ \frac{[\text{Sr}]}{[\text{Ca}]}_{\text{Fluid}} \right] \right.
\]

where \( ([\text{Sr}]/[\text{Ca}])_{\text{Fluid}} \) is the ratio of the measured concentration of dissolved \( \text{Sr} \) and \( \text{Ca} \) whereas \( ([\text{Sr}]/[\text{Ca}])_{\text{Anh}} \) is the mole fraction ratio of \( \text{Sr} \) and \( \text{Ca} \) in anhydrite determined by mass balance by comparison of the experimental system solution chemistry relative to the time-series solution samples. Calculated activities of the solution species and for the dissolved components of anhydrite were not used for
this calculation to remain comparable with past experimental and natural studies.

At the termination of each experiment, the Sr/Ca concentration ratio and Ca isotope composition of the product anhydrite was compared with the final time-series solution sample to confirm that the experimental system fulfills elemental and isotopic mass balance constraints. Finally, the experimental anhydrite product was analyzed by scanning electron microscopy (SEM) to characterize morphology after precipitation and recrystallization.

2.5. Ca isotope fractionation upon anhydrite precipitation

The fractionation of Ca isotopes between anhydrite and dissolved Ca\(^{2+}\) is reported as:

\[
\Delta C_{\text{Anh-Ca}^{2+}} = \delta^{43}C_{\text{Anh}} - \delta^{43}C_{\text{aq}} (3)
\]

where \(x = 43\) or 44. The Ca isotope composition of anhydrite is determined through mass balance constraints imposed by the measured composition of the experimental components taking account of mass loss during fluid samples:

\[
\delta^{43}C_{\text{System}} = X_{\text{Anh}} \delta^{43}C_{\text{Anh}} + X_{\text{aq}} \delta^{43}C_{\text{aq}} (4)
\]

where \(X\) is the mole fraction of Ca relative to each respective phase, anhydrite and dissolved Ca\(^{2+}\). In addition, the fractionation between anhydrite and dissolved Ca\(^{2+}\) produced by rapid initial precipitation of anhydrite from solution is calculated through linear regression analysis using a Rayleigh fractionation model:

\[
\ln(1000 + \delta^{43}C_{\text{aq}})) / (1000 + \delta^{43}C_{\text{eq}}) = \ln(f) (t_{\text{Anh-Ca}} - C) - 1 (5)
\]

where \(f\) is the fraction of Ca\(^{2+}\) remaining in solution, \(\delta^{43}C_{\text{aq}}\) is the isotopic composition of the solution at each stage of \(f\), and \(t_{\text{Anh-Ca}}\) is the associated fractionation produced between anhydrite and dissolved Ca\(^{2+}\) upon precipitation at each experimental condition. The Rayleigh fractionation model is appropriate to use during the precipitation event, since the rate of precipitation is rapid, and subsequent exchange between incipiently formed anhydrite and dissolved Ca\(^{2+}\) is limited.

2.6. Quantification of the extent and rate of anhydrite recrystallization using a \(^{43}\text{Ca}\) spike

To quantify the extent of Ca isotope mixing, \(F\), between anhydrite and coexisting dissolved constituents, we use the time-series solution Ca isotope data, \(\delta^{43}\text{Ca}\), the time-zero isotopic composition immediately after the \(^{43}\text{Ca}\) isotope injection, \(\delta^{43}_{0}\), and the isotopic composition of the entire system (i.e. anhydrite + dissolved Ca\(^{2+}\)) measured at the termination of each experiment, \(\delta^{43}_{\text{Anh-Ca}}\), as follows:

\[
F = \frac{\delta^{43}_{\text{Anh-Ca}} - \delta^{43}_{\text{Ca}}}{\delta^{43}_{0} - \delta^{43}_{\text{Ca}}} (6)
\]

Utilization of the bulk system Ca isotope composition, \(\delta^{43}_{\text{Anh-Ca}}\), for the calculation of \(F\) is justified since the introduction of the \(^{44}\text{Ca}\) spike into the reactor solution results in a highly anomalous initial \(\delta^{44}\text{Ca}\) composition (>100\%), significantly larger in magnitude than the equilibrium fractionation expected upon complete exchange between anhydrite and dissolved Ca\(^{2+}\) (Johnson et al., 2004). The \(\delta^{43}_{\text{Ca}}\) determined at the end of the experiment was used since the Ca isotope composition of the system evolves upon sampling and removal of solution throughout reaction progress. In addition, mass-balance constraints between the Ca reservoirs, anhydrite and fluid, are utilized to quantify the relative amount of anhydrite recrystallized with time for each experiment, as described further by Handler et al. (2014):

\[
F = \frac{m_{\text{Soln}} \delta^{43}\text{Ca} - \delta^{43}_{\text{Anh}}} {m^{43}_{\text{Anh}} \delta^{43}\text{Ca} - \delta^{43}_{\text{Anh}}} (7)
\]

Specifically, the calculation of \(F\) explicitly takes into account the relative mass balance of Ca in anhydrite and dissolved Ca reservoirs, where \(m^{43}_{\text{Anh}}\) and \(m^{43}_{\text{Anh}}\), represent the total moles of Ca in solution and anhydrite, respectively.

Linear regression of the \(\delta^{43}\text{Ca}\) fractional exchange, \(F\), with time, \(t\), is used to estimate the rate constant, \(k_{\text{Anh}}\) (s\(^{-1}\)), for Ca isotope exchange between anhydrite and dissolved Ca during recrystallization at each temperature. The time dependent isootope exchange data are applied to a first-order rate law (Cole et al., 1983, 1987, Criss et al., 1987; Gregory et al., 1989; Criss, 1999; Johnson et al., 2002):

\[
\ln(1 - F) = -k_{\text{Anh}} t (8)
\]

The first-order rate law was utilized since it provided a reasonable fit of the observed time-series \(\delta^{43}\text{Ca}\) solution data and is comparable to previous experimentally determined exchange rate data based on \(^{84}\text{O}\) isotope systematics between anhydrite and water (Chiba et al., 1981). The temperature dependence of the rate constant for Ca isotope exchange, \(k_{\text{Anh}}\), is quantified by determination of the activation energy, \(E_A\), where \(T\) is in Kelvin, \(R\) is the gas constant, and \(ln(\Delta)\) is the pre-exponential factor which describes the frequency of molecular collisions:

\[
\ln(k_{\text{Anh}}) = (-E_A / R) * (10^{3} / T) + \ln(\Delta') (9)
\]

With this specific Ca isotope exchange rate data, the degree and mechanism of exchange can be estimated for a range of hydrothermal conditions indicative of subseaflow hydrothermal systems, providing information fundamental to the interpretation of the Ca and Sr elemental/isotopic composition of natural hydrothermal fluids and anhydrite.

2.7. Geochemical modeling and speciation calculations

The saturation state of the batch experiments with respect to anhydrite was determined using the thermodynamic modeling software, Geochemist’s Workbench (Bethke, 2006). Model inputs include measured concentrations of the dissolved components, Ca\(^{2+}\), Sr\(^{2+}\), Na\(^{+}\), Cl\(^-\), SO\(_{4}^{2-}\) and pH at 25°C. A modified equilibrium database, required for the thermodynamic model to determine the
speciation of aqueous species was created by the DBCreate program developed by Kong et al. (2013) for the range of temperatures and pressure considered in this study, 25–400 °C and 500 bars. The algorithms required for the derivation of the thermodynamic data are consistent with Helgeson-Kirkham-Flowers equation of state and complementary revisions (Helgeson et al., 1981; Tanger and Helgeson, 1988; Shock et al., 1989, 1992, 1997; Sverjensky et al., 1997). The thermodynamic model assumes the activity of water and anhydrite is unity and the associated activity coefficients for charged dissolved species are approximated by the extended Debye-Hückel equation. The assumption of unity for the activity of water is justified since explicit calculation of the activity of water at experimental P-T-X conditions indicate a range in activity between 0.98 and 0.99 (Archer, 1992). The thermodynamic model takes advantage of the appropriate mass balance, mass action, and charge balance constraints coupled with the measured composition of the time-series solution samples to determine the distribution and speciation of dissolved constituents and mineral stability at experimental conditions (Fig. 3).

2.8. Natural analogue - TAG hydrothermal system

The TAG hydrothermal system, located at 26°08’N along a slow-spreading segment of the Mid-Atlantic Ridge, is a massive sulfide deposit that has been episodically active for over 20,000 years (You and Bickle, 1998). Since the discovery of high temperature vents at TAG in 1985, it has served as an important site for better understanding the geochemical, geophysical, and biological characteristics of active massive sulfide deposits at and below the seafloor of MOR hydrothermal systems (Rona et al., 1986; Humphris et al., 2015). TAG is particularly relevant to this study as drill cores from the massive sulfide mound of this site, recovered from ODP Leg 158, contain abundant anhydrite and there is extensive geochemical evidence that conductive heating and mixing of seawater with high temperature hydrothermal fluids play a key role controlling anhydrite formation (e.g. SF:HF mixing zone of Fig. 1) (Humphris et al., 1995; Herzig et al., 1998; Mills et al., 1998; Teagle et al., 1998a). This is confirmed by anhydrite fluid inclusion data and the intermediate Sr isotope composition of anhydrite relative to MORB and seawater (Petersen et al., 1998; Tivey et al., 1998). In addition, high temperature vent fluids (>350 °C) at TAG have recently been sampled, adding chemical and isotopic data with which the experimental data reported here can be compared.

The anhydrite samples collected from the TAG hydrothermal vent system are taken from a core drilled on a sulfide mound on IODP Leg 158 and contains sulfide/anhydrite stock-work as described by Humphris et al. (1995). The anhydrite collected for Ca isotope analysis is from the drill core half, 158-957C-7N-2, recovered from the depth interval of 20.25–20.31 m below the seafloor (Table 3). Anhydrite sampled from the same approximate depth and interval of the core half, 158-957C-7N-1, which has previously been analyzed for δ18O, δ34S, 87Sr/86Sr, and fluid inclusion homogenization temperatures, are used for comparison with this study (Chiba et al., 1998; Mills et al., 1998; Teagle et al., 1998a; Tivey et al., 1998; Mills and Tivey, 1999).

The high temperature vent fluids at TAG were retrieved by use of the remotely operated vehicle, JASON II, during a scientific expedition in 2008 aboard R/V Roger Revelle (KNOX189RR). The procedure for sampling the high-temperature solution was facilitated by use of a gas-tight Ti-sampler system, which is aided by a thermocouple to guide the sampler snorkel into the highest temperature fluid issuing from the sulfide mound, minimizing mixing of seawater with the hydrothermal fluid during sampling (Seewald et al., 2001; Wu et al., 2011). The sampled hydrothermal fluid is subsequently brought aboard ship, typically within a few hours, and allocated into acid-washed polyethylene bottles for cation, anion, and isotopic analysis (Table 3). An aliquot of the sampled hydrothermal solution was measured for pH by use of a Ross micro-electrode, which was calibrated with pH buffers of 4, 7, and 10. The solution samples for cations and anions were diluted with 0.1 N ultra-trace metal grade HCl and deionized water, respectively, and preserved for analysis by IC and ICP-OES at the University of Minnesota. The uncertainties associated with the chemical and isotopic analysis of the hydrothermal fluids are analogous to those described for the measurements of the experimental samples.
Table 3

Hydrothermal fluid samples and anhydrite sampled from the TAG hydrothermal vent field along the Mid-Atlantic Ridge.

<table>
<thead>
<tr>
<th>TAG fluid samples</th>
<th>Temp. (°C)</th>
<th>pH_{25°C}</th>
<th>[Mg^{2+}] (mM)</th>
<th>[Ca^{2+}] (mM)</th>
<th>[Sr^{2+}] (µM)</th>
<th>[Na^{+}] (mM)</th>
<th>[Cl(^-)] (mM)</th>
<th>87Sr/86Sr</th>
<th>δ^{44}Ca (‰)</th>
<th>δ^{44}Ca (‰)</th>
<th>2σ (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seawater</td>
<td>2</td>
<td>8</td>
<td>53.3</td>
<td>10.4</td>
<td>86</td>
<td>468</td>
<td>546</td>
<td>0.70918</td>
<td>0</td>
<td>0.05</td>
<td></td>
</tr>
<tr>
<td>Kremlin (363-IGT4)</td>
<td>140</td>
<td>5.83</td>
<td>32.3</td>
<td>22.8</td>
<td>101</td>
<td>510</td>
<td>592</td>
<td>n.a.</td>
<td>−0.74</td>
<td>−0.99</td>
<td>0.05</td>
</tr>
<tr>
<td>Complex (363-CGT-R)</td>
<td>363</td>
<td>4.09</td>
<td>5.9</td>
<td>35.3</td>
<td>112</td>
<td>555</td>
<td>649</td>
<td>n.a.</td>
<td>−0.80</td>
<td>−1.02</td>
<td>0.04</td>
</tr>
<tr>
<td>Complex (363-IGT5)</td>
<td>363</td>
<td>3.28</td>
<td>3.9</td>
<td>39.4</td>
<td>126</td>
<td>552</td>
<td>646</td>
<td>n.a.</td>
<td>−0.91</td>
<td>−1.21</td>
<td>0.07</td>
</tr>
<tr>
<td>b,c End-member</td>
<td>–</td>
<td>–</td>
<td>0</td>
<td>40.2</td>
<td>123</td>
<td>563</td>
<td>658</td>
<td>0.70380</td>
<td>−0.90</td>
<td>−1.19</td>
<td>0.12</td>
</tr>
<tr>
<td>TAG anhydrite</td>
<td>Depth (mbsf)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>158-957C-7N-2-25 cm</td>
<td>20.25</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.70166</td>
<td>−0.81</td>
<td>−1.03</td>
<td>0.02</td>
</tr>
<tr>
<td>158-957C-7N-2-31 cm</td>
<td>20.31</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.70208</td>
<td>−0.86</td>
<td>−1.12</td>
<td>0.04</td>
</tr>
<tr>
<td>Anhydrite-fluid</td>
<td>'δ^{44}Ca_{HF/SW} (%)'</td>
<td>2σ (%)</td>
<td>'δ^{44}Ca_{Anh/Fluid} (%)'</td>
<td>2σ (%)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>158-957C-7N-2-25 cm</td>
<td>−0.71</td>
<td>0.10</td>
<td>−0.32</td>
<td>0.10</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>158-957C-7N-2-31 cm</td>
<td>−0.71</td>
<td>0.10</td>
<td>−0.42</td>
<td>0.10</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>

All concentrations are represented in mmol/kg solution. The relative standard deviation (2σ) for the concentration of the individual dissolved components by ICP-OES and ion chromatography is ±2% for Na\(^+\), ±1% for Ca and Sr and ±1% for Cl\(^-\) and SO\(_4\)^{2\text{-}}\).

a The radiogenic Sr data are derived from Edmond et al. (1995) with an average 2σ error associated of ±0.000010. The solution mass reported is representative of the mass before each sampling event.

b The end-member hydrothermal composition of the fluid sampled from TAG is determined from linear regression of Mg/Ca versus the δ^{44}Ca of seawater and sampled high-temperature fluid.

c The end-member Ca isotope composition of the hydrothermal fluid sampled from TAG is determined by linear regression of Mg/Ca versus the δ^{44}Ca of seawater and sampled high-temperature fluid.

d The measured δ^{87}Sr/δ^{86}Sr ratio and fluid inclusion homogenization temperatures for sampled TAG anhydrite in this study is comparable in composition from the same section and depth of core previously reported by Mills and Tivey (1999) and Teagle et al. (1998a,b), 158-957C-7N-1.

e The fractionation between anhydrite and fluid, δ^{44}Ca_{Anh/Fluid}, is calculated from the Ca isotope composition of sampled anhydrite and the seawater/hydrothermal fluid mixture, from which it formed from, where the seawater/hydrothermal fluid Ca isotope composition was calculated through conservative mixing relationships, based on the 87Sr/86Sr and Ca isotope composition of anhydrite relative to seawater and hydrothermal fluid.
3. RESULTS

3.1. Experimental precipitation and recrystallization of anhydrite at 175, 250, and 350 °C

3.1.1. Major element chemistry and petrography of synthesized anhydrite

Over the first hours of the experiment, dissolved Ca and Sr concentrations decline by approximately 30–40% due to the formation of anhydrite as a result of the addition of the Na2SO4 solution into the gold cell reactor (Table 1). Dissolved SO4 concentrations in each experiment increase over the same time interval from effectively zero to peak values during the initial mixing stage of the reaction, 3.83–27.72 mmol/kg. Subsequently, SO4 concentrations decline, stabilizing at concentrations ranging from approximately 0.2 to 4 mmol/kg for the duration of the experiments. Dissolved Ca concentrations exhibit more complex behavior owing to the addition of the Na2SO4 and subsequent temperature dependent mixing and precipitation reactions. For example, dissolved Ca slowly declines at both 175 and 250 °C, but increases in the 350 °C experiment; however, all experiments eventually stabilize to values between approximately 7–10 mmol/kg Ca2+. Thermodynamic speciation calculations for all time-series solution samples after the precipitation event are at saturation with respect to anhydrite solubility at each prescribed pressure and temperature (Fig. 3). For the entirety of each experiment, all sampled solutions had a measured pH at 25 °C between 5.26 and 5.84, whereas the range in pH at experimental conditions is between 5.30 and 5.50. In addition, all speciation calculation of the time-series solution samples from each experiment confirm that Sr-bearing minerals, such as strontianite (SrCO3) and celestite (SrSO4), are undersaturated.

SEM imaging indicates that product anhydrite at 350 °C has an acicular morphology, indicative of fast precipitation rates, as demonstrated by a previous experimental examination of anhydrite precipitation at hydrothermal conditions by Shikazono and Holland (1983) (Fig. 4). The range in grain size for product anhydrite, precipitated and recrystallized at 350 °C, is approximately between 300 and 10 μm. Furthermore, the large grain size distribution, suggests that the mechanism of recrystallization is a favorable process that effectively facilitates anhydrite growth and exchange between the mineral and fluid elemental and isotopic reservoirs (Dubinina and Lakshtanov, 1997). Although not shown here, the morphology of anhydrite from experiments at 175 and 250 °C is also acicular in morphology with a similar grain size distribution. The anhydrite crystals were not tested for chemical heterogeneity, and thus, are assumed to be unzoned.

3.1.2. Sr/Ca partitioning during precipitation and recrystallization of anhydrite

The Sr/Ca partition coefficient for anhydrite-fluid (K_D), calculated for each sample indicates that the initial precipitation event is predominantly associated with elevated K_D values, 1.24, 0.97, 0.73, for 175, 250, and 350 °C, respectively, relative to long-term K_D values (Table 1 and Fig. 5). Time evolution paths of K_D and the rate of approach to steady state, however, differ for each temperature as anhydrite undergoes recrystallization. For example, at 175 °C, K_D data reaches a steady state value of 0.68, 95 h after the precipitation event and little change occurs subsequently in the course of 1991 h. However, at higher temperatures, 250 and 350 °C, the time-series K_D data continue to evolve towards lower values, ultimately 0.56 and 0.55, throughout the course of each experiment, 1192 and 436

Fig. 4. SEM image of product anhydrite precipitated and recrystallized at 350 °C throughout the course of the 366 h experiment. The product anhydrite exhibits crystalline acicular morphology and a range in grain size, suggesting that the mechanism of recrystallization was likely the dominant process that facilitated exchange between the mineral and fluid elemental and isotopic reservoirs. The acicular morphology and similar size distribution of product anhydrite is also demonstrated for the lower temperature experiments at 175 and 250 °C (not shown).

Fig. 5. Evolution of Sr/Ca partitioning, K_D, between anhydrite and dissolved Ca and Sr with reaction progress. Results based on mass calculations from solution chemistry (open symbols) and from the terminal measured compositions of anhydrite and dissolved Ca and Sr at the end of each experiment (black symbols). The calculated K_D values from the solution chemistry of early samples tend to exhibit relatively large range of values, approximately between 0.4 and 1.3 (see text). With increasing extents of anhydrite recrystallization with time the K_D values, from the perspective of solution chemistry, converge towards 0.65, 0.54, and 0.55 whereas the directly measured K_D from product anhydrite and the final solution sample have similar values, 0.66, 0.55, 0.52 for 175, 250, and 350 °C, respectively. The close agreement between the calculated and measured terminal K_D values demonstrates that the experimental system fulfills elemental mass balance constraints.
h, respectively. At the end of each experiment, the measured Sr/Ca ratio of product anhydrite was compared with the terminal solution sample, resulting in $K_D$ of 0.66, 0.55, and 0.52 for 175, 250, and 350 °C, respectively, similar to that calculated from solution chemistry, confirming that mass balance was maintained throughout each experiment.

3.1.3. Fractionation of natural abundance Ca isotopes during anhydrite precipitation

For each experiment, prior to anhydrite precipitation, the natural Ca isotope composition of the initial solution sample was measured to constrain the initial Ca concentration and isotope composition of the experimental system and isotope composition of the experimental system sample was measured to constrain the initial Ca concentration and isotope composition of the initial solution sample. Upon precipitation of anhydrite, dissolved Ca declines but becomes concentrated in heavy isotopes due to Rayleigh-type distillation. By utilizing a Rayleigh fractionation model with the time-series solution chemistry and associated Ca isotope data (Eq. (5)), precipitation induced fractionation yields values of −0.45, −0.22, and −0.02‰, for 175, 250, and 350 °C, respectively (Fig. 6). These data demonstrate a clear temperature dependent fractionation trend, which is consistent with natural anhydrite-fluid fractionation data reported for the Logatchev hydrothermal system and from past low temperature experimental studies.

3.1.4. Rates of anhydrite-dissolved Ca exchange at 175, 250, and 350 °C

After sufficient time was allowed for anhydrite to recrystallize and for the solution chemistry to reach chemical saturation conditions, an anomalous $^{43}$Ca solution was introduced to quantify the rate of exchange at each temperature. The time of injection is normalized as $t = 0$. The time-zero dissolved $^{43}$Ca composition sampled following injection was determined from linear extrapolation of the first three anomalous $^{43}$Ca samples, resulting in 180, 380, and 290‰ for the experiments at 175, 250, and 350 °C, respectively (Tables 1 and 2). This extrapolation procedure was conducted to prevent sampling of mixing effects effectively after the injection of the anomalous $^{43}$Ca solution into the gold cell reactor. As expected, with increasing recrystallization, the anomalous $^{43}$Ca tracer decreases. The fraction of isotopic mixing between anhydrite and dissolved Ca$^{2+}$, $F$ (calculated from the time-series Ca isotope composition) indicates that 13, 46, and 45% of the enriched $^{43}$Ca tracer in solution exchanged with anhydrite at 175, 250, and 350 °C in the course of 1322, 867, and 366 h, respectively (Eq. (6), Table 2, and Figs. 7 and 8). Comparison of the calculated initial Ca isotope composition of

anhydrite, prior to the spike, with that of the isotopically enriched anhydrite at the end of the experiment indicates that 4, 38, and 49% mixing occurred at 175, 250, and 350 °C, respectively. The difference between the calculated degree of mixing, \( F \), from the perspective of either Ca isotope reservoir, dissolved Ca\(^{2+} \) versus anhydrite, likely results from partial dissolution of product anhydrite during quench towards ambient conditions at the end of the experiment, since anhydrite becomes increasingly more soluble at lower temperature. As the reactor system allows sampling of the solution at experimental pressure and temperature conditions, estimated values of \( F \) from the time-series solution data are likely more accurate. The fraction of moles of Ca derived from anhydrite, \( F^* \) (Eq. (7)), to effectively cause the observed change in the time-series \(^{43}\text{Ca} \) solution data, \( F \) (Eq. (6)), requires approximately 6, 18, and 15% of Ca exchanged from anhydrite upon recrystallization at 175, 250, and 350 °C throughout the course of 1322, 867, and 366 h, respectively.

Measured rates of chemical exchange between anhydrite and fluid are a strong function of temperature and similar in magnitude to those observed in previous experimental studies utilizing \(^{18}\text{O} \) spikes to quantify the equilibration rate between anhydrite derived SO\(_4^{2-} \) and H\(_2\)O at hydrothermal conditions (Figs. 7 and 8) (Chiba et al., 1981). Linear regression of the rate constants derived from the Ca isotope exchange rate data, \( F \), between anhydrite and dissolved Ca\(^{2+} \) at each experimental temperature (Table 2 and Fig. 7), yield an activation energy of 46 kJ/mol, a value that is also similar to that determined by Chiba et al. (1981). This range in magnitude of E\(_A\) is indicative of exchange by dissolution and precipitation, where diffusion in anhydrite is likely the rate limiting effect on complete equilibration between anhydrite and dissolved constituents (Cole et al., 1983, 1987). In addition, Chiba et al. (1981) also noted that solution chemistry may impart a control on the rate of exchange, where fluids with a high H\(^+ \) activity, such as 1 M H\(_2\)SO\(_4 \) and 1 M HCl experimental solutions shown in Fig. 8, exhibit a lower activation energy, 29 kJ/mol, and thus, faster rates of exchange relative to experiments implementing circum-neutral fluids in NaCl solutions, 0.5 and 1.0 M, 42–46 kJ/mol. Furthermore, the rate data from experiments utilizing 0.5–1.0 M solutions, despite the small difference, suggests that an increase in the activity of Cl\(^- \) in solution, which effectively enhances the solubility of anhydrite by increasing the abundance of dissolved Ca–Cl complexes, may also increase the rate of elemental and isotopic exchange between anhydrite and the dissolved constituents in solution.

### 3.1.5. Equilibrium \( K_D \) values between anhydrite-fluid

To differentiate between equilibrium and kinetic \( K_D \) values, the rate constants derived from the \(^{43}\text{Ca} \) exchange data were used. This allowed the determination of the equilibrium \( K_D \) value in spite of incomplete exchange, by linear regression of the entire dataset of time-series \( K_D \) values against the predicted degree of mixing with time, \( F \) (Fig. 9). This approach, however, does not use the initial \( K_D \) values produced during the rapid anhydrite precipitation event, which are indicative of kinetic and mixing effects. Equilibrium \( K_D \) values, as determined from the linear regression of the time series data at each temperature, indicate a slight, but significant temperature effect, as follows: 0.43 ± 0.12, 0.34 ± 0.05, 0.29 ± 0.02 (1σ), for 175, 250, and 350 °C, respectively. Despite the large error associated with 175 °C experiment, due to the limited amount of exchange between anhydrite and fluid throughout the experiment, the equilibrium \( K_D \) values throughout the range of temperatures are of the same relative magnitude with the \( K_D \) values demonstrated from near equilibrium anhydrite precipitation experiments at 150–250 °C, 0.31–0.21, by Shikazono and Holland (1983), and estimated by comparison of natural anhydrite and hydrothermal fluids from a range of MOR hydrothermal vent sites, 0.60 ± 0.28, as determined by Teagle et al. (1998a). This approach of estimating equilibrium partitioning values by utilization of exchange rate data for Ca and extrapolation to complete exchange is valid since Sr is intimately associated with Ca and that SO\(_4^{2-} \) and Ca exchange congruently, as evidenced by the similar time scales at which these components exchange with anhydrite (Chiba et al., 1981) Fig. 8.

### 3.2. Ca and Sr Chemical and Isotopic Systematics from the TAG hydrothermal vent field

Hydrothermal fluids from TAG were derived from both diffuse and focused-flow structures, having temperatures of approximately 140 and 363 °C, respectively (Table 3). The chemical composition of the diffuse flow fluid is relatively seawater-like, as indicated by the high Mg\(^{2+} \) concentration. As noted above these fluids have undergone conductive
heating prior to seawater mixing. Evidence for this mixing includes elevated concentrations of dissolved Ca\(^{2+}\) and intermediate radiogenic Sr isotope composition between seawater and MORB, which is consistent with elevated homogenization temperatures in the subseafloor as determined from fluid inclusion analysis of anhydrite sampled from this same locality (Edmond et al., 1995; Tivey et al., 1995; Petersen et al., 1998; Teagle et al., 1998a). In contrast, the higher temperature vent fluids are typical of end-member hydrothermal fluids, which have near-zero dissolved Mg\(^{2+}\). Multiple hydrothermal fluid samples were recovered such that the calculation of end-member elemental concentration and Ca isotope compositional data could be determined. Thus, extrapolation of the Ca\(^{2+}\) and Sr\(^{2+}\) compositional data from all of the sampled hydrothermal fluids, together, towards zero Mg\(^{2+}\) indicates that the estimated end-member hydrothermal fluid is composed of 40.2 mmol/kg Ca\(^{2+}\) and 123 \(\mu\)mol/kg Sr\(^{2+}\), respectively. These data are similar to earlier reported data from TAG (Edmond et al., 1995; Rona et al., 1986). Following the same extrapolation procedure and assuming a \(\delta^{44}\)Ca value for seawater of 0‰, end-member \(\delta^{44}\)Ca value for the TAG hydrothermal fluid is estimated to be \(-1.19 \pm 0.12\)‰, approximately 0.0–0.2‰ lower than the measured \(\delta^{44}\)Ca value of the sampled TAG hydrothermal fluid at the seafloor and 0.2‰ lower than average MORB (\(-1.02 \pm 0.18\)‰, Feng et al., 2017).

The \(\delta^{44}\)Ca value of the anhydrite samples are \(-1.03 \pm 0.02\) and \(-1.12 \pm 0.04\)‰, for core sections 158-957C-7N-2-25 cm and 158-957C-7N-2-31 cm, respectively (Table 3). The measured \(^{87}\)Sr/\(^{86}\)Sr composition of anhydrite is 0.707166 and 0.707208, for 158-957C-7N-2-25 cm and 158-957C-7N-2-31 cm, respectively, similar in composition of anhydrite measured from the same depth interval of complementary drill core, 0.707102, by Teagle et al. (1998a).

By pairing of the \(^{87}\)Sr/\(^{86}\)Sr and Ca isotope compositional data of anhydrite, seawater, and end-member hydrothermal fluid composition through conservative mixing relationships, the Ca isotope composition of the chemically evolved fluid (hydrothermal fluid + seawater) from which anhydrite precipitated directly can be estimated. Coupling of these specific data result in an average \(\delta^{44}\)Ca composition of the evolved fluid from TAG, \(-0.71 \pm 0.07\)‰, where the \(\delta^{44}\)Ca fractionation between anhydrite and evolved fluid, \(\Delta^{44}\)Ca\(_{Anh-Ca^{aq}}\), is calculated to be \(-0.32\) and \(-0.42 \pm 0.10\)‰ (2σ), These values are similar in sign and magnitude with past field data from Logatchev, shown by Amini et al. (2008), \(-0.33 \pm 0.16\) and \(-0.62 \pm 0.09\)‰ (2σ), and with the experimental fractionation data reported from this study (Fig. 6).

4. DISCUSSION

4.1. Role of anhydrite in ocean hydrothermal environments and seawater chemistry

Anhydrite plays an important if somewhat enigmatic role in the geochemical cycling of Ca, Sr, and SO\(_4\) in MOR and back-arc hydrothermal systems (Teagle et al., 1998b; Sleep, 1991; Alt, 1995; Alt et al., 2003). The precipitation of anhydrite in and around hydrothermal recharge and discharge zones reduces porosity, permeability, and has a significant effect on the rate/geometry of hydrothermal fluid flow (Lowell et al., 2003). These physical effects engendered from anhydrite precipitation ultimately: (1)
insulate the oceanic crust from influx of large volumes of seawater, (2) focuses heat and flux of high-temperature fluids venting from below the seafloor; and, (3) may induce spatial variability in the extent of high temperature hydrothermal alteration of the oceanic crust (Sleep, 1991; Alt, 1995; Lowell and Yao, 2002). However, the overall mass balance of seawater sulfur predicted to be stored in young oceanic crust as anhydrite and sulfide produced from the reduction of seawater \( \text{SO}_4 \), is much less than that observed, as demonstrated by comparison of mineralogical and chemical data of drill core recovered from the relatively young oceanic drill site, DSDP/ODP Hole 504B (5.9 Ma) with thermal and geochemical estimates (Teagle et al., 1998b). This deficit, known as the anhydrite conundrum, raises the question of what particular zones in the subsurface predominantly sequester seawater Ca and \( \text{SO}_4 \) as anhydrite, for example, along the recharge path, deep-seated at or below the lava-dike transition, or as a consequence of mixing of conductively heated seawater with hydrothermal fluid within and below the massive sulfide mound (Fig. 1). Furthermore, each anhydrite-bearing zone in the subsurface differs in the susceptibility of anhydrite dissolution upon cooling of the hydrothermal system and pervasive seawater mixing, where shallow anhydrite formed along the recharge flow path and in or below the sulfide mound is particularly sensitive to dissolution effects whereas anhydrite formed in the deep-seated zone, at or below the lava-dike transition zone, may persist on geologically significant timescales (Teagle et al., 1998b; Sleep, 1991; Alt, 1995; Alt et al., 2003). The anhydrite samples recovered and used for this study are relevant for comparison with the processes taking place particularly at the zone of seawater/hydrothermal fluid mixing in and below the massive sulfide mound (Fig. 1). In other words, the Ca and Sr isotope composition of anhydrite formed in this zone will be distinct, compositionally, relative to anhydrite formed in the recharge zones and the deep-reaction zone.

The precipitation of anhydrite, in particular in recharge zones of axial MOR hydrothermal systems, will alter the chemical and isotopic composition of fluids penetrating more deeply into the ocean crust (e.g. SW recharge zone of Fig. 1). Anhydrite precipitation will remove seawater-derived Sr, Ca, and \( \text{SO}_4 \); relatively, \( \text{SO}_4 \) will be predominantly removed if the seawater Ca/\( \text{SO}_4 \) > 1, whereas, the majority of Ca and Sr will be removed if seawater Ca/\( \text{SO}_4 \) < 1. Accordingly, the extent to which seawater Ca, \( \text{SO}_4 \), and Sr are removed from solution by anhydrite precipitation will in turn determine the extent to which these elements are available for reaction with the oceanic crust in high-temperature hydrothermal systems and chemical and isotopic fluxes associated with recycling of oceanic crust into the mantle (Bickle and Teagle, 1992; Coogan, 2008). In modern seawater, we expect all of the seawater derived Ca to be removed during anhydrite precipitation, whereas a fraction of seawater Sr is removed, upon recharge into the seafloor and effectively replaced by Ca and Sr derived from the oceanic crust. In contrast, an ancient ocean with a Ca/\( \text{SO}_4 \) > 1 (i.e. the Eocene), seawater derived Ca and Sr may persist beyond the zone of anhydrite precipitation near the seafloor and the partially modified isotopic signature of seawater may be incorporated into secondary Ca silicates during exchange and alteration at high temperature in the reaction zone or possibly retained in composition in the hydrothermal fluid emitting from the subsurface (Lowenstein et al., 2001; Wortmann and Paytan, 2012; Coogan and Dosso, 2015; Antonelli et al., 2017). Although seawater Ca/\( \text{SO}_4 \) ratios likely play an important role in determining the \( ^{87}\text{Sr}/^{86}\text{Sr} \) value of altered oceanic crust, these specific cause and effect processes are not well understood and require experimental and further field data to constrain the Ca isotope composition of altered and subducted oceanic crust.

4.2. Sr/Ca partitioning effects associated with anhydrite precipitation

The mass transfer of Ca and Sr between seawater, hydrothermal fluid, and the oceanic crust is sensitive to anhydrite precipitation and recrystallization in the subsurface (Chiba et al., 1998; Petersen et al., 1998; Teagle et al., 1998a; Tivey et al., 1998; Mills and Tivey, 1999). This is demonstrated from the experimental anhydrite-fluid Sr/Ca partition data presented in this study, where the early stages of incipient anhydrite precipitation predominantly exhibit elevated K\( _D \) values between anhydrite and dissolved constituents; 1.24–0.77 (Fig. 5), relative to the long-term steady state values, 0.66–0.52. The elevated K\( _D \) values demonstrate that the distribution of Sr relative to Ca between the anhydrite-fluid reservoirs is sensitive to the initial formation conditions, in accordance with past experimental data from Shikazono and Holland (1983). Both experimental studies demonstrate that increasingly supersaturated fluids and high rates of anhydrite precipitation result in elevated K\( _D \) values, effectively approaching values of unity, similar to experimental observations made in the calcite-fluid system at wide range of temperatures (Lorenz, 1981; Beck et al., 1992; Lemarchand et al., 2004; Tang et al., 2008; Lopez et al., 2009; DePaolo, 2011). Comparison of time-series K\( _D \) values from this study and K\( _D \) values from prescribed supersaturation experiments performed by Shikazono and Holland (1983) relative to equilibrium K\( _D \) values inferred from both studies, 0.20–0.40, may demonstrate reversibility throughout the range of temperatures indicative of MOR hydrothermal systems.

The observed K\( _D \) values from TAG are likely indicative of kinetic and/or reservoirs effects due to rapid anhydrite precipitation from hydrothermal solution(s), with limited exchange towards equilibrium thereafter, especially at temperatures < 350 °C (Figs. 5 and 10) (Teagle et al., 1998a; Mills and Tivey, 1999). On average, a K\( _D \) value of approximately 0.50–0.60 would represent the mean Sr/Ca partitioning value between anhydrite and hydrothermal fluid/seawater mixtures in the upper subsurface (>50 m below the seafloor). Interestingly, K\( _D \) values from TAG vary with depth, where values exceed unity near the seafloor and have intermediate radiogenic Sr isotope compositions between MORB and seawater. However, anhydrite-fluid K\( _D \) values at >50 m below the seafloor approach values that are consistent with equilibrium conditions and have a mostly radiogenic Sr isotope signature (seawater dominated).
correlation suggests that anhydrite formed along pathways where conductively heated seawater mixed with minor amounts of end-member hydrothermal fluids, precipitates relatively slowly at near equilibrium conditions. Conversely, anhydrite that formed near the seafloor, at and below the sulfide mound, forms rapidly at far from equilibrium conditions due to dynamic mixing of end-member hydrothermal fluids and seawater (Figs. 1 and 10). These particular observations of anhydrite-fluid Sr/Ca partitioning have the potential to differentiate fluid flow pathways at hydrothermal systems.

4.3. Modern MOR Ca isotope fractionation systematics

The Ca isotope composition of the end-member TAG hydrothermal fluid, $-1.19 \pm 0.12$‰, is lighter relative to seawater, 0‰, similar to end-member hydrothermal fluids from the Logatchev and EPR 17–19’S MOR hydrothermal systems, $-0.95 \pm 0.07$‰ and $-0.96 \pm 0.19$‰, respectively, and within error of MORB, $-1.02 \pm 0.18$‰ (Schmitt et al., 2003; Amini et al., 2008; Feng et al., 2017). In comparison, the composition of the anhydrite sampled from TAG ranges from $-1.03$ to $-1.12$‰. The effect of anhydrite precipitation in the subsurface, at and below the sulfide mound, would effectively cause the Ca isotope composition of the mixed fluids (end-member hydrothermal fluid + seawater) to become isotopically heavier (Figs. 1 and 6). Interestingly, the Ca isotope composition of seawater/hydrothermal fluid mixtures, between $-0.77$ and $-0.46$‰, determined from the coupling of the radiogenic Sr and Ca isotope composition of sampled anhydrite from the active sulfide mound of TAG and Logatchev do indicate $\delta^{44}$Ca values greater than MORB, likely as a consequence of precipitation induced fractionation (Table 3 and Amini et al. (2008)). However, as demonstrated by TAG, Logatchev, and EPR 17–19’S end-member vent fluids, the degree of anhydrite formation in the subsurface is not sufficient to significantly shift the Ca isotope composition of end-member fluids relative to MORB. This lack of Ca isotope compositional variability between end-member vent fluids and MORB is very likely due to the significant flux of hydrothermally derived-Ca leached from MORB, effectively overwhelming any reservoir effects caused by anhydrite precipitation in the subsurface (Scheuermann et al., 2018). Overall, a range of $-0.30$ to $-0.60$‰ is appropriate to describe the magnitude of fractionation between anhydrite and fluid in MOR hydrothermal systems.

Primary anhydrite fluid inclusion homogenization temperature data derived from Petersen et al. (1998) from the corresponding core and similar depth interval as the sampled anhydrite used for Ca isotope analysis in this study, indicate a temperature 300–360 °C. This range of temperatures is well above that predicted from the natural anhydrite-fluid Ca isotope fractionation data reported here, approximately 150–200 °C, when compared to the tentative temperature dependent fractionation trend shown in this study (Fig. 4). This disparity is likely due to the experimental data not being indicative of equilibrium fractionation conditions upon precipitation or a reflection of the dynamic nature of fluids circulating and mixing in the subsurface at TAG (Chiba et al., 1981, 1998; Petersen et al., 1998; Teagle et al., 1998a; Tivey et al., 1998; Shanks, 2001). Further experiments are needed to constrain the uncertainty between equilibrium and kinetic effects produced upon anhydrite precipitation to better understand the interpretation of the Ca elemental and isotopic budget of MOR hydrothermal systems.

5. CONCLUSIONS

The experimental data presented in this study address the effect of temperature and time on Sr/Ca partitioning and exchange and fractionation of Ca isotopes between fluid and anhydrite. Experimental results indicate that Sr/Ca partitioning between anhydrite and dissolved components are subject to disequilibrium effects, which change with time as anhydrite undergoes recrystallization and crystal growth. In addition, the associated Ca isotope fractionation between anhydrite and dissolved Ca$^{2+}$ ($\Delta^{44/40}$Ca(Anh−Fluid)) results in enrichment of heavy isotopes of Ca in solution, although the magnitude of this decreases with increasing temperature, consistent with previous experimental data. The addition of an enriched $^{43}$Ca source to the anhydrite-bearing system allows quantification of the rate of isotopic exchange between the mineral and fluid reservoirs. Data indicate that the rate of exchange is a clear function of temperature and the apparent mechanism of anhydrite is recrystallization. Coupling of experimental $^{43}$Ca exchange rate data with time-series Sr/Ca partitioning data permitted extrapolation towards complete exchange between the anhydrite-fluid reservoirs, resulting in equilibrium Sr/Ca isotope partitioning values, $K_{Sc}$, at 175, 250,
Physiochemical processes controlling the chemical and isotopic composition of anhydrite and hydrothermal fluid recovered from the TAG and Logatchev hydrothermal systems are informed by the experimental results. Specifically, conditions in the subseafloor where mixing of end-member hydrothermal fluid with conductively heated seawater can result in elevated $K_D$ values owing to kinetic effects imposed by anhydrite formation from supersaturated fluids. The opposite is likely true of anhydrite precipitation along fluid pathways where temperature increase is more gradual, such as seawater recharge zones. In this case, equilibrium Sr/Ca partitioning values can be realized. Although the precipitation of anhydrite is a significant process in the subseafloor of modern MOR hydrothermal systems, the Ca isotope composition of end-member vent fluid from TAG and Logatchev remain within error of the composition of MORB. Further experimental studies and additional analyses of Ca isotope fractionation data between anhydrite, altered and secondary silicates, and dissolved Ca$^{2+}$ from MOR hydrothermal systems are needed to properly constrain equilibrium fractionation at a range of temperatures and to better understand the mass transfer of Ca and associated elements, such as Sr, between circulating fluids in the subseafloor with the oceanic crust.

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