

# **Fluid mixing and spatial geochemical variability in the Lost City hydrothermal field chimneys**

**Karina A. Aquino<sup>1,2</sup>, Gretchen L. Früh-Green<sup>1</sup>, Stefano M. Bernasconi<sup>1</sup>, Jörg Rickli<sup>1</sup>, Susan Q. Lang<sup>3</sup>, and Marvin D. Lilley<sup>4</sup>**

<sup>1</sup>Department of Earth Sciences, ETH Zürich, Sonneggstrasse 5, 8092, Zürich, Switzerland.

<sup>2</sup>Department of Science and Technology - Philippine Nuclear Research Institute, Commonwealth Avenue, Diliman, Quezon City 1101, Philippines.

<sup>3</sup>Woods Hole Oceanographic Institution, 360 Woods Hole Road, Woods Hole, MA 02543, USA.

<sup>4</sup>School of Oceanography, University of Washington, Seattle, WA 98195, USA

Corresponding author: Karina A. Aquino ([karmina.aquino@erdw.ethz.ch](mailto:karmina.aquino@erdw.ethz.ch))

## **Key Points:**

- The mineralogy and geochemistry of Lost City chimneys are controlled by the extent of mixing between hydrothermal fluids and seawater
- Brucite and calcite precipitate in vent fluid dominated zones while aragonite forms in the exterior of the structures in seawater-rich zones
- Carbonates precipitate in isotopic disequilibrium and record the O and C stable isotope composition of seawater dissolved inorganic carbon.

## **Abstract**

Carbonate-brucite chimneys are a characteristic of low- to moderate-temperature, ultramafic-hosted alkaline hydrothermal systems, such as the Lost City hydrothermal field located on the

Atlantis Massif at 30°N near the Mid-Atlantic Ridge. These chimneys form as a result of mixing between warm, serpentinization-derived vent fluids and cold seawater. Previous work has documented the evolution in mineralogy and geochemistry associated with the aging of the chimneys as hydrothermal activity wanes. However, little is known about spatial heterogeneities within and among actively venting chimneys. New mineralogical and geochemical data ( $^{87}\text{Sr}/^{86}\text{Sr}$  and stable C, O, and clumped isotope) indicate that brucite and calcite precipitate at elevated temperatures in vent fluid-dominated domains in the interior of chimneys. Exterior zones dominated by seawater are brucite-poor and aragonite is the main carbonate mineral. Carbonates form mostly out of oxygen and clumped isotope equilibrium due to rapid precipitation upon vent fluid-seawater mixing. In contrast, the carbonates precipitate close to carbon isotope equilibrium, with dissolved inorganic carbon in seawater as the dominant carbon source, and have  $\delta^{13}\text{C}$  values within the range of marine carbonates. Our data suggest that calcite is a primary mineral in the active hydrothermal chimneys and does not exclusively form as a replacement of aragonite during later alteration with seawater. Elevated formation temperatures and lower  $^{87}\text{Sr}/^{86}\text{Sr}$  relative to aragonite in the same sample suggest that calcite may be the first carbonate mineral to precipitate.

#### **Plain Language Summary**

At the Lost City hydrothermal field, warm, alkaline fluids are discharging in uplifted mantle rocks. When vent fluids mix with seawater at the seafloor, carbonate and brucite minerals form spectacular towers up to 60 m high. Systems like Lost City are important because the reaction between water and rocks provide carbon and energy sources for microbial life. However, we still do not fully understand what controls the mineralogy and geochemistry of the Lost City hydrothermal chimneys. In this paper, we suggest that the extent of mixing between the hydrothermal fluids and seawater influence the mineralogy and geochemistry of the chimneys. Calcite, which is previously known to form only during alteration of aragonite by seawater, can also form during seawater-hydrothermal fluid mixing. Both calcite and brucite form in the interior of the chimneys where vent fluid is more dominant. Aragonite, on the other hand, forms in the exterior of the structures from seawater-rich fluids. Lastly, because minerals precipitate rapidly during fluid mixing, the stable isotope geochemistry of the carbonates mostly record the composition and temperature of seawater and not the mixed fluid. Thus, care should be exercised in interpreting geochemical data from similar systems.

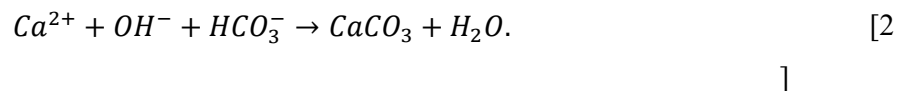
## 60 1 Introduction

61 Since the discovery of the Lost City hydrothermal field (LCHF; Kelley et al., 2001),  
 62 carbonate-brucite deposits associated with serpentinite-hosted hydrothermal systems are  
 63 increasingly being recognized in different tectonic settings and are possibly widespread  
 64 components of serpentinizing environments on the seafloor. Carbonate-brucite deposits occur  
 65 at or near slow- and ultraslow-spreading mid-oceanic ridges (Kelley et al., 2001; Lartaud et  
 66 al., 2011; Lecoeuvre et al., 2020), in subduction forearc regions (Ohara et al., 2012; Okumura  
 67 et al., 2016), in magma-poor continental margins (Klein et al., 2015; Schwarzenbach, Früh-  
 68 Green, et al., 2013), and they are also associated with continental ophiolitic settings (Launay  
 69 & Fontes, 1985; Monnin et al., 2014; Pisapia et al., 2017). Unlike high-temperature black-  
 70 smoker hydrothermal systems (>350°C) that are characterized by acidic (pH 2-6), metal- and  
 71 sulfide-rich vent fluids (Von Damm, 1990; German & Seyfried, 2013), fluids derived from  
 72 the serpentinization of ultramafic rocks are relatively cool (~30-120°C), alkaline (pH 9-11),  
 73 metal- and CO<sub>2</sub>-poor, and often rich in dissolved volatiles such as H<sub>2</sub>, CH<sub>4</sub>, and other short  
 74 chain hydrocarbons (Cipolli et al., 2004; Eickenbusch et al., 2019; Kelley et al., 2005; Lang  
 75 et al., 2010; Monnin et al., 2014; Schwarzenbach, Lang, et al., 2013; Seyfried et al., 2015).  
 76 Instead of sulfide structures typical of black-smoker systems, delicate carbonate-brucite  
 77 chimneys are formed in zones where serpentinization-derived hydrothermal fluids mix with  
 78 oxygenated, Mg- and CO<sub>2</sub>-rich seawater (Früh-Green et al., 2003; Kelley et al., 2001;  
 79 Ludwig et al., 2006; Okumura et al., 2016; Pisapia et al., 2017):



seawate    vent fluid    brucite

r



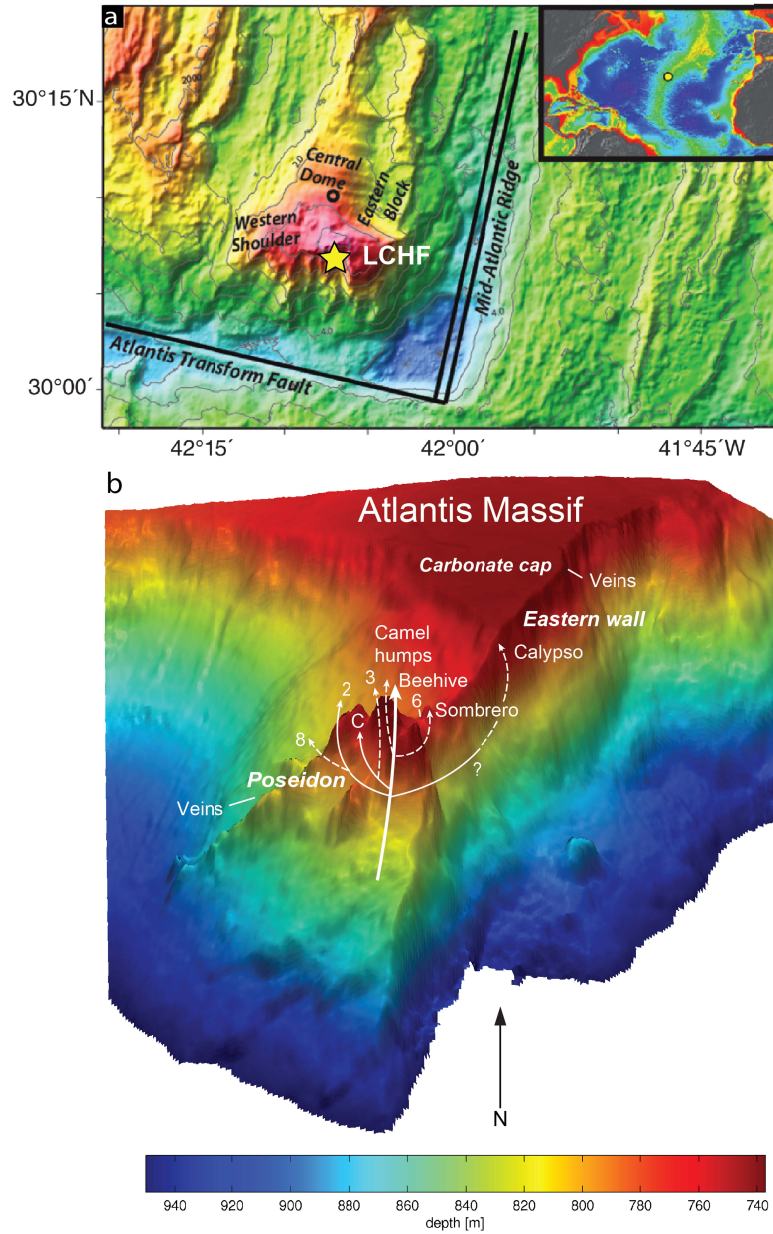
vent fluid    seawater    Ca-carbonate

80 Serpentinization reactions, which can provide both carbon and energy sources for  
 81 chemoautotrophy (Amend et al., 2011; Dick & Shock, 2021; McCollom, 2007; McCollom &  
 82 Seewald, 2007), are likely to have been widespread in the early Earth (Sleep, 2010) and could  
 83 potentially be occurring in other planetary bodies such as Enceladus and Europa (Glein et al.,

2015; McKay et al., 2008; Waite et al., 2017). Therefore, systems like the LCHF can improve our understanding of the geochemical evolution of life on Earth and elsewhere.

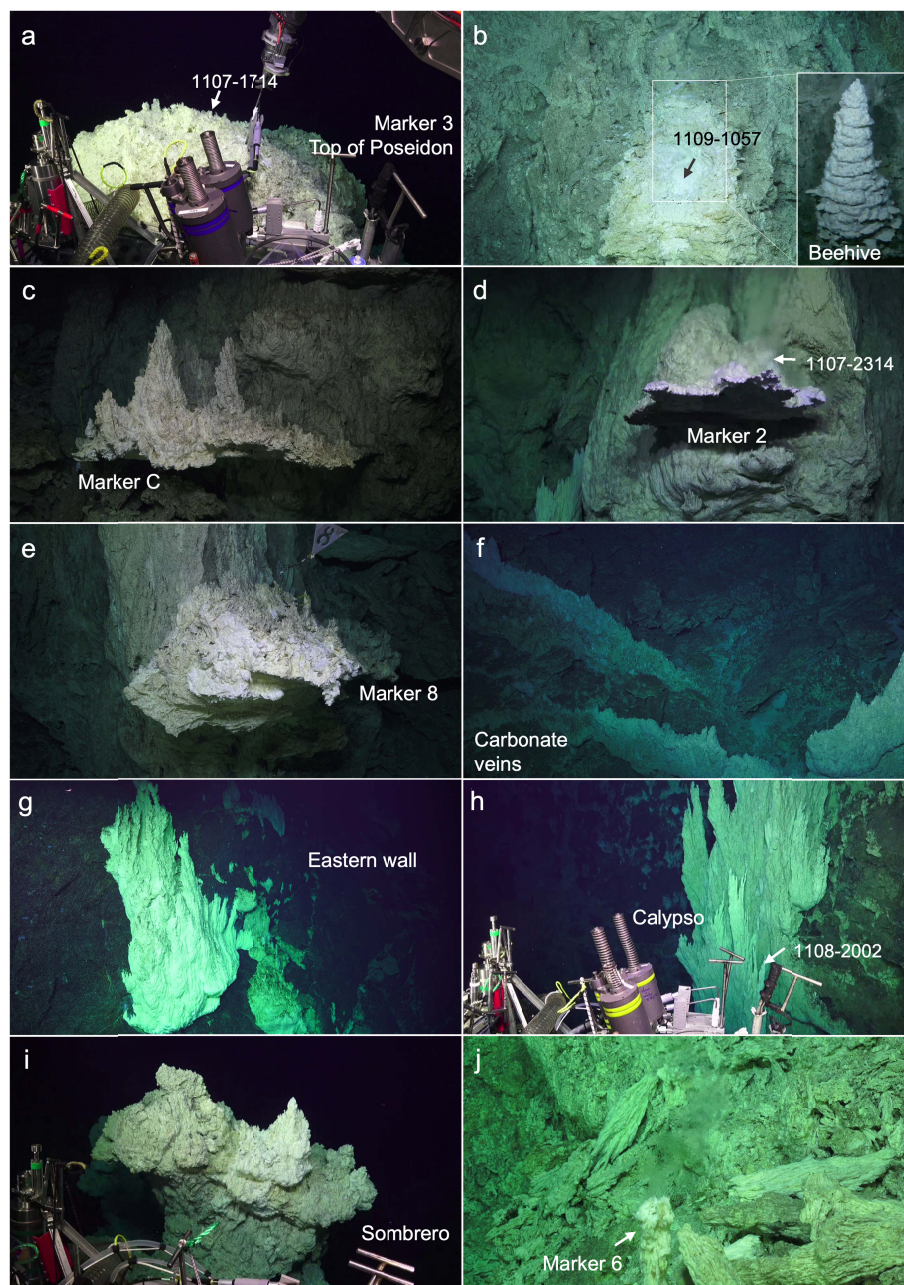
The LCHF is the first documented and remains one of the best studied examples of such low temperature, marine hydrothermal systems to date. It is located approximately 15 km west of the Mid-Atlantic Ridge at 30°N at the inside corner of the intersection with the Atlantis Transform Fault (Figure 1). The field is at the top of a terrace on the southern wall of the Atlantis Massif at about 700-800 m below sea level. At least 30 carbonate-brucite structures (Figure 2), up to 60 m high, span across the approximately 400 m wide vent field and include actively venting and inactive spires, pinnacles, parasitic flanges, and veins (Kelley et al., 2001). Ludwig et al. (2006; 2011) extensively studied the mineralogical, geochemical, and textural transformation during ageing of the chimneys from nascent to inactive and extinct structures. They documented that young, active chimneys precipitate predominantly as aragonite and brucite, characterized by low  $^{87}\text{Sr}/^{86}\text{Sr}$  and low trace element contents. Progressive reaction with seawater results in the conversion of aragonite to calcite, dissolution of brucite, as well as increases in  $^{87}\text{Sr}/^{86}\text{Sr}$  and trace element concentrations (Ludwig et al., 2006). In our companion paper (Aquino et al., 2023), we show that calcite is also a primary mineral in active chimneys and does not only form from aragonite alteration. Calcite is associated with brucite and occurs in the interior of actively venting spires as well as the downward-facing side of flanges. In contrast, aragonite generally comprises the exterior of the chimneys (Figure 3).





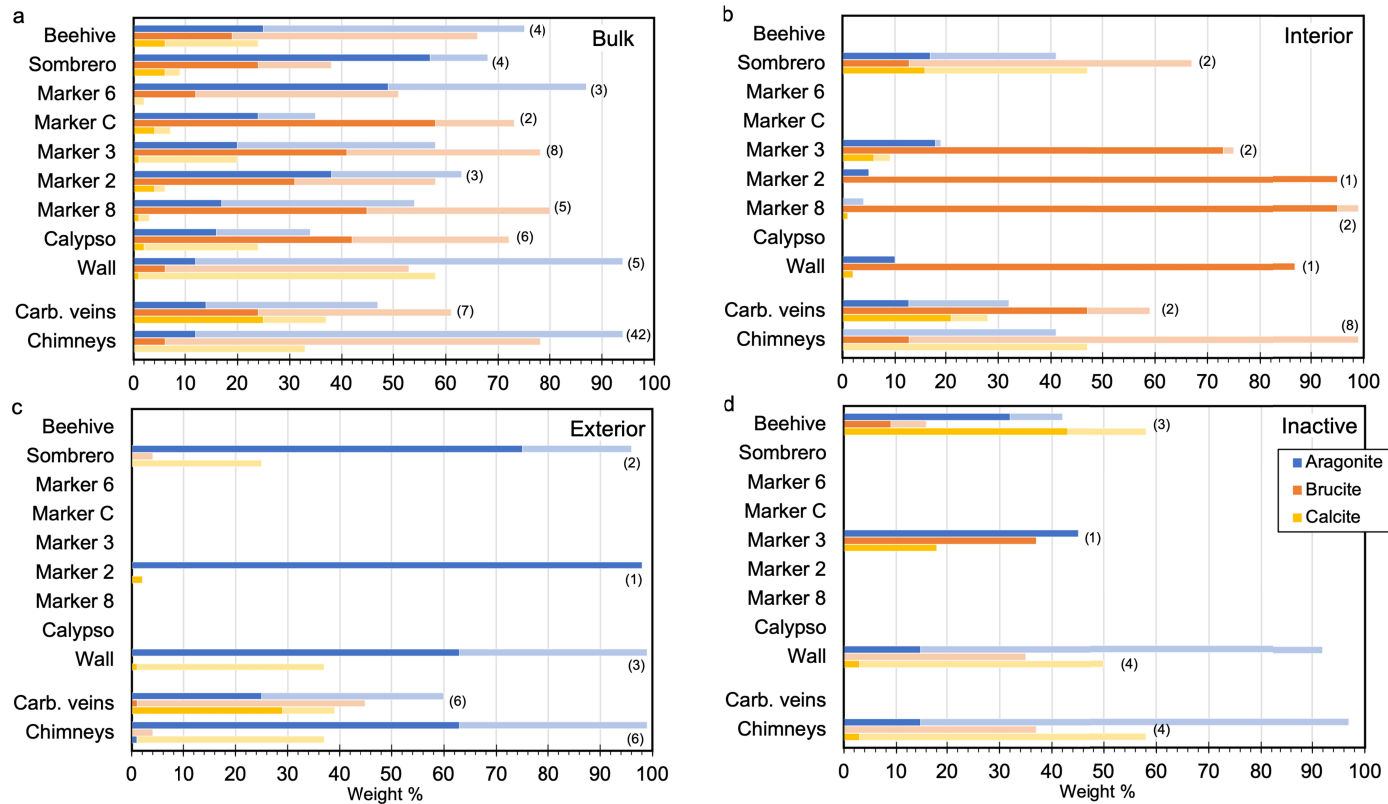
**Figure 1.** Location of the Lost City hydrothermal field (LCHF) and the vent sites investigated in this study. (a) Overview of the LCHF situated on the southern wall of the Atlantis Massif at 30°N west of the Mid-Atlantic Ridge. (b) 3-Dimensional map of the LCHF (after Kelley et al., 2005) including the vents studied here. The massive 60-m structure Poseidon is a composite of multiple active vents in the center of the field. Active vents are also found along the northeast trending escarpment called eastern wall. Carbonate veins were collected close to the eastern wall and west of Marker 8. Also shown are vent fluid flow paths showing the relationship between vent sites after Aquino et al. (2022): thick solid line, 1st

113 order vent - Beehive; thin solid line, 2nd order vents - Marker C and Marker 2; thin dashed  
114 lines, 3rd order vents.



**Figure 2.** Dive photos of vent sites investigated. (a) Marker 3 located at the top of the Poseidon structure. (b) Location of active venting at what used to be the Beehive structure. White box indicates the general location of the Beehive structure and inset shows the structure in 2005 (Kelley et al., 2005). (c) Marker C flange. (d) Marker 2 or IMAX flange. (e) Marker 8 flange. (f) Carbonate veins or fissure-filling deposits located west of Marker 8. (g) Carbonate spires along the eastern wall. (h) Calypso spire, also located along the eastern wall. White arrow points to the spire collected as sample 1108-2002. (i) Top of the Sombrero

123 structure. (j) Marker 6 spire. Images courtesy of S. Lang, UofSC / NSF / ROV Jason / 2018  
124 © WHOI



125

126 **Figure 3.** Mineral abundances of samples collected at each site. (a) Subsamples representative of the bulk sample. (b) Chimney interior. (c)  
 127 chimney exterior. (d) Inactive chimneys. Notes: Number of samples analyzed for each site is shown in parentheses. Blue = aragonite; orange =  
 128 brucite; yellow = calcite. Dark and light shades represent minimum and maximum mineral abundances, respectively. Data from Aquino et al.  
 129 (2023). Samples classified as “inactive” from the active structures Beehive and Marker 3 are older parts of the chimney that are yellowish and  
 130 show a higher degree of induration and form the substrate for new bright white chimney growth.

The objective of this study is to understand the controls on the spatial variations observed in the mineralogy of the active chimneys, more specifically, to determine when and where calcite or aragonite precipitates. Our geochemical data show that heterogeneous precipitation conditions (e.g., pH, temperature, Mg/Ca ratios) at Lost City are largely controlled by the variability in mixing proportions of vent fluid and seawater, which affects the mineralogy and geochemistry of the resulting hydrothermal chimneys. Stable carbon and oxygen isotope data suggest that during the very rapid precipitation of carbonate upon fluid mixing, the dissolved inorganic carbon (DIC) behaves as a closed system preventing equilibration between DIC, which is mostly sourced from seawater, and  $\text{H}_2\text{O}$  or  $\text{OH}^-$ . As a result, the carbonates record oxygen and clumped isotope compositions that do not reflect the temperature of the formation fluid.

## 2 Sampling sites

The locations of the sampled vent structures are shown in Figure 1 and listed in Table S1. Here we only briefly summarize the morphology of the structures which has been discussed in previous studies (Früh-Green et al., 2003; Kelley et al., 2001, 2005; Ludwig et al., 2006). We use location names described in earlier studies and names of field markers deployed during previous sampling campaigns. Carbonate-brucite chimneys occur in a wide variety of morphologies. The central and most prominent part of the field is the 60-m high Poseidon structure, which is a composite of actively venting towers and spires including Marker 3 and Camel humps, and parasitic vent structures forming cone- and flange-like morphologies (Figures 1, 2). The Beehive vent is a ~1-m high, cone-like parasitic structure that vents the hottest fluid (up to 96 to 116°C; Table S1) which can be considered the primary, unaltered endmember fluid at LCHF (Aquino et al., 2022; Kelley et al., 2005; Ludwig et al., 2006). The structure itself was no longer present during our sampling in 2018, but we were able to collect chimney samples from a venting orifice, presumably representing the interior of the former structure (Table S2, Figure 2b). Marker C and Marker 2 (also called IMAX flange), are parasitic flanges on the Poseidon structure which include horizontally growing structures that trap warm vent fluids (Figure 2c, d). In some cases, chimney spires grow on the top of these flanges. We also sampled Marker 8 (Figure 2e), another parasitic flange west of Poseidon, as well as several veins or fissure filling carbonate-brucite deposits located west of Marker 8 (Figure 2f).



The eastern wall is a steep escarpment where carbonate veins and morphologically diverse chimney structures (Figure 2g) grow directly along faults or subhorizontal foliations in the serpentinites (Kelley et al., 2005). It is located along a ridge extending towards the northeast from the main vent field (Figure 1). We sampled several structures, including actively venting spires such as Calypso (Figure 2h), and others that showed no evidence of venting at the time of sampling. Veins that crosscut the relatively flat-lying, carbonate-cap sedimentary sequences at the top of the southern Atlantis Massif (Kelley et al., 2005) were also sampled. Marker 6 and Sombrero are two sites located east of Poseidon and close to the eastern wall. Sombrero is an actively venting spire with fresh growth at the top of the structure (Figure 2i). The venting structure on Marker 6 on the other hand is a small, delicate, actively venting spire that grew atop rubble of inactive chimneys (Figure 2j). Despite their location away from Poseidon at the center of the field, these vents are thought to share a common flow path with Beehive (Figure 1c; Aquino et al., 2022; Seyfried et al., 2015).

### 3 Materials and Methods

#### 3.1 Samples

Grab and suction (slurp) samples of hydrothermal chimneys and fissure-filling veins were collected with the remotely operated vehicle (ROV) *Jason* during the Lost City 2018 Return expedition (*R/V Atlantis* cruise AT42-01). Chimney samples were mostly obtained in sites where active venting, recognized primarily by shimmering water and increased water temperatures, was observed. During six dives (Dives 1107 to 1112), we sampled sites that had been previously investigated (Figure 1), such as Beehive and sites named after field markers laid out in 2003 (e.g., Marker 2, Marker C; Kelley et al., 2005). Most of these vents are associated with the massive Poseidon structure at the center of the field (Figure 1). Other active vents along the eastern wall were sampled for the first time in 2018. A total of 31 chimneys and 6 vein samples were collected and described macroscopically. These samples were assigned a unique ID including the *Jason* dive number, the date and time of collection, and the sample type (e.g., J.1109.19Sep.0756 CHIM for a chimney sample collected on September 19, 2018, 07:56 GMT during *Jason* Dive 1109). In this paper, we use a shortened version of the sample ID which includes the dive number and time of collection (e.g., 1109-0756). Each sample was subdivided onboard into aliquots and distributed among the science party members. Each of these samples were assigned a unique cruise ID (e.g., sample 1109-0756 is further subdivided into samples with cruise ID LC01349, LC01353, etc.). Specimens

that exhibited heterogeneities such as distinct colors, textures, or layers were subsampled to analyze their mineralogy and geochemistry (e.g., LC01349a, LC01349b). Samples representing a general average or bulk composition of a chimney or vein were also analyzed separately. Mineral separation via hand picking was carried out on selected samples to determine texture- and/or mineralogy-specific isotopic compositions.

### 3.2 Analytical methods

A total of 76 samples from 31 chimneys and 6 carbonate veins were analysed for mineralogy, chemistry, and stable C and O isotope compositions, 33 for  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios, and 24 for carbonate clumped isotopes ( $\Delta_{47}$ ). 14 and 12 mineral separates were also measured for their  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $\Delta_{47}$  values, respectively. For major and trace element analyses, powdered samples were washed with distilled water and centrifuged several times to remove salt and dirt, and subsequently freeze dried. 300  $\mu\text{g}$  of sample material were then dissolved in 300  $\mu\text{l}$  of 2%  $\text{HNO}_3$  and measured on an Agilent 8800 Triple Quadrupole inductively coupled plasma mass spectrometer (ICP-MS) at the Geological Institute, ETH Zürich. The reproducibility of the analyses is better than  $\sim 5\%$  (RSD) based on repeated analyses of standards.

C and O isotope analyses were conducted on 90 to 140  $\mu\text{g}$  of carbonates (powdered and washed) using a GasBench II system coupled with a Delta V isotope ratio mass spectrometer (IRMS, Thermo Fischer Scientific, Bremen, Germany) at the Geological Institute, ETH Zürich following the methodology of Breitenbach & Bernasconi (2011). Results are reported in the conventional delta notation relative to the Vienna Pee-Dee Belemnite (VPDB). The average standard deviation of repeated measurements of standards during different sessions is less than 0.1‰ for both  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ .  $\delta^{18}\text{O}$ -based precipitation temperatures were calculated using the calibration of O'Neil et al. (1969) as recalculated by Friedman & O'Neil (1977).

Carbonate clumped isotope analyses were performed with a Kiel IV carbonate device coupled to a 253Plus isotope ratio mass spectrometer (IRMS) (both Thermo Fisher Scientific) at the Geological Institute, ETH Zürich as described by Müller et al. (2017) and Meckler (2014). The carbonate samples were reacted with three drops of 104%  $\text{H}_3\text{PO}_4$  at 70°C via the Kiel IV device. The  $\text{CO}_2$  gas released from this reaction was then purified in a custom-built Porapak Q resin held at -40°C to remove potential isobaric contaminations. Samples were measured with a maximum of three replicate measurements per sample and session, which



generally consists of 24 measurements of 100-120 µg sample carbonate interspersed with 5 replicates each of the carbonate standards ETH-1, ETH-2 and 10 replicates of ETH-3 (Bernasconi et al., 2018, 2021). The samples were analysed in LIDI mode with 400 seconds of integration of sample and reference gas. Raw data processing and background corrections were performed using the Easotope software (John & Bowen, 2016). Sample compositions are reported as the excess abundance of the CO<sub>2</sub> isotopologue with mass 47 relative to the abundance expected from a random distribution of isotopes:

$$\Delta_{47} = \frac{R^{47}}{R^{47*} - 1}, \quad [3]$$

where  $R^{47}$  is the ratio of the abundance of the CO<sub>2</sub> isotopologue with mass 47 relative to the abundance of the most common CO<sub>2</sub> isotopologue with mass 44 and  $R^{47*}$  is the ratio expected from a random distribution calculated from the measured  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values. Sample  $\Delta_{47}$  measurements were converted to the  $\Delta_{47}$  Intercarb-Carbon Dioxide Equilibrium Scale (ICDES) (Bernasconi et al., 2021) using the ETH-1, ETH-2 and ETH-3 carbonate standards described previously by Bernasconi et al. (2018); Meckler et al. (2014); Müller et al. (2017) with the community-derived  $\Delta_{47}$  values reported by Bernasconi et al. (2021).  $\Delta_{47}$  based temperatures were calculated using the calibration of Anderson et al. (2021).

Radiogenic Sr isotope and Sr concentration analysis were performed at the Institute of Geochemistry and Petrology, ETH Zürich. Twenty to thirty mg of powdered carbonate (bulk samples, separated aragonite or calcite) were first reacted in 5 ml of ammonium acetate buffered to pH 5 to leach the carbonate fraction. Leached sample fractions were subsequently dried down and measured dilute for Sr concentrations by ICP-MS (Element XR, Thermo Fisher Scientific). 200 to 500 ng of Sr were separated from the carbonate matrix using Sr spec resin following the methods of Deniel & Pin (2001) and de Souza et al. (2010). Strontium isotope ratios ( $^{87}\text{Sr}/^{86}\text{Sr}$ ) were measured at Sr concentrations of ~100 ppb on a multicollector ICP-MS (Neptune Plus, Thermo Fischer Scientific). Instrumental mass fractionation was corrected for using the exponential law and  $^{86}\text{Sr}/^{88}\text{Sr}$  of 0.11940 (Nier, 1938). The uncertainty of the measurements estimated from repeated measurements of Nist SRM 987 is < 24 ppm (2 SD, n = 15 per session). Sample  $^{87}\text{Sr}/^{86}\text{Sr}$  values were renormalized to the accepted value of Nist SRM 987 ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.710248$ , Thirlwall, 1991).

## 4 Results

The mineralogy of the samples, summarized in Figure 3, is discussed in detail in the companion paper (Aquino et al., 2023). The Lost City chimneys are composed of variable mixtures of aragonite, brucite, and calcite (Figure 3a). In general, the interior of these chimneys is composed mostly of brucite and minor calcite/aragonite (Figure 3b), while the exterior is dominated by aragonite (Figure 3c). Scanning electron microscopy of samples from the interior of the chimneys highlighted the spatial association between brucite and calcite (Aquino et al., 2023). Several samples from relatively high temperature vents (e.g., Beehive and Sombrero) as well as from eastern wall sites (Calypso and carbonate veins), contain significant amounts of calcite (>20%). Many samples exhibit flow textures preserved as brucite mineral membranes upon which aragonite or calcite may precipitate on. In the following we compare the results of our geochemical analyses with previous mineralogical studies (Tables S2 and S3).

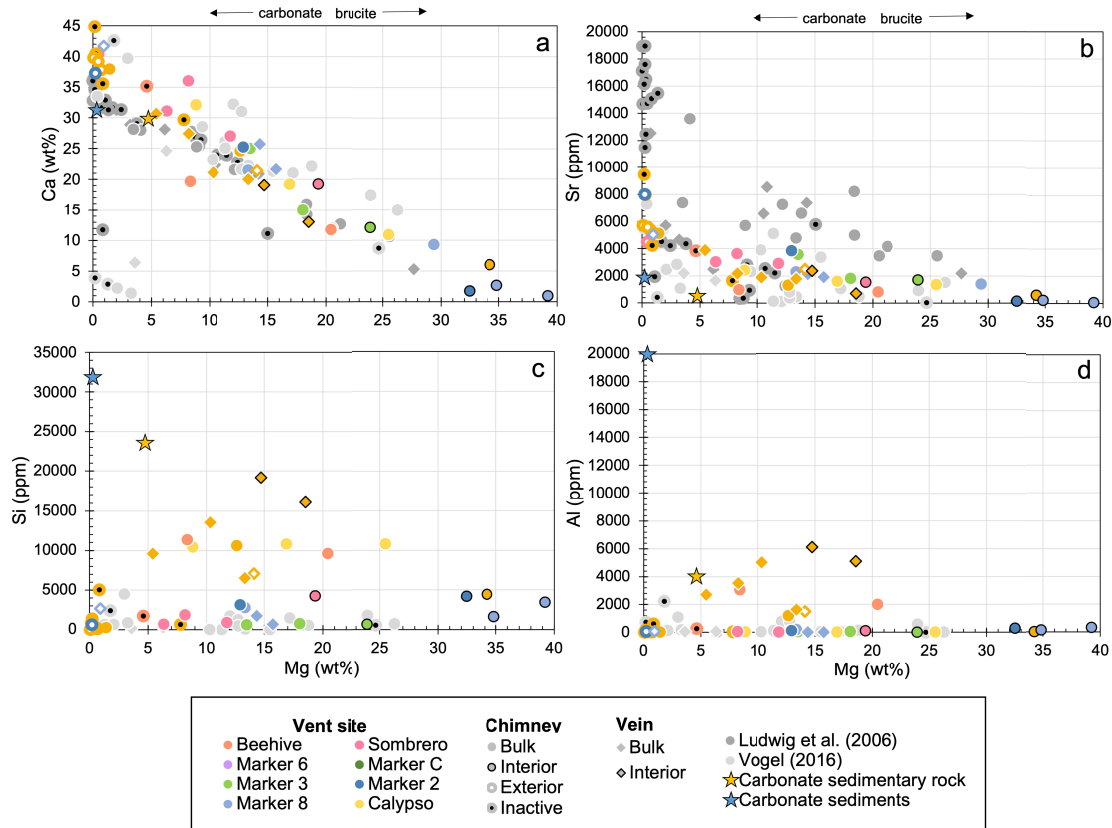
### 4.1 Major and trace elements

The major and trace element compositions of the samples are shown in Figure 4. In general, variations in Ca, Mg, and Sr concentrations (Figure 4a, b) roughly reflect the relative proportion of carbonate and brucite in the chimneys, consistent with previous studies (Ludwig et al., 2006; Vogel, 2016). The brucite-rich chimney interior contains up to 40 wt% Mg whereas the carbonate-rich exterior and inactive chimneys have up to 45 wt% Ca (Figure 4a). Brucite does not appear to incorporate significant Sr because the Sr contents of the brucite-rich (>95 wt% brucite, >30 wt% Mg) samples are low (70 to 237 ppm). In comparison, the chimney exterior has more than 4000 ppm Sr, reflecting a higher proportion of aragonite (Figure 4b). Inactive structures contain even more elevated amounts of Sr (up to about 19,000 ppm). Interestingly, most carbonate veins and samples from the eastern wall and Beehive show elevated Si and Al contents (>6000 ppm and 1000 ppm, respectively) relative to samples from other sites (<5000 ppm Si, < 1000 ppm Al). These values are similar to the average marine carbonate sedimentary rock and lower than those of the average marine unlithified carbonate sediments (Figure 4c, d). Most trace element (Cr, Mn, Ni, Fe, Zn, Pb) concentrations are significantly lower than those of the average marine carbonate rock and sediments (Figure S1). On the other hand, Sr, Cd, U, P, and Si have similar or slightly higher concentrations relative to the average marine carbonate rock and carbonate sediments (Figure

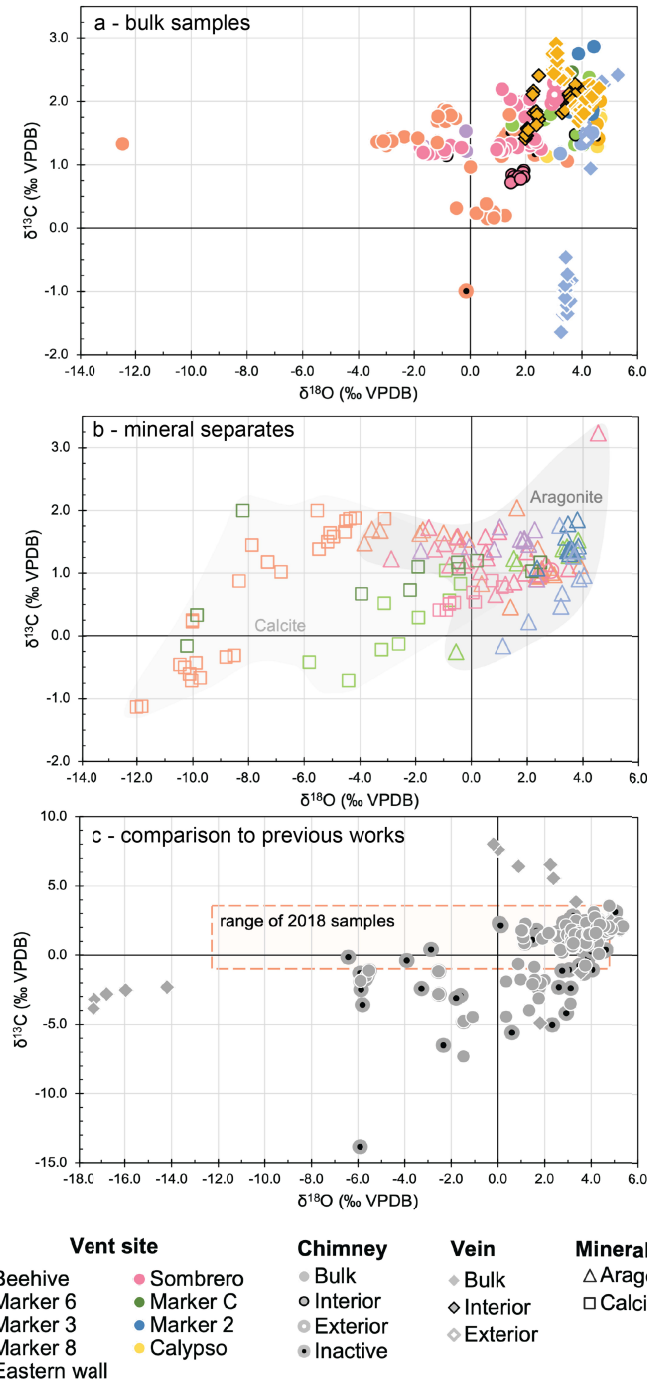
S1). Ba, U, P, and S show a generally increasing trend with decreasing Mg while Zn roughly increases with Mg (Figure S1).

#### 4.2 Oxygen isotopes

The Lost City chimney carbonates are isotopically variable and show within-sample heterogeneity at the 100  $\mu\text{g}$  sample size used for the measurements (Fig. 5). The  $\delta^{18}\text{O}$  of mixed calcite and aragonite bulk samples of active chimneys span a wide isotopic range from -12.45 to 4.67 ‰ (111 analyses of 35 samples) whereas the veins have a narrower range from 1.99 to 5.29 ‰ (116 analyses of 10 samples) (Figure 5a; Table S2). Beehive, Sombrero, and Marker 6, which belong to the vent group with the lowest dissolved sulfide concentrations of 0.3 to 0.4 mmol/kg (Aquino et al., 2022), have the lowest and most variable bulk  $\delta^{18}\text{O}$  values (-12.45 to +3.49 ‰; 62 replicates of 10 samples). The interior of the chimneys and veins from the same site have similar or lower  $\delta^{18}\text{O}$  than the bulk, whereas samples from the exterior have similar or higher  $\delta^{18}\text{O}$  than the bulk (e.g., Figures 5, S2b, e, f). The  $\delta^{18}\text{O}$  of inactive structures are generally similar to those of the active ones (Figure S2a e, i). The  $\delta^{18}\text{O}$  of calcite crystals and aggregates separated by hand picking ranges between -12.0 and +2.5‰ (n=54) whereas aragonite  $\delta^{18}\text{O}$  values are -3.8 to + 4.0‰ (n=85) (Tables S3, S4; Figure 5b, c). Within each site, calcite  $\delta^{18}\text{O}$  values are generally lower than or similar to values of the mixed samples (Figure S2a, b, d, e).  $\delta^{18}\text{O}$  values of separated aragonites are mostly similar to or slightly higher than the bulk compositions (Figure S2a-g). Except for a few fissure-filling deposits, the oxygen isotope composition of samples collected in 2018 are similar to those reported previously (Figure 5c) (Früh-Green et al., 2003; Vogel, 2016).



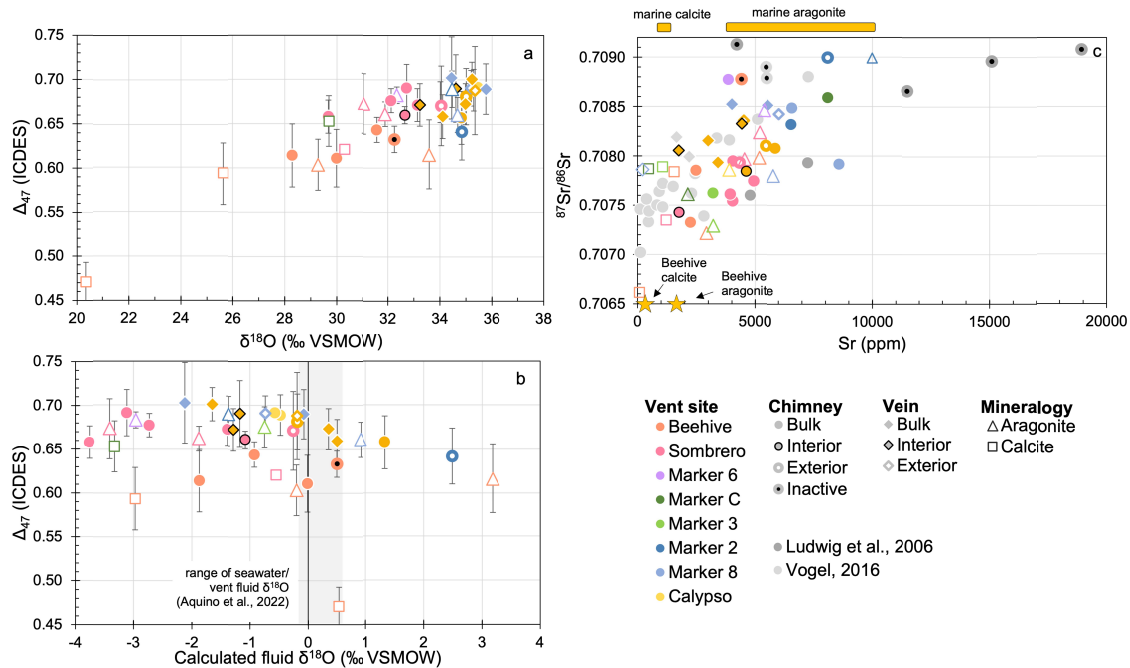
**Figure 4.** Major element concentrations vs. Mg of active carbonate chimneys and fissure filling veins including data from Ludwig et al. (2006) and Vogel (2016). (a) Variation in Ca and Mg generally reflects the relative proportions of carbonate and brucite in the samples. Note that the interior of the chimneys are generally Mg rich relative to the exterior. (b) Sr contents increase with increasing Ca and decreasing Mg concentrations. Active structures have significantly lower Sr contents than the inactive chimneys. (c-d) Most samples have minor amounts of Si and Al except for a few samples from the eastern wall and Beehive. Data for marine carbonate sedimentary rock and unlithified marine carbonate sediments are from Turekian and Wedepohl (1961).



**Figure 5.** Carbon and oxygen isotope compositions of the Lost City hydrothermal chimneys and veins: (a) bulk samples and (b) mineral separates. Gray fields indicate the range of compositions of calcite and aragonite crystals. (c) Comparison of samples collected in 2018 (mostly active samples) and previous work (active, inactive, fissure filling) (Früh-Green et al., 2003; Vogel, 2016).

### 4.3 Carbon isotopes

The bulk  $\delta^{13}\text{C}$  of the active chimneys (0.15 to 2.46‰;  $n=101$ , Figure 5, Table S2) lie within a narrower range than the  $\delta^{18}\text{O}$ . Although the inactive chimneys have generally slightly higher bulk  $\delta^{13}\text{C}$  values (1.52 to 2.26‰;  $n=25$ ), these isotope compositions still lie within the range of active chimneys (Figure S2a, e, i). Most of the veins (0.94 to 2.91‰;  $n=102$ ) have  $\delta^{13}\text{C}$  values very similar to the chimneys except for one vein west of Marker 8 (average  $\delta^{13}\text{C} = -1.06$ ‰; Table S2, Figures 5a, S2j). As with  $\delta^{18}\text{O}$ , for samples from the same site, subsamples from the interior of the chimneys have similar or slightly lower  $\delta^{13}\text{C}$  than the bulk, while subsamples from the exterior of the chimneys have similar or slightly higher  $\delta^{13}\text{C}$  values than the bulk (Figure S2b, e, j). Both pure calcite and aragonite have  $\delta^{13}\text{C}$  values that lie in a relatively wide range between -1.1 to +2.0‰ ( $n=54$ ) and between -0.3 to +2.0‰ ( $n=85$ ), respectively. The average  $\delta^{13}\text{C}$  for calcite is lower than for aragonite (Tables S3, S4). The carbon isotope composition of LCHF samples reported in earlier studies span a wider range than those of samples collected in 2018 (Figure 5c) (Früh-Green et al., 2003; Vogel, 2016).



**Figure 6.** Carbonate clumped isotope compositions and Sr geochemistry of the Lost City chimneys and veins.  $\Delta_{47}$  values plotted against (a) oxygen isotope compositions and (b) formation fluid oxygen isotope compositions calculated from carbonate clumped isotope

compositions. (c)  $^{87}\text{Sr}/^{86}\text{Sr}$  values and Sr concentrations of selected Lost City carbonate samples. Also shown are the range of marine calcite and aragonite from Kinsman (1969). Note: For bulk samples, carbonate Sr contents plotted here are estimated from measured Sr concentrations and the relative abundance of carbonates vs. brucite (Table S3).

#### 4.4 Clumped isotopes

The bulk  $\Delta_{47}$  values of the chimneys vary from 0.611 to 0.702‰ which correspond to temperatures of -7 to 20°C (Table S2, Figure 6a, b). Aragonite from these samples have  $\Delta_{47}$  values that lie within a slightly lower but similar range (0.604 to 0.690‰; -3 to 22°C). Calcite, on the other hand, has generally lower  $\Delta_{47}$  than the bulk samples and aragonite (0.471 to 0.653‰; 7 to 78°C) (Tables S2, S3, Figure 6a, b). Samples from Beehive have the lowest and most variable  $\Delta_{47}$  while the carbonate veins have the highest and least variable  $\Delta_{47}$  values. In samples where both aragonite and calcite could be measured individually, calcite has consistently lower  $\Delta_{47}$  than aragonite (Table S3, Figure 6a).

#### 4.5 Sr isotopes

Strontium concentrations and  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios of bulk chimney and vein samples are listed in Table S2 and plotted in Figure 6c together with literature data (Ludwig et al., 2006; Vogel, 2016).  $^{87}\text{Sr}/^{86}\text{Sr}$  isotope ratios, ranging between 0.70732 and 0.70900, broadly increase with Sr concentrations (Figure 6c). Not all samples were analyzed for Sr geochemistry, nevertheless, we observe that samples from the interior have among the lowest and generally have low  $^{87}\text{Sr}/^{86}\text{Sr}$  and Sr concentrations (e.g., Sombrero, eastern wall chimney Table S2, Figure 6c). The opposite is true for the exterior of the chimneys, which have among the highest  $^{87}\text{Sr}/^{86}\text{Sr}$  and Sr concentrations (e.g., Marker 2, Figure 6c). There is no clear relationship between the Sr geochemistry of chimneys and veins and their location within the LCHF, and samples taken from each of the vent groups have overlapping ranges of  $^{87}\text{Sr}/^{86}\text{Sr}$  and Sr concentrations. One sample taken further away from the site of active venting at Beehive (LC02454b) has a significantly higher  $^{87}\text{Sr}/^{86}\text{Sr}$  and Sr concentration ( $^{87}\text{Sr}/^{86}\text{Sr}$  = 0.70878, 3889 ppm Sr) than the samples collected directly at the vent ( $^{87}\text{Sr}/^{86}\text{Sr}$  = 0.70732 to 0.70785, 808 to 1246 ppm Sr). Aragonite separates generally have higher  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios (0.70722 to 0.70899) and Sr concentrations (2131 to 9988 ppm Sr) than calcite separates (0.70661 to 0.70789; 77 to 1552 ppm Sr) (Table S3, S4).

## 5 Discussion

Our results show that the mineralogy and the isotope and element geochemistry of the Lost City hydrothermal chimneys vary spatially within individual chimneys and among different vent sites. In the following, we first discuss the effect of varying proportions of vent fluid to seawater on the mineralogy of the chimneys and veins. Then, we discuss how rapid, instantaneous precipitation during fluid mixing controls the carbonate carbon, oxygen, and clumped isotope compositions.

### 5.1 Variable mixing of vent fluid and seawater

In general, the interior of the chimneys from active vents is brucite( $\pm$ calcite  $\pm$ aragonite)-rich, with our scanning electron microscopy data highlighting a spatial association between brucite and calcite (Aquino et al., 2023). The chimney interior displays lower  $^{87}\text{Sr}/^{86}\text{Sr}$  values and Sr concentrations than the aragonite-rich exterior (Figures 3, 6). This gradient in Sr concentrations and isotopic composition reflect the dominant mineral phases and their affinity to incorporate Sr. The Sr partitioning coefficient of calcite is lower than that of aragonite (Kinsman, 1969); therefore, the aragonite-dominated chimney exterior is expected to incorporate more Sr (Figure 6c). Moreover, the  $^{87}\text{Sr}/^{86}\text{Sr}$  values of carbonates record the  $^{87}\text{Sr}/^{86}\text{Sr}$  of the fluids from which they precipitate (Coggon & Teagle, 2011). Given that both the endmember hydrothermal fluid ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.70650$ , Aquino, Fröh-Green, Rickli, et al., 2022; Ludwig et al., 2006) and seawater ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.70917$ , Palmer & Edmond, 1989) compositions are known, the strontium isotope composition of the carbonates can be used to calculate the mixing proportion of these two fluids during mineral precipitation. Since the temperatures of the vents and seawater are also known, the temperature of the mixed fluid can be estimated (see section 5.3.1).  $^{87}\text{Sr}/^{86}\text{Sr}$  compositions of bulk chimneys and veins suggest mixing of 13 to 69% vent fluid with seawater (n=22). Carbonates from the chimney interior generally formed from higher proportions of vent fluid (32 to 65% vent fluid; n=4) compared to the exterior (6 to 46% vent fluid; n=4). Calcite is formed from a solution containing 48 to 96% vent fluid (n=5) while aragonite is formed with a relatively lower vent fluid contribution (7 to 73%; n=9).

To investigate the effect of varying proportions of vent fluid and seawater on the chemistry (e.g., Mg/Ca), temperature, pH, and mineral saturation during mixing of seawater and LCHF vent fluids, we calculated a mixing model (Figure 7) using the Geochemist's Workbench® (Bethke et al., 2020). In the model, 1 kg of Beehive endmember fluid with

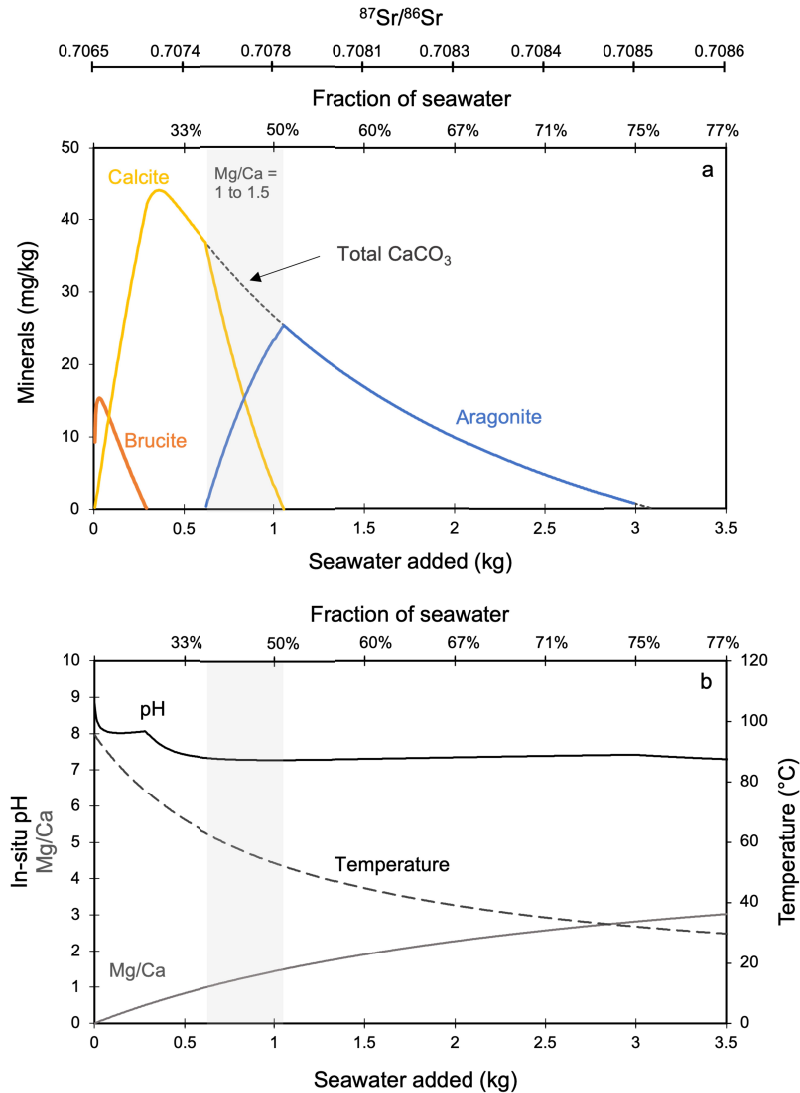


96°C is mixed gradually with seawater at 10°C. The compositions of the vent fluid and seawater used in the calculation are listed in Table S5. The modeling results show that brucite, aragonite, and/or calcite are supersaturated during mixing of up to 3 kg of seawater with 1 kg of Beehive vent fluids (up to 75% seawater). Brucite, whose solubility is highly sensitive to pH (Pokrovsky & Schott, 2004), is supersaturated only at <30% seawater where the pH of the solution is predicted to be greater than 8. Note that pH decreases with increasing temperature because of the temperature dependence of the dissociation constant of water (White, 2009). Thus, the actual in-situ pH of Beehive vent fluids is lower (8.9) than the pH measured in the laboratory at 22°C which is the one commonly reported in the literature (10.8, Table S5; Aquino, Früh-Green, Rickli, et al., 2022; Kelley et al., 2005; Lang et al., 2012). Calcium carbonates, on the other hand, are supersaturated at higher seawater mixing proportions, 0 to ~50% for calcite, and ~35 to ~75% seawater for aragonite. This calculation yields comparable results to other studies, which predicted the precipitation of brucite at lower seawater fractions and increasing calcium carbonate precipitation with continued seawater admixture (Allen & Seyfried, 2004; Palandri & Reed, 2004).

## 5.2 Carbonate mineralogy

The mixing calculation (Figure 7) is consistent with the observation that brucite is generally more abundant in the interior of the structures, while calcium carbonate (Figure 3) is dominant in the exterior. The model cannot be used to determine which carbonate polymorph precipitates, as both calcite and aragonite precipitate when the precipitation of the other mineral is suppressed in the calculations. Our petrographic observations suggest that calcite is an early precipitate. In the companion paper (Aquino et al., 2023), we provide mineralogical and textural evidence that the interior of at least some of the chimneys contain significant calcite whereas their exterior is dominated by aragonite, where a higher proportion of seawater is dominant. Calcite is in most cases intimately associated with brucite and occurs as euhedral, well-formed crystals (Aquino et al., 2023). These mineralogical observations are in agreement with the oxygen- and clumped isotope-derived precipitation temperatures for calcite, which are generally higher than those for aragonite (Tables S3, S4). Temperature and fluid Mg/Ca are the main controls on the calcium carbonate polymorph that precipitates from seawater (Berner, 1975; Davis et al., 2000; Morse et al., 1997, 2007; Sun et al., 2015). At 25°C, calcite precipitates when Mg/Ca is below  $1.4 \pm 0.1$  (Morse et al., 1997) while concurrent precipitation of aragonite and calcite is thermodynamically possible at

Mg/Ca of around 2 (Sun et al., 2015). This threshold Mg/Ca value for calcite precipitation decreases to about 1 at temperatures  $>30^{\circ}\text{C}$ . Based on these considerations, for Beehive, we estimate that co-precipitation of the two calcium carbonate polymorphs is possible at Mg/Ca ratios between  $\sim 1$  and 1.5 (Figure 7a). Below this Mg/Ca, at very low vent fluid/seawater ratios, calcite can coexist with brucite (Figure 3). With increasing seawater contribution, the Mg/Ca ratio of the fluid is higher than the threshold for calcite precipitation and aragonite precipitates.



**Figure 7.** Mixing calculations involving the addition of seawater to 1kg of Beehive vent fluids. (a) Mixing results in the supersaturation of brucite and calcium carbonate. Brucite solubility is highly sensitive to pH and brucite is only saturated at low seawater fractions  $< 30\%$ . Calcite is thermodynamically stable at lower Mg/Ca ratios whereas aragonite is stable

at higher Mg/Ca ratios. (b) Gradients in temperature, pH, and Mg/Ca ratios during mixing. Seawater and Beehive endmember fluid compositions are listed in Table S5.

The model results are broadly supported by the lower  $^{87}\text{Sr}/^{86}\text{Sr}$  values of calcite (Figure 6c), which imply calcite precipitates from solutions with more vent fluid, and higher temperatures (calculated from  $\delta^{18}\text{O}$  and  $\Delta_{47}$ ) than the solutions leading to aragonite formation (Tables S3, S4). Overall, the geochemistry and of the petrography of calcite in the interior of the chimneys (Aquino et al., 2023) are consistent with calcite being the first carbonate mineral to precipitate in the actively venting structures. The occurrence of primary calcite is not uncommon in hydrothermal chimneys and veins. The interior of active hydrothermal chimneys from the Shinkai Seep Field in the southern Mariana forearc have also been reported to contain more calcite than aragonite (Okumura et al., 2016), and brucite-calcite hydrothermal veins were inferred to form in a fluid-mixing zone at a Lost City-type hydrothermal system at the Iberian Margin (Klein et al., 2015). Furthermore, chimneys from the Old City hydrothermal field (southwest Indian Ridge), like those from the LCHF, are composed of variable mixtures of brucite, calcite, and aragonite including a sample from an active structure comprising of brucite and calcite (Lecoeuvre et al., 2020).

Calcite is also abundant in chimneys from the eastern wall and in veins (Figure 3). Veins have calcite contents of up to 39% while samples from Calypso and other small chimneys in the eastern wall contain up to 24% and 58% calcite, respectively (Table S2). We suggest that the relatively low vent temperatures ( $\sim 30^\circ\text{C}$  for Calypso,  $11^\circ\text{C}$  for the vein on the carbonate cap), may explain the abundance of calcite. Mixing of small amounts of cold seawater with these relatively low temperature vent fluids will bring the fluid temperatures to  $<30^\circ\text{C}$  at which the minimum Mg/Ca ratio for aragonite precipitation is higher (Morse et al., 1997).

In addition to the effects of the variation in the mixing proportions of seawater and hydrothermal fluid, carbonate mineralogy and Sr geochemistry in the hydrothermal chimneys are also affected by the ageing of the structures. Observations from both active and inactive chimneys indicate that aragonite in young, active chimneys is converted to calcite by reaction with seawater over time (Aquino et al., 2023; Ludwig et al., 2006). Ludwig et al. (2006) and Vogel (2016) have shown that active chimneys have generally lower  $^{87}\text{Sr}/^{86}\text{Sr}$  and Sr concentrations (average 6328 ppm) than inactive structures where Sr concentrations increase (average 9015 ppm) and  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios approach those of seawater. During aging we also

observe the addition of carbonate particles (foraminifera and coccoliths) filtered from seawater circulating through the porous structures.

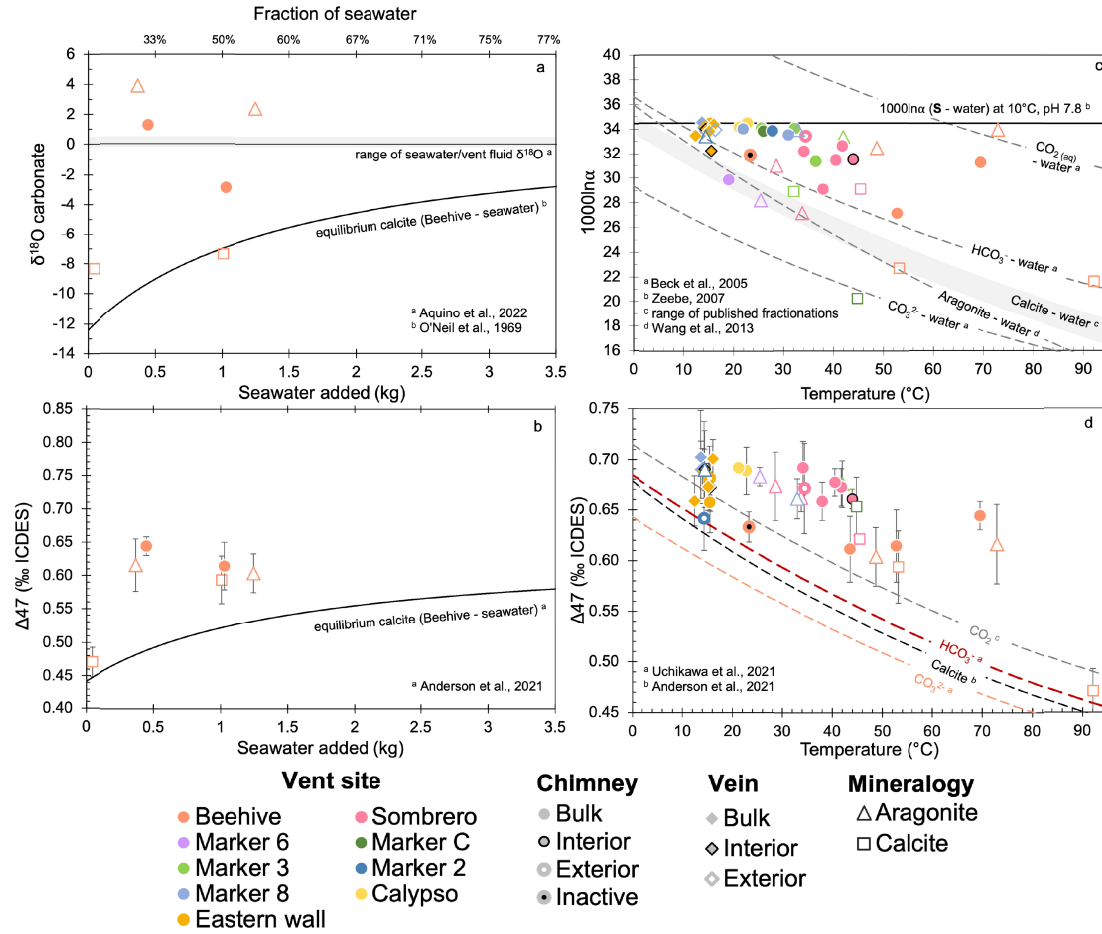
### 5.3 Multiple controls on $\delta^{13}\text{C}$ , $\delta^{18}\text{O}$ , and $\Delta_{47}$ values

Dissolved inorganic carbon (DIC) in the carbonate system exists in three main forms:  $\text{CO}_{2(\text{aq})}$  (aqueous carbon dioxide),  $\text{HCO}_3^-$  (bicarbonate), and  $\text{CO}_3^{2-}$  (carbonate ion), with the relative proportion of these species changing with pH (Zeebe & Wolf-Gladrow, 2001). The oxygen isotope fractionation between each species and water is different for each DIC species and is dependent on temperature (Figure 8c) (Beck et al., 2005). Therefore, the overall oxygen isotope fractionation between the sum of the DIC species (denoted as S, Figure 8c) and water is a function of both pH and temperature (Zeebe, 2007). At Lost City, carbonates are formed during mixing of seawater and vent fluids with different endmember pH, temperature, and chemistry (Kelley et al., 2005; Lang et al., 2012; Ludwig et al., 2006; Proskurowski et al., 2006). Therefore, multiple factors, including variable mixing proportions of vent fluid and seawater, and the initial pH, temperature, and stable isotope compositions of the vent fluid is expected to affect the compositions of the carbonates.

#### 5.3.1 Estimation of the temperature of the mixed fluid

The carbonates show a large range in  $\delta^{18}\text{O}$  and  $\Delta_{47}$  values (Figures 5, 6). Precipitation temperatures calculated from clumped and oxygen isotope analyses range from  $-7^\circ\text{C}$  to  $78^\circ\text{C}$  and  $-1^\circ\text{C}$  to  $76^\circ\text{C}$ , respectively (Tables S2, S3). These calculated temperatures are below the measured vent fluid temperatures (Table S1) which is expected as the carbonates precipitate from a mixture of vent fluid and seawater (Ludwig et al., 2006; see section 5.1). However, a significant number of samples yield precipitation temperatures below that of ambient seawater ( $\sim 10^\circ\text{C}$ ) and even below zero, indicating precipitation under disequilibrium conditions. We can estimate the temperature of the mixed fluid that produced each carbonate ( $T_{\text{est}}$ ) combining  $^{87}\text{Sr}/^{86}\text{Sr}$  derived seawater and vent fluid fractions (see section 5.1) with the measured exit temperatures of the vent fluids and the temperature of ambient seawater (Figure 6c, Table S1). For veins and chimneys where venting was not observed, we use the venting temperature of  $22^\circ\text{C}$ , measured in an eastern wall vein northeast of Marker 7 (Aquino et al., 2022). Using this approach, we calculate the expected  $\delta^{18}\text{O}$  and  $\Delta_{47}$  of a carbonate that precipitated in equilibrium with this mixed fluid (black line in Figure 8a, b for Beehive). Beehive carbonates have  $\delta^{18}\text{O}$  and  $\Delta_{47}$  values that are much higher than expected at

equilibrium (Figure 8a, b). Calcite, however, appears to precipitate closer to equilibrium than aragonite and the bulk sample.



**Figure 8.** Effect of seawater mixing and temperature to the equilibrium oxygen isotope and clumped isotope compositions. Expected (a)  $\delta^{18}\text{O}$  and (b)  $\Delta_{47}$  values of carbonate produced by mixing of seawater and Beehive vent fluids assuming equilibrium conditions. Also shown are  $\delta^{18}\text{O}$  and  $\Delta_{47}$  values of aragonite, calcite, and bulk samples from Beehive plotted against the fraction of seawater estimated from  $^{87}\text{Sr}/^{86}\text{Sr}$  in the carbonates. Note that for both  $\delta^{18}\text{O}$  and  $\Delta_{47}$ , samples from Beehive do not appear to have precipitated in equilibrium. (c) Oxygen isotope fractionation ( $1000\ln\alpha$ ) between carbonate and water (vent fluid-seawater solution) of the Lost City carbonate samples plotted against estimated temperature during carbonate precipitation ( $T_{\text{est}}$ ). Also shown are equilibrium fractionation lines between calcite and various DIC species and water. S refers to the overall oxygen isotope fractionation between the sum of the DIC species and water at the pH and temperature of the background seawater at Lost City. (d) Carbonate clumped isotope composition of the Lost City samples plotted

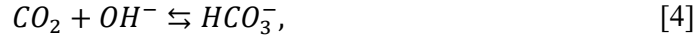
against estimated temperatures ( $T_{\text{est}}$ ). Also shown are  $\Delta_{47}$  values at equilibrium. Temperature is estimated from  $^{87}\text{Sr}/^{86}\text{Sr}$  derived vent fluid fractions and measured vent temperatures ( $T_{\text{est}}$ , see section 5.3.1). °Upper and lower limits for the calcite-water fractionation are from Coplen (2007) and O'Neil et al. (1969), respectively.

The oxygen isotope fractionation between the carbonates and water ( $1000\ln\alpha_{(\text{CaCO}_3\text{-H}_2\text{O})}$ ); calculated from the  $\delta^{18}\text{O}$  of the vent fluids and background seawater, Aquino et al., 2022) and  $\Delta_{47}$  values vs.  $T_{\text{est}}$  for the samples with  $^{87}\text{Sr}/^{86}\text{Sr}$  data are shown in Figure 8c, d. Also shown are the equilibrium oxygen isotope fractionation curves between water (i.e., mixture between seawater and vent fluids) and the different DIC species, the sum of all species (S), calcite, and aragonite. Most samples have oxygen and clumped isotope compositions far from the expected equilibrium values of calcite at their estimated formation temperature (Figure 8c, d). For  $1000\ln\alpha_{(\text{CaCO}_3\text{-H}_2\text{O})}$ , most of the data fall between the equilibrium  $\text{HCO}_3^-$  and  $\text{CO}_3^{2-}$  compositions at  $10^\circ\text{C}$ , the temperature of background seawater at Lost City. Exception to this are several calcite samples from Beehive and Marker C which show lower values. A similar observation is made for  $\Delta_{47}$ , but for fewer samples.

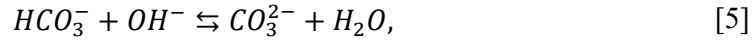
### 5.3.2 Disequilibrium precipitation

Calcium carbonate minerals often precipitate under non-equilibrium conditions in laboratory and most natural settings and thus display  $\delta^{18}\text{O}$  values that deviate from equilibrium (Daëron et al., 2019; Tripathi et al., 2015; Watkins et al., 2013). Disequilibrium, kinetic isotope effects are due to crystal growth reactions as well as the slow isotope exchange between DIC species and water relative to mineral precipitation (Daëron et al., 2019; Watkins et al., 2013; Zeebe et al., 1999). Carbonate minerals at Lost City precipitate very rapidly whereas equilibrium between DIC and  $\text{H}_2\text{O}$  may take hours to several days to be achieved especially at higher pH (Dietzel et al., 2009; Watkins et al., 2013; Zeebe et al., 1999; Zeebe & Wolf-Gladrow, 2001).

Oxygen atoms that are incorporated into the carbonate minerals at Lost City are sourced from seawater DIC and/or vent fluid-derived  $\text{OH}^-_{(\text{aq})}$ :



seawater    vent fluid



seawater    vent fluid



As described above, the carbonates record oxygen and clumped isotope compositions that are mostly higher than the equilibrium values for calcite (Figure 8c, d). Kinetic isotope effects during  $\text{CO}_2$  hydroxylation (equation 4) reported in terrestrial hyperalkaline systems (including natural springs and laboratory experiments) where atmospheric  $\text{CO}_2$  is the principal source of carbon produce strong depletions in  $^{18}\text{O}$  and  $^{13}\text{C}$  (e.g., Clark et al., 1992; Schwarzenbach, Lang, et al., 2013) which is the opposite of what we observe at Lost City. In contrast to terrestrial hyperalkaline springs, seawater DIC which is mostly in the form of  $\text{HCO}_3^-$  is the principal source of carbon at Lost City.

Quantitative precipitation of seawater DIC at  $10^\circ\text{C}$  and  $\text{pH} = 7.8$  will produce a carbonate with a  $1000\ln\alpha_{(\text{CaCO}_3\text{-H}_2\text{O})}$  value of 34.5 (Figure 8c). A number of samples, especially from relatively low temperature vents (e.g., eastern wall, Calypso, Marker 8), seem to record the  $\delta^{18}\text{O}$  of DIC of background seawater with little variability in  $1000\ln\alpha_{(\text{CaCO}_3\text{-H}_2\text{O})}$  across many vents (Figure 8c). Addition of warm ( $\sim 20$  to  $96^\circ\text{C}$ ) Ca-OH vent fluids to an isotopically equilibrated seawater at  $10^\circ\text{C}$  will increase the temperature and pH of the solution and will promote the forward reaction in equation 5 forming  $\text{CO}_3^{2-}$  that can be rapidly incorporated into a growing crystal prior to isotope equilibration at the new fluid temperature. Our data show, especially for the samples from the lower temperature vents, that seawater DIC behaves as a generally closed system, and that there is no or very limited reaction between DIC and  $\text{H}_2\text{O}$  or  $\text{OH}^-$  preceding carbonate precipitation. In addition, since Ca concentrations (10 to 40 mmol/kg) are significantly higher than DIC contents (0 to 2 mmol/kg), quantitative precipitation of the DIC may occur. Thus, the carbonates may directly record the  $\delta^{18}\text{O}$  and  $\Delta_{47}$  composition of the starting seawater DIC.

The samples from the relatively higher temperature vents (e.g., Beehive, Sombrero), on the other hand, display lower values which can be explained by a combination of multiple

processes (Figure 8c). Carbonates forming at relatively higher temperatures generally also form at relatively higher pH (Figure 7b). At higher pH, the oxygen isotope fractionation between DIC and water decreases (Zeebe, 2007) and the equilibration time for oxygen isotopes in DIC increases (Figure S3; Watkins et al., 2013; Zeebe, 1999). However, time required for oxygen isotope equilibration also decreases with increasing temperature (Watkins et al., 2013). At 96°C and pH = 8.9 (e.g., Beehive vent fluid), oxygen isotope equilibration time is relatively short and may take about 17 seconds (Figure S3). We suggest that the carbonates from higher temperature, higher pH vents have formed at conditions closer to equilibrium, which explains the overall lower  $\delta^{18}\text{O}$  compositions, especially for calcite formed from vent fluid-rich fluids. In contrast, carbonates formed at lower temperatures, either because they are from lower temperature vents or they formed from extensive mixing with seawater, need more time to achieve equilibrium (Figure S3).

In contrast to oxygen isotopes, carbon isotopes achieve equilibrium at time scales on the order of less than 30 seconds (Zeebe & Wolf-Gladrow, 2001). Serpentinite-hosted systems such as the Lost City are characterized by fluids that are DIC-poor because of precipitation of carbonate minerals along the flow paths at depth due to the increased pH and/or reduction to hydrocarbons (Cipolli et al., 2004; Delacour et al., 2008; Kelley et al., 2005; Monnin et al., 2014; Proskurowski et al., 2008; Ternieten et al., 2021a, 2021b). Thus, upon mixing with seawater, C is dominantly sourced from seawater DIC. As a result, carbonates collected from actively venting structures at Lost City predominantly record  $\delta^{13}\text{C}$  within the range of marine carbonates (-2‰ to +2‰; Wefer & Berger, 1991) (Table S2; Figure 5). However,  $\delta^{13}\text{C}$  values as low as -7‰ and as high as +13‰ have been previously reported from samples collected near the eastern wall (e.g., active chimney from Wall Marker H; vein near Marker 7; (Früh-Green et al., 2003; Kelley et al., 2005; Vogel, 2016). We did not observe such low  $\delta^{13}\text{C}$  values, instead our samples from the eastern wall generally record the highest  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values. Low  $\delta^{13}\text{C}$  values observed in the some carbonate samples have been explained by methane oxidation and/or incorporation of mantle-derived carbon (Früh-Green et al., 2003; Vogel, 2016).

An interesting aspect of the data is the strong correlation between  $\delta^{18}\text{O}$  and  $\Delta_{47}$  values ( $R^2 = 0.75$ ) which indicates a similar control on these isotope systems. As discussed above, most of the samples record the oxygen and carbon isotope composition of seawater DIC. In contrast, most of the samples have  $\Delta_{47}$  values that are higher than seawater bicarbonate  $\Delta_{47}$ , dominating the clumped isotope signal. We do not have an explanation for these elevated  $\Delta_{47}$ .



A significantly lower background seawater temperature than what has been measured for more than a decade at Lost City ( $\sim 10^{\circ}\text{C}$ , Aquino, Fröh-Green, Rickli, et al., 2022; Lang et al., 2012; Ludwig et al., 2006) is unlikely. It is worth noting that the  $\Delta_{47}$  of the individual DIC species, especially  $\text{HCO}_3^-$ , are not very well constrained. In particular, experimentally derived  $\Delta_{47}$  values for  $\text{HCO}_3^-$  determined by Tripathi et al. (2015) are 0.03‰ higher than values predicted from theory in the same study. Other factors may also affect the temperature dependence of the equilibrium  $\Delta_{47}$  values of the DIC species. For example, the presence of cations may slightly increase equilibrium clumped isotope compositions (Hill et al., 2020). Uchikawa et al. (2021) also noted a discrepancy in  $\Delta_{47}$  values of up to 0.106‰ and 0.081‰ for  $\text{HCO}_3^-$  and  $\text{CO}_3^{2-}$ , respectively, between values derived from *ab initio* calculations by Hill et al. (2020) which consider the presence of  $\text{Na}^+$  in the solution and experimental results of Tripathi et al. (2015). Overall, despite uncertainties in the clumped isotope composition of DIC, our data indicate that a large number of our samples record a similar  $\Delta_{47}$  composition which may imply a common DIC source pool for most samples, just like for  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ .

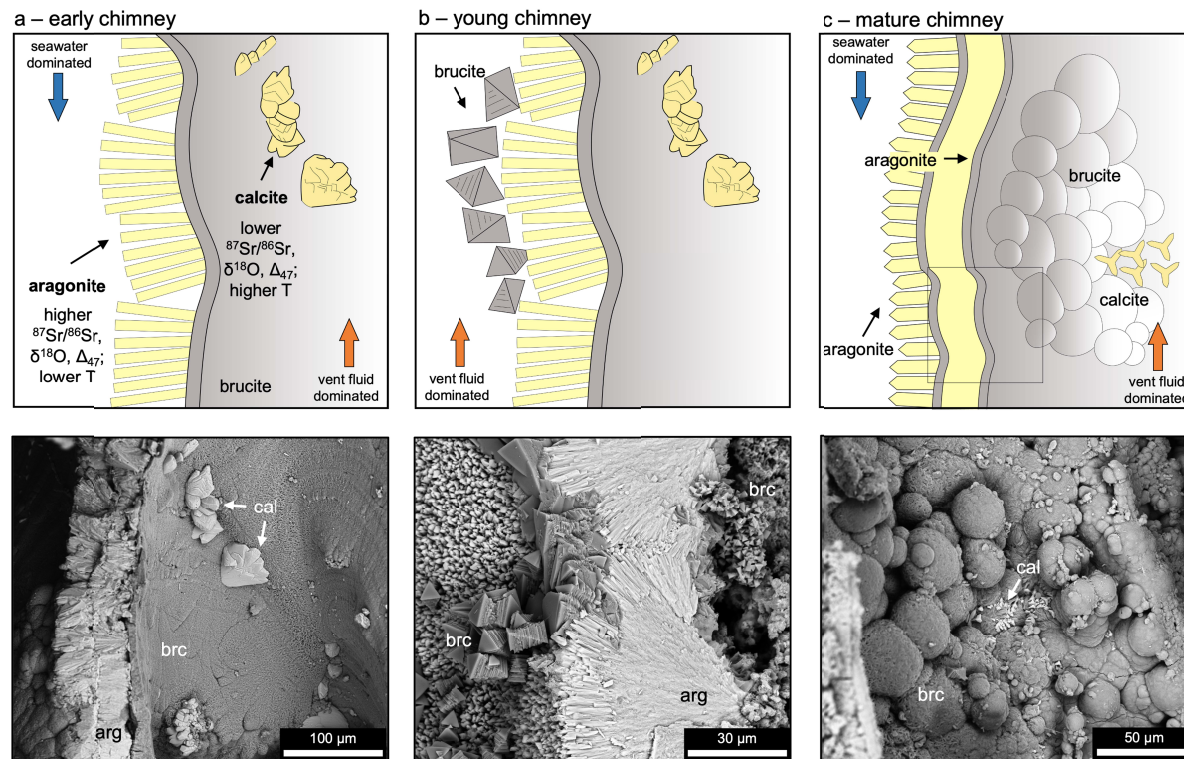
### 5.3.3 Other effects

There is no significant difference in  $\delta^{13}\text{C}$  between aragonite and calcite, although calcite tends to be slightly depleted in  $^{13}\text{C}$  relative to aragonite (Tables S3, S4; Figure 5b). This can be attributed to the lower isotopic fractionation between calcite and DIC compared to aragonite (Lécuyer et al., 2012; Romanek et al., 1992). Similarly, calcite records lower  $\delta^{18}\text{O}$  and  $\Delta_{47}$  values than aragonite (Tables S3, S4, Figure S2a, b, e), reflecting higher temperatures of precipitation. Calcite and aragonite crystals taken from the same sample show very heterogeneous isotopic compositions. Individual calcite crystals from Beehive have  $\delta^{18}\text{O}$  ranging from -12.0 to -3.1‰ and  $\delta^{13}\text{C}$  between -1.1 to 2.0‰ (Figure S2a). In general, there is a positive correlation between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ , especially in calcite (Figure S2a, b, d, e). We attribute this correlation to local Rayleigh fractionation during progressive carbonate precipitation. Carbonates generally have a lower  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  than the DIC they precipitate from, resulting in a slight enrichment in the isotopic composition of the remaining DIC. Carbonate crystals that subsequently precipitate from this residual mixed fluid are then slightly enriched in  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  relative to the earlier-formed carbonate. However, we observe a significantly larger variation in  $\delta^{18}\text{O}$  relative to  $\delta^{13}\text{C}$  ( $\sim 9$  vs.  $\sim 1$ ‰) in calcite. Vent fluids from Beehive have a pH of 10.7 at  $25^{\circ}\text{C}$  which is equivalent to about 0.45 mmol/kg of  $\text{OH}^-$ . We estimate that mixing with less than 20% of seawater is enough to consume the vent fluid derived  $\text{OH}^-$  and can locally decrease the pH. The oxygen isotope fractionation between

seawater and DIC increases with decreasing pH (Tripathi et al., 2015; Zeebe, 2007). Overall, Rayleigh fractionation combined with a local decrease in pH during progressive carbonate precipitation may explain the trend we observe in composition of the mineral separates collected from the same sample.

#### 5.4 Relationship between mineralogy and geochemistry of the Lost City chimneys

Here, we link our geochemical results with previous mineralogical and textural investigation of the Lost City chimneys (Aquino et al., 2023; Ludwig et al., 2006). The chimneys and veins are characterized by channel wall flow structures which serve as paleofluid flow paths. These mineral channel walls have been shown to be initially composed of brucite in young and active chimneys (Figure 9a; Aquino, Früh-Green, Bernasconi, et al., 2022; Ludwig et al., 2006). Brucite is stable at relatively higher pH (Pokrovsky & Schott, 2004) and precipitates only from solutions dominated by vent fluids predominantly in the interior of the chimneys (Figures 3, 7). With continued addition of seawater, carbonates may precipitate on the pre-existing brucite layer. Calcite will form at lower Mg/Ca ratios, i.e. from high proportions of vent fluid, in the interior of chimney structures and/or channel walls (Figures 7, 9). The coprecipitation of brucite and calcite from vent fluid-rich solutions is consistent with the observed association of these minerals at multiple vent locations (Aquino et al., 2023). Aragonite, in contrast, precipitates from solutions with higher Mg/Ca ratios and nucleates more commonly toward the exterior of brucite mineral channel walls from a seawater-dominated fluid (Figure 9a). Our interpretation that primary calcite and aragonite precipitate from distinct solutions is supported by the textural investigation of a fresh spire from the actively venting site Calypso. In this sample, both aragonite and calcite were observed to be present, but they precipitated on opposite sides of the brucite channel walls (Figure 9a). Mineral separates were unavailable for this sample because of the small size of the crystals ( $<50\text{ }\mu\text{m}$ ). Nevertheless, calcite in most samples has overall lower  $^{87}\text{Sr}/^{86}\text{Sr}$ ,  $\delta^{18}\text{O}$ , and  $\Delta_{47}$  values, indicating formation from more pristine hydrothermal fluid at higher temperatures, than coexisting aragonite (Figures 5b, 6). Overall, these suggest that the channel walls separate the chimney structure into distinct domains which may allow solutions of different compositions to flow and distinct carbonate minerals to precipitate.



674

675 **Figure 9.** Conceptual model of mineral precipitation in young, active chimneys modified from Aquino et al. (2023). (a) Chimneys initially form  
 676 brucite mineral membranes that bound fluid flow paths. Carbonate minerals precipitate on these channel walls. In vent fluid-dominated zones,  
 677 calcite is the carbonate mineral that forms, while aragonite precipitates in seawater-dominated zones (b-c) Multiple venting events and mixing  
 678 with seawater leads to the thickening of the channel wall by the precipitation of brucite on earlier formed aragonite or continued aragonite  
 679 precipitation, depending on the vent fluid-seawater mixing proportions. Box shows the approximate area drawn in a and b. Lower row shows  
 680 corresponding scanning electron images from Aquino et al. (2023).

Aquino, Früh-Green, Bernasconi, et al. (2022) also reported brucite minerals that nucleated upon earlier formed aragonite or calcite crystals. As brucite is only supersaturated at vent-fluid rich solutions, this brucite may form from a new influx of vent fluid (Figure 9b), showing a dynamic fluid circulation through the chimneys. The mineral channel walls thicken as mineral precipitation continues with the mineral assemblage being controlled by the changing composition of the fluid (Figure 9c). Overtime, as hydrothermal activity wanes, brucite and calcite are no longer thermodynamically stable and only aragonite precipitates. After hydrothermal fluid circulation stops, previously formed aragonite is converted to secondary calcite (Ludwig et al., 2006).

## 6 Conclusions

Hydrothermal chimneys collected from actively venting structures at the LCHF are composed of variable mixtures of brucite, calcite, and aragonite. Brucite is the first mineral to form upon mixing of the hydrothermal fluid with seawater forming an intricate channel system bounded by mineral membrane walls. During continued formation of the chimney, carbonate minerals precipitate on the preexisting channel walls from fluids of varying composition that impart a strong inhomogeneity to the mineralogy and isotope geochemistry of the towers. The mineralogy of the carbonate depends on the Mg/Ca ratio of the mixed fluid which is controlled by variation in the amount of vent fluid and seawater in the mixed fluid, which can be estimated using  $^{87}\text{Sr}/^{86}\text{Sr}$ . Vent fluid-dominated solutions have low Mg/Ca ratios and primarily precipitate brucite and calcite. In contrast, in seawater-dominated environments, chimney structures are often dominated by aragonite. Most carbonates, especially those from relatively lower temperature vents, record the  $\delta^{18}\text{O}$  of seawater bicarbonate at background seawater temperature ( $\sim 10^\circ\text{C}$ ) and have  $\delta^{13}\text{C}$  compositions within the range of marine carbonates. The clumped isotope compositions, on the other hand, are within a narrow range above the composition of seawater bicarbonate. Overall, our stable isotope data suggest that at Lost City, dissolved inorganic carbon (DIC) is often quantitatively precipitated during fluid mixing and equilibration between DIC and  $\text{H}_2\text{O}$  or  $\text{OH}^-$  does not occur. As a result, the carbonates record oxygen and clumped isotope compositions that do not reflect the temperature of the formation.

## Acknowledgments

We thank the captain and crew of the R/V Atlantis, ROV Jason, and the scientific party of the 2018 expedition AT42-01. Funding was provided by NSF awards OCE- 1536702

/ 1536405 / 1535962, the Swiss National Science Foundation project No. 200021\_163187, the Joint Institute for the Study of the Atmosphere and Ocean (JISAO) under NOAA Cooperative Agreement NA15OAR4320063 - Contribution No. 2021-1132, the Deep Carbon Observatory, and the Philippine Department of Science and Technology–Science Education Institute (DOST-SEI). We thank Madalina Jaggi and Stewart Bishop for assistance during the stable isotope analyses. We also thank James Watkins for providing the curves in Figure S3 and for discussions that helped improved the manuscript.

## Data Availability Statement

Individual analyses of samples are available at <https://doi.org/10.3929/ethz-b-000579720>.

## References

- Allen, D. E., & Seyfried, W. E. (2004). Serpentinization and heat generation: Constraints from Lost City and Rainbow hydrothermal systems. *Geochimica et Cosmochimica Acta*, 68(6), 1347–1354. <https://doi.org/10.1016/j.gca.2003.09.003>
- Amend, J. P., McCollom, T. M., Hentscher, M., & Bach, W. (2011). Catabolic and anabolic energy for chemolithoautotrophs in deep-sea hydrothermal systems hosted in different rock types. *Geochimica et Cosmochimica Acta*, 75(19), 5736–5748. <https://doi.org/10.1016/j.gca.2011.07.041>
- Anderson, N. T., Kelson, J. R., Kele, S., Daëron, M., Bonifacie, M., Horita, J., et al. (2021). A Unified Clumped Isotope Thermometer Calibration (0.5–1,100°C) Using Carbonate-Based Standardization. *Geophysical Research Letters*, 48(7), 1–11. <https://doi.org/10.1029/2020GL092069>
- Aquino, K. A., Früh-Green, G. L., Rickli, J., Bernasconi, S. M., Lang, S. Q., Lilley, M. D., & Butterfield, D. A. (2022). Multi-stage evolution of the Lost City hydrothermal vent fluids. *Geochimica et Cosmochimica Acta*, 332, 239–262. <https://doi.org/10.1016/j.gca.2022.06.027>
- Aquino, K. A., Früh-Green, G. L., Bernasconi, S. M., Bontognali, T., Foubert, A., & Lang, S. Q. (2023). *Biotic and abiotic mineral formation at the Lost City hydrothermal field*.
- Beck, W. C., Grossman, E. L., & Morse, J. W. (2005). Experimental studies of oxygen isotope fractionation in the carbonic acid system at 15°, 25°, and 40°C. *Geochimica et*

- Cosmochimica Acta*, 69(14), 3493–3503. <https://doi.org/10.1016/j.gca.2005.02.003>
- Bernasconi, S. M., Müller, I. A., Bergmann, K. D., Breitenbach, S. F. M., Fernandez, A., Hodell, D. A., et al. (2018). Reducing Uncertainties in Carbonate Clumped Isotope Analysis Through Consistent Carbonate-Based Standardization. *Geochemistry, Geophysics, Geosystems*, 19(9), 2895–2914. <https://doi.org/10.1029/2017GC007385>
- Bernasconi, S. M., Bergmann, K. D., Bonifacie, M., Meckler, A. N., Affek, H. P., Bajnai, D., et al. (2021). InterCarb: A Community Effort to Improve Interlaboratory Standardization of the Carbonate Clumped Isotope Thermometer Using Carbonate Standards. *Geochemistry, Geophysics, Geosystems*, 22(5), 1–25. <https://doi.org/10.1029/2020GC009588>
- Berner, R. A. (1975). The role of magnesium in the crystal growth of calcite and aragonite from sea water. *Geochimica et Cosmochimica Acta*, 39(4). [https://doi.org/10.1016/0016-7037\(75\)90102-7](https://doi.org/10.1016/0016-7037(75)90102-7)
- Bethke, C. M., Farrel, B., & Sharifi, M. (2020). *GWB Essentials Guide*. Aqueous Solutions, LLC Champaign, Illinois. Retrieved from <https://www.gwb.com/pdf/GWB14/GWBessentials.pdf>
- Breitenbach, S. F. M., & Bernasconi, S. M. (2011). Carbon and oxygen isotope analysis of small carbonate samples (20 to 100 µg) with a GasBench II preparation device. *Rapid Communications in Mass Spectrometry*, 25(13), 1910–1914. <https://doi.org/10.1002/rcm.5052>
- Cipolli, F., Gambardella, B., Marini, L., Ottonello, G., & Zuccolini, M. V. (2004). Geochemistry of high-pH waters from serpentinites of the Gruppo di Voltri (Genova, Italy) and reaction path modeling of CO<sub>2</sub> sequestration in serpentinite aquifers. *Applied Geochemistry*, 19(5), 787–802. <https://doi.org/10.1016/j.apgeochem.2003.10.007>
- Clark, I. D., Fontes, J.-C., & Fritz, P. (1992). Stable isotope disequilibria in travertine from high pH waters: Laboratory investigations and field observations from Oman. *Geochimica et Cosmochimica Acta*, 56(5), 2041–2050. [https://doi.org/10.1016/0016-7037\(92\)90328-G](https://doi.org/10.1016/0016-7037(92)90328-G)
- Coggon, R. M., & Teagle, D. A. H. (2011). Hydrothermal calcium-carbonate veins reveal past ocean chemistry. *TrAC - Trends in Analytical Chemistry*, 30(8), 1252–1268. <https://doi.org/10.1016/j.trac.2011.02.011>

- Coplen, T. B. (2007). Calibration of the calcite-water oxygen-isotope geothermometer at Devils Hole, Nevada, a natural laboratory. *Geochimica et Cosmochimica Acta*, 71(16), 3948–3957. <https://doi.org/10.1016/j.gca.2007.05.028>
- Daëron, M., Drysdale, R. N., Peral, M., Huyghe, D., Blamart, D., Coplen, T. B., et al. (2019). Most Earth-surface calcites precipitate out of isotopic equilibrium. *Nature Communications*, 10(1), 1–7. <https://doi.org/10.1038/s41467-019-08336-5>
- Von Damm, K. L. (1990). Seafloor hydrothermal activity: black smoker chemistry and chimneys. *Annual Review of Earth & Planetary Sciences*, 18, 173–204. <https://doi.org/10.1146/annurev.ea.18.050190.001133>
- David J. J. Kinsman. (1969). Interpretation of Sr<sup>+2</sup> Concentrations in Carbonate Minerals and Rocks. *SEPM Journal of Sedimentary Research*, Vol. 39. <https://doi.org/10.1306/74D71CB7-2B21-11D7-8648000102C1865D>
- Davis, K. J., Dove, P. M., & De Yoreo, J. J. (2000). The role of Mg<sup>2+</sup> as an impurity in calcite growth. *Science*, 290(5494), 1134–1137. <https://doi.org/10.1126/science.290.5494.1134>
- Delacour, A., Früh-Green, G. L., Bernasconi, S. M., Schaeffer, P., & Kelley, D. S. (2008). Carbon geochemistry of serpentinites in the Lost City Hydrothermal System (30°N, MAR). *Geochimica et Cosmochimica Acta*, 72(15), 3681–3702. <https://doi.org/10.1016/j.gca.2008.04.039>
- Deniel, C., & Pin, C. (2001). Single-stage method for the simultaneous isolation of lead and strontium from silicate samples for isotopic measurements. *Analytica Chimica Acta*, 426(1), 95–103. [https://doi.org/10.1016/S0003-2670\(00\)01185-5](https://doi.org/10.1016/S0003-2670(00)01185-5)
- Dick, J. M., & Shock, E. L. (2021). The Release of Energy during Protein Synthesis at Ultramafic-Hosted Submarine Hydrothermal Ecosystems. *Journal of Geophysical Research: Biogeosciences*, (1), 1–12. <https://doi.org/10.1029/2021jg006436>
- Dietzel, M., Tang, J., Leis, A., & Köhler, S. J. (2009). Oxygen isotopic fractionation during inorganic calcite precipitation - Effects of temperature, precipitation rate and pH. *Chemical Geology*, 268(1–2), 107–115. <https://doi.org/10.1016/j.chemgeo.2009.07.015>
- Eickenbusch, P., Takai, K., Sissman, O., Suzuki, S., Menzies, C., Sakai, S., et al. (2019). Origin of short-chain organic acids in serpentinite mud volcanoes of the Mariana convergent margin. *Frontiers in Microbiology*, 10(JULY), 1–21.

<https://doi.org/10.3389/fmicb.2019.01729>

Friedman, I., & O'Neil, J. R. (1977). Data of Geochemistry, Chapter KK: Compilation of Stable Isotope Fractionation Factors of Geochemical Interest. *U.S. Geological Survey Professional Paper 440-KK*, (Sixth Edition), 117.

Früh-Green, G. L., Kelley, D. S., Bernasconi, S. M., Karson, J. A., Ludwig, K. A., Butterfield, D. A., et al. (2003). 30,000 Years of Hydrothermal Vent Field. *Science*, *301*(July), 495–498.

German, C. R., & Seyfried, W. E. (2013). *Hydrothermal Processes. Treatise on Geochemistry: Second Edition* (2nd ed., Vol. 8). Elsevier Ltd.  
<https://doi.org/10.1016/B978-0-08-095975-7.00607-0>

Glein, C. R., Baross, J. A., & Waite, J. H. (2015). The pH of Enceladus' ocean. *Geochimica et Cosmochimica Acta*, *162*, 202–219. <https://doi.org/10.1016/j.gca.2015.04.017>

Hill, P. S., Schauble, E. A., & Tripathi, A. (2020). Theoretical constraints on the effects of added cations on clumped, oxygen, and carbon isotope signatures of dissolved inorganic carbon species and minerals. *Geochimica et Cosmochimica Acta*, *269*, 496–539.  
<https://doi.org/10.1016/j.gca.2019.10.016>

John, C. M., & Bowen, D. (2016). Community software for challenging isotope analysis: First applications of 'Easotope' to clumped isotopes. *Rapid Communications in Mass Spectrometry*, *30*(21), 2285–2300. <https://doi.org/10.1002/rcm.7720>

Kelley, D. S., Karson, J. A., Blackman, D. K., Früh-Green, G. L., Butterfield, D. A., Lilley, M. D., et al. (2001). An off-axis hydrothermal vent field near the Mid-Atlantic Ridge at 30° N. *Nature*, *412*(6843), 145–149. <https://doi.org/10.1038/35084000>

Kelley, D. S., Karson, J. A., Früh-Green, G. L., Yoerger, D. R., Shank, T. M., Butterfield, D. A., et al. (2005). A serpentinite-hosted ecosystem: The Lost City hydrothermal field. *Science*, *307*(5714), 1428–1434. <https://doi.org/10.1126/science.1102556>

Klein, F., Humphris, S. E., Guo, W., Schubotz, F., Schwarzenbach, E. M., Orsi, W. D., & Karl, D. M. (2015). Fluid mixing and the deep biosphere of a fossil Lost City-type hydrothermal system at the Iberia Margin. *Proceedings of the National Academy of Sciences of the United States of America*, *112*(39), 12036–12041.  
<https://doi.org/10.1073/pnas.1504674112>

Lang, S. Q., Butterfield, D. A., Schulte, M., Kelley, D. S., & Lilley, M. D. (2010). Elevated



concentrations of formate , acetate and dissolved organic carbon found at the Lost City hydrothermal field. *Geochimica et Cosmochimica Acta*, 74(3), 941–952.

<https://doi.org/10.1016/j.gca.2009.10.045>

Lang, S. Q., Fröh-Green, G. L., Bernasconi, S. M., Lilley, M. D., Proskurowski, G., Méhay, S., & Butterfield, D. A. (2012). Microbial utilization of abiogenic carbon and hydrogen in a serpentinite-hosted system. *Geochimica et Cosmochimica Acta*, 92, 82–99.

<https://doi.org/10.1016/j.gca.2012.06.006>

Lartaud, F., Little, C. T. S., Rafelis, M. De, Bayon, G., Dymont, J., & Ildefonse, B. (2011). Fossil evidence for serpentinization fluids fueling chemosynthetic assemblages.

<https://doi.org/10.1073/pnas.1009383108/->

[/DCSupplemental.www.pnas.org/cgi/doi/10.1073/pnas.1009383108](https://www.pnas.org/cgi/doi/10.1073/pnas.1009383108)

Launay, J., & Fontes, J. C. (1985). Les sources thermales de Prony (Nouvelle-Calédonie) et leurs précipités chimiques. Exemple de formation de brucite primaire. *Geologie de La France*, 83–99. Retrieved from <https://www.documentation.ird.fr/hor/fdi:19982>

Lecoeuvre, A., Ménez, B., Cannat, M., Chavagnac, V., & Gérard, E. (2020). Microbial ecology of the newly discovered serpentinite-hosted Old City hydrothermal field (southwest Indian ridge). *ISME Journal*, 15(3), 818–832.

<https://doi.org/10.1038/s41396-020-00816-7>

Lécuyer, C., Hutzler, A., Amiot, R., Daux, V., Grosheny, D., Otero, O., et al. (2012). Carbon and oxygen isotope fractionations between aragonite and calcite of shells from modern molluscs. *Chemical Geology*, 332–333, 92–101.

<https://doi.org/10.1016/j.chemgeo.2012.08.034>

Ludwig, K. A., Kelley, D. S., Butterfield, D. A., Nelson, B. K., Fröh-Green, G., Fru, G., & Fröh-Green, G. (2006). Formation and evolution of carbonate chimneys at the Lost City Hydrothermal Field. *Geochimica et Cosmochimica Acta*, 70(14), 3625–3645.

<https://doi.org/10.1016/j.gca.2006.04.016>

Ludwig, K. A., Shen, C. C., Kelley, D. S., Cheng, H., & Edwards, R. L. (2011). U-Th systematics and <sup>230</sup>Th ages of carbonate chimneys at the Lost City Hydrothermal Field. *Geochimica et Cosmochimica Acta*, 75(7), 1869–1888.

<https://doi.org/10.1016/j.gca.2011.01.008>

McCollom, T. M. (2007). Geochemical Constraints on Sources of Metabolic Energy for

- 867 Chemolithoautotrophy in Ultramafic-Hosted Deep-Sea Hydrothermal Systems.  
868 *Astrobiology*, 7(6), 933–950. <https://doi.org/10.1089/ast.2006.0119>
- 869 McCollom, T. M., & Seewald, J. S. (2007). Abiotic synthesis of organic compounds in deep-  
870 sea hydrothermal environments. *Chemical Reviews*, 107(2), 382–401.  
871 <https://doi.org/10.1021/cr0503660>
- 872 McKay, C. P., Porco, C. C., Altheide, T., Davis, W. L., & Kral, T. A. (2008). The possible  
873 origin and persistence of life on enceladus and detection of biomarkers in the plume.  
874 *Astrobiology*, 8(5), 909–919. <https://doi.org/10.1089/ast.2008.0265>
- 875 Meckler, A. N., Ziegler, M., Millán, M. I., Breitenbach, S. F. M., & Bernasconi, S. M.  
876 (2014). Long-term performance of the Kiel carbonate device with a new correction  
877 scheme for clumped isotope measurements. *Rapid Communications in Mass*  
878 *Spectrometry*, 28(15), 1705–1715. <https://doi.org/10.1002/rcm.6949>
- 879 Monnin, C., Chavagnac, V., Boulart, C., Ménez, B., Gérard, M., Gérard, E., et al. (2014).  
880 Fluid chemistry of the low temperature hyperalkaline hydrothermal system of Prony bay  
881 (New Caledonia). *Biogeosciences*, 11(20), 5687–5706. [https://doi.org/10.5194/bg-11-](https://doi.org/10.5194/bg-11-5687-2014)  
882 [5687-2014](https://doi.org/10.5194/bg-11-5687-2014)
- 883 Morse, J. W., Wang, Q., & Tsio, M. Y. (1997). Influences of temperature and Mg : Ca ratio  
884 on the mineralogy of CaCO<sub>3</sub> precipitated from seawater. *Geology*, 25, 85–87.  
885 [https://doi.org/10.1130/0091-7613\(1997\)025<0085:IOTAMC>2.3.CO;2](https://doi.org/10.1130/0091-7613(1997)025<0085:IOTAMC>2.3.CO;2)
- 886 Morse, J. W., Arvidson, R. S., & Lüttge, A. (2007). Calcium carbonate formation and  
887 dissolution. *Chemical Reviews*, 107(2), 342–381. <https://doi.org/10.1021/cr050358j>
- 888 Müller, I. A., Fernandez, A., Radke, J., van Dijk, J., Bowen, D., Schwieters, J., & Bernasconi,  
889 S. M. (2017). Carbonate clumped isotope analyses with the long-integration dual-inlet  
890 (LIDI) workflow: scratching at the lower sample weight boundaries. *Rapid*  
891 *Communications in Mass Spectrometry*, 31(12), 1057–1066.  
892 <https://doi.org/10.1002/rcm.7878>
- 893 Nier, A. O. (1938). The Isotopic Constitution of Strontium, Barium, Bismuth, Thallium and  
894 Mercury. *Physical Review*, 54(4), 275–278. <https://doi.org/10.1103/PhysRev.54.275>
- 895 O’Neil, J. R., Clayton, R. N., & Mayeda, T. K. (1969). Oxygen isotope fractionation in  
896 divalent metal carbonates. *The Journal of Chemical Physics*, 51(12), 5547–5558.  
897 <https://doi.org/10.1063/1.1671982>

- 898 Ohara, Y., Reagan, M. K., Fujikura, K., Watanabe, H., Michibayashi, K., Ishii, T., et al.  
899 (2012). A serpentinite-hosted ecosystem in the Southern Mariana Forearc. *Proceedings*  
900 *of the National Academy of Sciences of the United States of America*, 109(8), 2831–  
901 2835. <https://doi.org/10.1073/pnas.1112005109>
- 902 Okumura, T., Ohara, Y., Stern, R. J., Yamanaka, T., Onishi, Y., Watanabe, H., et al. (2016).  
903 Brucite chimney formation and carbonate alteration at the Shinkai Seep Field, a  
904 serpentinite-hosted vent system in the southern Mariana forearc. *Geochemistry,*  
905 *Geophysics, Geosystems*, 17(9), 3775–3796. <https://doi.org/10.1002/2016GC006449>
- 906 Palandri, J. L., & Reed, M. H. (2004). Geochemical models of metasomatism in ultramafic  
907 systems: Serpentinization, rodingitization, and sea floor carbonate chimney  
908 precipitation. *Geochimica et Cosmochimica Acta*, 68(5), 1115–1133.  
909 <https://doi.org/10.1016/j.gca.2003.08.006>
- 910 Palmer, M. R., & Edmond, J. M. (1989). The strontium isotope budget of the modern ocean.  
911 *Earth and Planetary Science Letters*, 92(1), 11–26. [https://doi.org/10.1016/0012-](https://doi.org/10.1016/0012-821X(89)90017-4)  
912 [821X\(89\)90017-4](https://doi.org/10.1016/0012-821X(89)90017-4)
- 913 Pisapia, C., Gérard, E., Gérard, M., Lecourt, L., Lang, S. Q., Pelletier, B., et al. (2017).  
914 Mineralizing filamentous bacteria from the prony bay hydrothermal field give new  
915 insights into the functioning of serpentinization-based subseafloor ecosystems. *Frontiers*  
916 *in Microbiology*, 8(JAN), 1–18. <https://doi.org/10.3389/fmicb.2017.00057>
- 917 Pokrovsky, O. S., & Schott, J. (2004). Experimental study of brucite dissolution and  
918 precipitation in aqueous solutions: surface speciation and chemical affinity control.  
919 *Geochimica et Cosmochimica Acta*, 68(1), 31–45. [https://doi.org/10.1016/S0016-](https://doi.org/10.1016/S0016-7037(03)00238-2)  
920 [7037\(03\)00238-2](https://doi.org/10.1016/S0016-7037(03)00238-2)
- 921 Proskurowski, G., Lilley, M. D., Kelley, D. S., & Olson, E. J. (2006). Low temperature  
922 volatile production at the Lost City Hydrothermal Field, evidence from a hydrogen  
923 stable isotope geothermometer. *Chemical Geology*, 229(4), 331–343.  
924 <https://doi.org/10.1016/j.chemgeo.2005.11.005>
- 925 Proskurowski, G., Lilley, M. D., Seewald, J. S., Früh-Green, G. L., Olson, E. J., Lupton, J. E.,  
926 et al. (2008). Abiogenic hydrocarbon production at lost city hydrothermal field. *Science*,  
927 319(5863), 604–607. <https://doi.org/10.1126/science.1151194>
- 928 Romanek, C. S., Grossman, E. L., & Morse, J. W. (1992). Carbon isotopic fractionation in

synthetic aragonite and calcite: Effects of temperature and precipitation rate.

*Geochimica et Cosmochimica Acta*, 56(1), 419–430. [https://doi.org/10.1016/0016-7037\(92\)90142-6](https://doi.org/10.1016/0016-7037(92)90142-6)

Schwarzenbach, E. M., Früh-Green, G. L., Bernasconi, S. M., Alt, J. C., & Plas, A. (2013). Serpentinization and carbon sequestration: A study of two ancient peridotite-hosted hydrothermal systems. *Chemical Geology*, 351, 115–133. <https://doi.org/10.1016/j.chemgeo.2013.05.016>

Schwarzenbach, E. M., Lang, S. Q., Früh-Green, G. L., Lilley, M. D., Bernasconi, S. M., & Méhay, S. (2013). Sources and cycling of carbon in continental, serpentinite-hosted alkaline springs in the Voltri Massif, Italy. *Lithos*, 177, 226–244. <https://doi.org/10.1016/j.lithos.2013.07.009>

Seyfried, W. E., Pester, N. J., Tutolo, B. M., & Ding, K. (2015). The Lost City hydrothermal system: Constraints imposed by vent fluid chemistry and reaction path models on subsurface heat and mass transfer processes. *Geochimica et Cosmochimica Acta*, 163, 59–79. <https://doi.org/10.1016/j.gca.2015.04.040>

Sleep, N. H. (2010). The Hadean-Archaeon environment. *Cold Spring Harbor Perspectives in Biology*, 2(6), 1–14. <https://doi.org/10.1101/cshperspect.a002527>

de Souza, G. F., Reynolds, B. C., Kiczka, M., & Bourdon, B. (2010). Evidence for mass-dependent isotopic fractionation of strontium in a glaciated granitic watershed. *Geochimica et Cosmochimica Acta*, 74(9), 2596–2614. <https://doi.org/10.1016/j.gca.2010.02.012>

Sun, W., Jayaraman, S., Chen, W., Persson, K. A., & Ceder, G. (2015). Nucleation of metastable aragonite CaCO<sub>3</sub> in seawater. *Proceedings of the National Academy of Sciences*, 112(11), 3199–3204. <https://doi.org/10.1073/pnas.1423898112>

Ternieten, L., Früh-Green, G. L., & Bernasconi, S. M. (2021a). Carbonate Mineralogy in Mantle Peridotites of the Atlantis Massif (IODP Expedition 357). *Journal of Geophysical Research: Solid Earth*, 126(12). <https://doi.org/10.1029/2021jb021885>

Ternieten, L., Früh-Green, G. L., & Bernasconi, S. M. (2021b). Distribution and Sources of Carbon in Serpentinized Mantle Peridotites at the Atlantis Massif (IODP Expedition 357). *Journal of Geophysical Research: Solid Earth*, 126(10). <https://doi.org/10.1029/2021JB021973>

- Thirlwall, M. F. (1991). Long-term reproducibility of multicollector Sr and Nd isotope ratio analysis. *Chemical Geology*, 94(2), 85–104. [https://doi.org/10.1016/S0009-2541\(10\)80021-X](https://doi.org/10.1016/S0009-2541(10)80021-X)
- Tripathi, A. K., Hill, P. S., Eagle, R. A., Mosenfelder, J. L., Tang, J., Schauble, E. A., et al. (2015). Beyond temperature: Clumped isotope signatures in dissolved inorganic carbon species and the influence of solution chemistry on carbonate mineral composition. *Geochimica et Cosmochimica Acta*, 166, 344–371. <https://doi.org/10.1016/j.gca.2015.06.021>
- Turekian, K. K., & Wedepohl, K. H. (1961). Distribution of the Elements in Some Major Units of the Earth's Crust. *GSA Bulletin*, 72(2), 175–192. [https://doi.org/https://doi.org/10.1130/0016-7606\(1961\)72\[175:DOTEIS\]2.0.CO;2](https://doi.org/https://doi.org/10.1130/0016-7606(1961)72[175:DOTEIS]2.0.CO;2)
- Uchikawa, J., Chen, S., Eiler, J. M., Adkins, J. F., & Zeebe, R. E. (2021). Trajectory and timescale of oxygen and clumped isotope equilibration in the dissolved carbonate system under normal and enzymatically-catalyzed conditions at 25 °C. *Geochimica et Cosmochimica Acta*, 314, 313–333. <https://doi.org/10.1016/j.gca.2021.08.014>
- Vogel, M. (2016). *Peridotite-hosted hydrothermal systems past and present: serpentinization, metasomatism and carbonate precipitation in modern and jurassic ultramafic seafloor*. <https://doi.org/https://doi.org/10.3929/ethz-a-010679089>
- Waite, J. H., Glein, C. R., Perryman, R. S., Teolis, B. D., Magee, B. A., Miller, G., et al. (2017). Cassini finds molecular hydrogen in the Enceladus plume: Evidence for hydrothermal processes. *Science*, 356(6334), 155–159. <https://doi.org/10.1126/science.aai8703>
- Watkins, J. M., Nielsen, L. C., Ryerson, F. J., & DePaolo, D. J. (2013). The influence of kinetics on the oxygen isotope composition of calcium carbonate. *Earth and Planetary Science Letters*, 375, 349–360. <https://doi.org/10.1016/j.epsl.2013.05.054>
- Wefer, G., & Berger, W. H. (1991). Isotope paleontology: growth and composition of extant calcareous species. *Marine Geology*, 100(207–248).
- White, W. M. (2009). 6. Aquatic Chemistry. *Geochemistry*, 210–257. <https://doi.org/http://doi.ieeeecomputersociety.org/10.1109/ISIP.2010.108>
- Zeebe, R. E. (1999). An explanation of the effect of seawater carbonate concentration on foraminiferal oxygen isotopes. *Geochimica et Cosmochimica Acta*, 63(13–14), 2001–

2007. [https://doi.org/10.1016/S0016-7037\(99\)00091-5](https://doi.org/10.1016/S0016-7037(99)00091-5)

Zeebe, R. E. (2007). An expression for the overall oxygen isotope fractionation between the sum of dissolved inorganic carbon and water. *Geochemistry, Geophysics, Geosystems*, 8(9), 1–7. <https://doi.org/10.1029/2007GC001663>

Zeebe, R. E., & Wolf-Gladrow, D. A. (2001). *CO<sub>2</sub> in Seawater: Equilibrium, Kinetics, Isotopes* (Vol. 65). Retrieved from <https://www.sciencedirect.com/bookseries/elsevier-oceanography-series/vol/65/suppl/C>

Zeebe, R. E., Wolf-Gladrow, D. A., & Jansen, H. (1999). On the time required to establish chemical and isotopic equilibrium in the carbon dioxide system in seawater. *Marine Chemistry*, 65(3–4), 135–153. [https://doi.org/10.1016/S0304-4203\(98\)00092-9](https://doi.org/10.1016/S0304-4203(98)00092-9)