Plain Language Summary  Carbonates form in diverse geological settings within the continental and oceanic lithosphere. They present a safe and permanent geological storage for CO$_2$, which is an important greenhouse gas. Thus, understanding the mechanism that drives carbonate formation will help to find a solution to stabilize atmospheric CO$_2$ and reduce global warming. This study uses petrographic and chemical investigations of drilled lithosphere samples from the Atlantis Massif to create a model for carbonate formation. The Atlantis Massif is located in the Atlantic Ocean close to the Mid-Atlantic-Ridge. It is characterized by a high amount of mantle rocks exposed at the seafloor. Mantle rocks react in contact with seawater to form serpentine and create an environment that leads to the transformation of CO$_2$ into carbonates. Our results demonstrate a high variability in carbonate chemistry as well as timing and temperature of formation within the Atlantis Massif. It shows that carbonate formation not only depends on the chemistry of the fluid from which it precipitates but also on the physical properties of the fluid. Furthermore, the study shows that Mg-rich carbonate (magnesite) formation can occur in an oceanic setting, which is predicted but rarely observed in natural settings.

1. Introduction

Mantle peridotites occur in many tectonic settings, such as at mid-ocean ridges (MORs), subduction zone fore- arcs, continental margins, and in ophiolites, and undergo hydrothermal alteration when exposed to aqueous fluids. Alteration processes have significant consequences for the thermal structure and rheology of the oceanic lithosphere, geochemical budgets of the ocean, and microbial processes at and below the seafloor (e.g., Früh-Green et al., 2004; Lister, 1972; Proskurowski et al., 2006; Wheat & Mottl, 2004). Seawater circulation through the oceanic crust occurs at rates such that circulation of the entire volume of the ocean through the oceanic crust requires less than 20 Myr (German & Von Damm, 2003; Wheat & Mottl, 2004). The resulting hydrothermal alteration of peridotites leads to the formation of hydrous silicates, oxides, sulfides, and carbonates (e.g., Frost, 1985).

The mechanisms that drive carbonate precipitation have been a subject of study for many years (Frost, 1985; Greenwood, 1967; Hansen et al., 2005; Hess, 1933; Koons, 1981; Korzhinskii, 1959; Schandl & Naldrett, 1992;
Trommsdorff & Evans, 1977). Carbonates can form in different geological settings under various pressure and temperature conditions. In shallow domains of the oceanic lithosphere, carbonates can precipitate during conductive heating of seawater, conductive cooling of hydrothermal fluids, mixing of hydrothermal fluids with seawater, and metasomatic replacement reactions (e.g., Alt et al., 2013; Bach et al., 2011; Grozeva et al., 2017; Klein et al., 2015; Schroeder et al., 2015; Schwarzenbach, Früh-Green, et al., 2013). More recently, natural carbonate formation has received significant attention as a mechanism to sequester atmospheric CO₂ and reduce global warming (Kelemen et al., 2011; Seifritz, 1990). Previous studies indicate that 5%–9% of the basement exposed along slow- and ultraslow-spreading MORs consist of hydrothermally altered peridotite (Cannat et al., 2010; Carlson, 2001) that can host carbonates. This makes carbonate formation in ultramafic rocks a theoretically important sink for carbon in the shallow oceanic lithosphere, but one that remains poorly quantified.

Several field, experimental and theoretical studies have been conducted with a main focus on carbonation of olivine to magnesite (Andreani et al., 2009; Bruni et al., 2002; Cipollri et al., 2004; Giammar et al., 2005; Hansen et al., 2005; Hövelmann et al., 2011; Kelemen & Matter, 2008; King et al., 2010; Klein & Garrido, 2011; Klein & McCollom, 2013; Lafay et al., 2014; Paukert et al., 2012; Peuble, Andreani, et al., 2015; Van Noort et al., 2013). Magnesite formation is a widespread alteration reaction of hydrothermally altered peridotites on land and can be found mainly within ophiolites in exhumed mélange rocks in paleo-subduction zones and orogenic belts (Barnes et al., 1973; Beinlich et al., 2012, 2020; Bohlke, 1989; Boschi et al., 2009; Dabitizias, 1980; Falk & Kelemen, 2015; Hansen et al., 2005; Naldrett, 1966; Robinson et al., 2005; Schandl & Wicks, 1991; Spandler et al., 2008; Ulrich et al., 2014), or within highly serpentinized peridotites that have been affected by the alteration with Mg- and HCO₃-rich meteoric waters at low temperatures (Barnes & O’Neil, 1969; Bruni et al., 2002; Kelemen et al., 2011; Kelemen & Matter, 2008; Paukert et al., 2012; Schwarzenbach et al., 2016). However, magnesite precipitation in oceanic peridotites has only been reported in a small number of studies (Gabrina et al., 2006). Instead, aragonite, calcite, and less commonly dolomite occur in hydrated mafic and ultramafic rocks in oceanic settings (Bach et al., 2011; Bonatti et al., 1974; Eickmann, Bach, Rosner, & Peckmann, 2009; Kelemen et al., 2011; Ludwig et al., 2006; Schroeder et al., 2015; Schwarzenbach, Lang, et al., 2013). Hydration and alteration experiments with ultramafic rocks show that dolomite or magnesite do not precipitate even in supersaturated systems in which dolomite and magnesite formation is thermodynamically predicted (Grozeva et al., 2017; Hövelmann et al., 2011). Grozeva et al. (2017) assumed that the Mg/Ca ratio of the reacting fluid at the mineral scale ultimately controls whether magnesite, dolomite, or calcite forms in seafloor serpentinization systems. In contrast, Hövelmann et al. (2011) suggest that kinetic factors impede the formation of Mg-bearing carbonates relative to Ca-carbonates. It is known that during hydrothermal alteration of mantle peridotites, variations in fluid flow significantly change the local solution chemistry, which in return influences carbonate precipitation. However, it remains unclear what factors directly control carbonate mineralogy in a natural system and what role fluid-rock interactions play during carbonate precipitation.

Here we present a study of serpentinized peridotites and metabasic rocks from the Atlantis Massif (AM) close to the Lost City Hydrothermal Field (LCHF) that were recovered during International Ocean Discovery Program (IODP) Expedition 357 (Früh-Green et al., 2017). The Lost City Hydrothermal Field located on the southern wall of the Atlantis Massif is considered an end-member for serpentinite-hosted, hydrothermally active systems at the slow-spreading Mid-Atlantic Ridge (MAR). It is an off-axis, low-temperature, peridotite-dominated submarine venting environment driven by seawater migration along deeply penetrating fault systems that facilitate hydration of the plutonic crust and serpentinization of the upper mantle (Kelley et al., 2001, 2005). Similar to other hydrothermal systems, the LCHF is characterized by fluids that are strongly enriched in Ca and depleted in Mg (e.g., Seyfried et al., 2015), whereby Ca is believed to be derived from dissolution of pyroxene (Douville et al., 2002; Gamo et al., 2001; Kelley et al., 2005) or related to silica metasomatism of surrounding mafic domains (Berndt et al., 1989; Bischoff & Dickson 1975). However, unlike fluids venting from unsedimented, basalt-hosted hydrothermal systems that are typically enriched in CO₂ (Lilley et al., 2003; Von Damm et al., 1998), the LCHF fluids are highly alkaline and depleted in CO₂, suggesting that they have lost oxidized carbon at depth by carbonate formation, either due to metasomatic replacement of minerals or precipitation within fractures (Proskurowski et al., 2008).

In this study, we investigated the carbonate population at the Atlantis Massif to constrain the physico-chemical properties and composition of the circulating fluid and to better understand the hydrothermal evolution of the Atlantis Massif. We performed detailed analyses of the chemical and petrological characteristics of the carbonates...
within the basement, highlighting the complexity of this unique system and allowing us to examine the crucial roles that protolith and fluid composition play in controlling carbonate precipitation during the interaction of hydrothermal fluids with mantle rocks.

2. Geological Setting and Sampling

The Atlantis Massif, located at 30°N along the slow-spreading Mid-Atlantic Ridge (MAR), is a 1.5–2 Myr old, dome-like massif forming the inside corner of the intersection between the MAR and the Atlantis Transform Fault (ATF) (Figure 1). The massif is interpreted as an oceanic core complex (OCC), comprised of lower crustal and upper mantle rocks uplifted and exposed by long-lived, low-angle detachment faulting (Blackman et al., 1998, 2002; Cann et al., 1997; Karson et al., 2006). Three lithologic domains can be distinguished: the corrugated central dome comprising mafic rocks ranging from olivine-rich troctolites to oxide gabbros; the variably altered peridotite-dominated southern wall with intermittent mafic plutonic intrusions, which is part of the Southern Ridge and defines the edge of the ATF; and the eastern basaltic block, which is interpreted as the hanging wall of the OCC. The summit peak rises to depths of less than 750 m and is located about midway along the length of the Southern Ridge. The top of the massif is covered by pelagic sediments, rubble, and sedimentary breccias (Blackman et al., 2002; Boschi et al., 2006; Früh-Green et al., 2003; Karson et al., 2006; Schroeder et al., 2002).

The Lost City Hydrothermal Field is located on a fault-bounded terrace just below the top of the southern wall near the summit. It consists of various active and inactive, up to 60 m tall, carbonate-brucite chimneys that vent low-temperature (40–95°C), alkaline (pH 9–11) fluids (Kelley et al., 2001, 2005; Ludwig et al., 2006; Seyfried et al., 2015). The vent fluids are characterized by low concentrations of silica, metals, and CO₂ and high concentrations of CH₄ (1–2 mmol/kg), H₂ (up to 15 mmol/kg), Ca (∼30 mmol/kg), and low-molecular-weight...
hydrocarbons. Hydrothermal circulation is believed to be driven by residual crustal heat and lithosphere cooling, with the composition of the fluids controlled by serpentinization reactions in the underlying peridotite. The hydrothermal activity has been ongoing for at least 30,000 years (Früh-Green et al., 2003; Kelley et al., 2001, 2005), and subsequent U-Th analyses indicated that this is a minimum age as some of the carbonate deposits yielded ages of up to 120,000 years (Ludwig et al., 2011).

2.1. IODP Expedition 357
IODP Expedition 357 drilled seventeen shallow boreholes (from 1.3 mbsf to maximum 16.4 mbsf) at nine sites along an east-west trending transect across the AM (Figure 1). Two sites are on the eastern end of the southern wall (Sites M0068 and M0075), three in the central section north of the LCHF (Sites M0069, M0072, and M0076), two on the western end (Sites M0071 and M0073), and two sites are to the north in the direction of the central dome and IODP Site U1309 (Sites M0070 and M0074). The recovered ultramafic rocks reveal a high and heterogeneous degree of serpentinization, metasomatic talc-amphibole-chlorite overprinting, and local rodosilatization (Früh-Green et al., 2017, 2018). Contacts between ultramafic and gabbroic rocks are marked by silica metasomatism with talc, tremolite, and chlorite (central IODP Holes M0068B, M0072B, and M0076B) in part replacing pyroxenes (Rouméjon, Früh-Green, et al., 2018). This indicates that hydration may have started at temperatures in the range of 400–500°C and that orthopyroxene alteration pre-dates the beginning of serpentinization of olivine (Rouméjon, Früh-Green, et al., 2018; Schroeder & John, 2004). Field and geophysical studies estimated that at least ~20% of the massif is completely serpentinized (Detrick & Collins, 1998; Früh-Green et al., 2003; Nooner et al., 2003). Two main stages of hydration can be distinguished; an initial stage of pervasive serpentinization, at 200–350°C (Boschi et al., 2008; Rouméjon, Früh-Green, et al., 2018; Rouméjon, Williams, et al., 2018), followed by a second stage of serpentinization at similar temperatures but higher SiO2 concentration dominated by focused fluid flow and leading to serpentine recrystallization and vein formation dominated by chrysotile or antigorite (Rouméjon, Früh-Green, et al., 2018).

This study focuses primarily on the five sites containing altered peridotite (Sites M0071, M0072, M0069, M0076, and M0068, Figure 2). The four remaining sites include foraminifera-bearing carbonate ooze, rubble blocks of variable lithologies, and sedimentary breccia with volcanic clasts and carbonate sediments (Früh-Green et al., 2017). The sites considered here have a core recovery ranging between 30% (Hole M0071C) and 75% (Hole M0069A), with an average recovery of 58%. Drill cores of IODP Expedition 357 are described in detail in Früh-Green et al., 2017, 2018 (Supporting Information), and Rouméjon, Früh-Green, et al., 2018 (Supporting Information).

3. Analytical Methods
Petrographic investigations were made on 37 doubly polished thin sections of representative samples distributed over five sites (7 holes), covering the diversity of rock types and alteration textures. The sample set consists of 24 peridotites (21 harzburgites and 3 dunites), nine mafic rocks (4 gabbroic rocks, 5 doleritic rocks), two talc-amphibole ± chlorite schists, and two brecciated samples (Table 1). Sample locations within the cores are indicated in Figure 2, together with a simplified lithostratigraphy and location of identified carbonate occurrences. A full identification for a sample is defined by IODP guidelines as: Expedition_Drilling-Platform, Site, Hole_Core, Coretype_Section_Interval in cm (e.g., 357_M0076B_9R_1_5-9). For simplicity, we have removed the expedition and drilling platform information which is the same for all samples (e.g., 76B_9R_1_5-9 is a shortened version of the sample above taken from Site M0076, core 9R during Exp. 357). We used clumped isotope data and radiocarbon ages from a parallel study that analyzed the identical set of carbonate veins (Ternieten et al., 2021).

3.1. Powder X-Ray Diffraction
For mineralogical identification of carbonate veins, rock samples were cleaned with 2-propanol and compressed air. In the first step, the surfaces of the veins were removed and discarded, followed by sampling of the vein interiors with a hand-held drill. In a second step, the obtained vein material was homogenized using an agate mortar previously cleaned with dichloromethane (DCM). Crystallographic analyses were made using a Bruker AXS D8 Advance Powder X-ray Diffractometer (XRD) equipped with a Lynxeye superspeed detector (Bruker Corporation, Billerica, United States) with Cu Kα X-ray radiation 2θ ranging between 10° to 60°, a voltage of 45 kV and...
a current of 40 mA, step size of 0.01°, and measurement time of 1 s per step. Quantification of the mineralogy of powdered samples was carried out using the program PowDil (Kourkoumelis, 2013) and the RRUFF database (Lafuente et al., 2015).

3.2. Optical and Electron Microscopy

Petrographic analyses were conducted using a polarizing microscope (Carl Zeiss Microscopy GmbH, Göttingen, Germany) on polished thin sections. Characterization, distribution, and textures of carbonates on polished thin sections were conducted using a cold cathode cathodoluminescence (CL) CL8200 Mk5-2 (Cambridge Image Technology Ltd, Hertfordshire, UK) with a Nikon microscope (Nikon, Japan) interfaced with a Zeiss Axiocam camera. Operation conditions were set at an accelerating voltage of 15 kV with a beam current of 250–300 μA. Major element composition of selected carbonates was determined using a five-spectrometer JEOL JXA-8200 and JEOL JXA-8230 Electron Probe Microanalyzer (EPMA) (JEOL, Akishima, Japan) operating in wavelength-dispersion mode with an accelerating voltage of 15 kV and a current of 10–5 nA. EPMA analyses were carried out on polished and carbon-coated thin sections. Beam diameter was 10, 5, and 2 μm, and count times varied from 20 to 60 s per element. Where possible, a defocused beam (10 μm) was used to reduce beam damage; where crystal sizes were small, a more focused beam (5 μm, 2 μm) was used. The program Probe for EPMA (Probe Software Inc., Eugene, United States) was used along with the time-dependent intensity (TDI) correction to counteract the effect of beam damage when a drift in the intensity was observed. The mean atomic number background correction was used to correct for the background (Donovan et al., 2016; Donovan & Tingle, 1996). Elements analyzed are Si, Mg, Al, Ba, Ca, Fe, Mn, and Sr, and concentrations were quantified based on natural and synthetic standards. CO₂ was calculated assuming stoichiometric mineral compositions, and the analyses were normalized to 100 wt%. For most elements with concentrations >1 wt%, average precision is better than 1%, and the average accuracy is better than 2%. Analyses of phases that yielded SiO₂ concentrations above 0.1 wt% have been excluded from our study, as they likely reflect mixed phases with surrounding serpentine or other silicate minerals. Because the crystal size of carbonates within the serpentine mesh cores was commonly very small,
| Site | Hole | Core | Section | Interval (cm) | Top | Bot. | Depth (mbsf) | Lithology | Carbonate occurrences | Carbonate mineralogy |
|------|------|------|---------|--------------|-----|------|-------------|-----------|----------------------|---------------------|
| 71   | A    | 1    | 1       | 45 48       | 0.45| 0.48| Serpentinitized dunite | Vein | Arg, Cal IIa, Cal IIb, Dol |
| 71   | B    | 3    | 1       | 27 29       | 3.71| 3.73| Serpentinitized harzburgite | None |
| 71   | C    | 1    | 1       | 46 50       | 0.46| 0.50| Serpentinitized harzburgite | None |
| 71   | C    | 2    | 1       | 67 77       | 3.35| 3.45| Serpentinitized harzburgite | None |
| 71   | C    | 2    | CC      | 5 9         | 3.70| 3.74| Serpentinitized harzburgite | None |
| 71   | C    | 3    | 1       | 10 13       | 5.12| 5.15| Serpentinitized harzburgite | None |
| 71   | C    | 5    | 1       | 20 23       | 7.58| 7.61| Serpentinitized harzburgite | None |
| 71   | C    | 6    | 1       | 70 74       | 10.33| 10.37| Dolerite | None |
| 69   | A    | 5    | 1       | 137 151     | 8.25| 8.39| Metadolerite | Void-filling grains | Cal* |
| 69   | A    | 6    | 1       | 127 130     | 9.87| 9.90| Metadolerite | None |
| 69   | A    | 8    | 1       | 14 18       | 11.46| 11.50| Metadolerite | Void-filling grains | Cal* |
| 69   | A    | 10   | 2       | 14 16       | 16.08| 16.10| Serpentinitized harzburgite | Vein | Arg |
| 72   | B    | 6    | 1       | 15 18       | 7.86| 7.89| Rodingitized gabbro | None |
| 72   | B    | 6    | 1       | 66 68       | 8.37| 8.39| Rodingitized dolerite | Void-filling grains | Cal |
| 72   | B    | 7    | 1       | 3 5         | 9.02| 9.04| Talc/amphi./chlorite schist | Void-filling grains | Cal* |
| 72   | B    | 7    | 1       | 55 58       | 9.54| 9.57| Talc/amphi./chlorite schist | Void-filling grains | Cal* |
| 72   | B    | 7    | CC      | 9 13        | 10.38| 10.42| Rodingitized gabbro | None |
| 72   | B    | 8    | 1       | 19 22       | 10.90| 10.93| Serpentinitized harzburgite | Vein, serpentine mesh core | Cal IIIb |
| 72   | B    | 8    | 1       | 69 72       | 11.40| 11.43| Talc/amphi./schist & serp. harzburgite | Vein, serpentine mesh core | Cal IIb |
| 72   | B    | 8    | 2       | 26 30       | 11.77| 11.81| Serpentinitized harzburgite | Vein, serpentine mesh core | Cal I, Cal IIa, Cal IIb |
| 72   | B    | 8    | 2       | 50 51       | 12.00| 12.02| Serpentinitized harzburgite | Serpentine mesh core | Cal* |
| 72   | B    | 8    | 2       | 67 70       | 12.18| 12.20| Serpentinitized harzburgite | Vein, serpentine mesh core | Cal I, Cal IIa, Cal IIb |
| 72   | B    | 8    | CC      | 0 5         | 12.28| 12.33| Serpentinitized harzburgite | Serpentine mesh core | Cal IIa, Cal IIb |
| 76   | B    | 3    | 1       | 78 83       | 4.22| 4.27| Serpentinitized dunite | None |
| 76   | B    | 5    | 1       | 55 59       | 7.28| 7.32| Serpentinitized dunite | Vein, serpentine mesh core | Cal III, Dol |
| 76   | B    | 5    | 1       | 77 80       | 7.50| 7.53| Metagabbro | Vein, void-filling grains | Cal III, Dol |
| 76   | B    | 6    | 1       | 63 65       | 8.61| 8.63| Serpentinitized harzburgite | None |
| 76   | B    | 6    | 1       | 118 122     | 9.16| 9.20| Serpentinitized harzburgite | Vein | Cal IIa, Cal IIb |
| 76   | B    | 7    | 1       | 60 62       | 10.32| 10.34| Serpentinitized harzburgite | None |
| 76   | B    | 8    | 1       | 90 94       | 12.05| 12.09| Serpentinitized harzburgite | None |
| 76   | B    | 8    | 1       | 112 117     | 12.27| 12.32| Serpentinitized harzburgite | Vein | Cal III, Cal IIb, Dol, Mgs |
| 76   | B    | 8    | 1       | 148 151     | 12.63| 12.66| Serpentinitized harzburgite | None |
| 76   | B    | 9    | 1       | 5 9         | 12.92| 12.96| Serpentinitized harzburgite | Vein | Arg, Cal III, Cal IIb, Dol, Mgs |
| 76   | B    | 10   | 1       | 91 111      | 15.50| 15.70| Metagabbro | None |
| 68   | B    | 1    | 1       | 10 13       | 0.10| 0.13| Serpentinitized harzburgite | Vein | Arg* |
| 68   | B    | 4    | 1       | 84 87       | 4.84| 4.87| Serpentinitized dunite | None |
| 68   | B    | 7    | 1       | 3 5         | 7.60| 7.62| Serpentinitized dunite | None |

Arg, aragonite; Cal, calcite; Dol, dolomite; Mgs, magnesite.
*Carbonate mineralogy based on optical and cathodoluminescence microscopy.
it is difficult to obtain good analyses by electron microprobe and twenty-three analyses of dolomite, replacing serpentine after olivine, with SiO$_2$ < 4 wt%, are included here.

4. Results

4.1. Basement Rocks

The carbonates are hosted in ultramafic, mafic, and clastic sedimentary rocks (Figure 3a–3c). Cores from the northwestern (M0071) and most eastern sites (M0068) contain varying proportions of ultramafic-, mafic, and sedimentary domains that have been interpreted as originating from mass-wasting and local faulting processes (Früh-Green et al., 2017; Rouméjon, Früh-Green, et al., 2018). Ultramafic and mafic host rocks were recovered from the central part of the southern wall (IODP Sites M0069, M0072, M0076) and are described as in-situ portions of the AM indicated by coherent, decimeter- to meter-long sections of cores, as well as consistent structural measurements (Früh-Green et al., 2017, 2018) (Figure 2). The central sites also make up the deepest boreholes, and the following discussion mainly focuses on observations from these cores.

Modal proportions of primary minerals in the peridotites range from 60%–100% olivine, 0%–40% pyroxene (including small amounts of clinopyroxene), and <1–2% spinel (Rouméjon, Früh-Green, et al., 2018). However, some serpentinized peridotites from Site M0072 have as much as 5 to 8 vol% magnetite and differ considerably from other typical serpentinized ultramafic domains (Figure 3a). Metamorphosed gabbroic and doleritic intrusions are largely altered to tremolite, chlorite, and/or talc, which locally overprint the serpentinization textures.

A significant observation in the basement rocks is the high variability in the occurrence, composition, and distribution of carbonate phases. The ultramafic samples, in particular, are characterized by complex networks of distinct generations of carbonates, specifically aragonite, calcite, dolomite, and magnesite. Combining textural features, chemical composition, timing, and formation temperatures allow the distinction of different generations of calcite (Table 2). In the following, we describe the most pronounced variations that lead to the classification of calcite I, calcite IIa, calcite IIb and calcite III. Dolomite also shows distinct generations in one sample but no clear correlation to major element composition. Thus dolomite is not further differentiated in this study.

4.2. Carbonate Occurrences

Carbonates in the basement of the southern wall occur as (a) void-filling carbonates, (b) carbonates replacing serpentine mesh cores after olivine, and (c) carbonate veins (Tables 1 and 2 and Figure 2). Carbonate veins were observed in both mafic and ultramafic rocks. Void-filling carbonates were only observed in mafic rocks, and carbonates replacing serpentine after olivine were only found within ultramafic rocks. Void-filling carbonates are typically found in metadolerites, talc-amphibole ± chlorite schists, and metagabbro. Calcite is the only carbonate phase and has grain sizes <20 μm, except for one metagabbro with carbonate grains <0.1 mm (Figures 3b and 3d). Void-filling carbonates are limited to the central sites at the southern wall (M0069, M0072, M0076). Carbonates replacing serpentine after olivine are dominantly calcite, rarely dolomite, and have an average grain size <50 μm (Figures 3e and 3f). Calcite within the serpentine mesh cores is restricted to Site M0072 and can occur as one or two calcite generations in the same sample, whereas dolomite is limited to Site M0076. Based on textural features, no clear distinction between whether the carbonates replace serpentine or directly olivine is possible.

4.2.1. Carbonate Veins

Veins are the most abundant carbonate occurrence in the Atlantis Massif basement and tend to be concentrated in the central drill sites closest to the LCHF (M0069, M0072, M0076) (Figure 2). The carbonate veins typically dissect grain boundaries, have kinked to irregular shapes, and show crosscutting or, more rarely, branching geometries (Figure 3g). They are less than 1 cm wide and are predominantly hosted by variably serpentinized peridotites and crosscut most other textures, indicating a late formation stage (Figure 4a). The abundance of carbonate veins is highly variable and can be as low as one vein or make up to half of a rock sample (∼40 vol%) as in some ultramafic rocks from Site M0076. An example of the size of a sample from the rock cores is given in Figure 4a. The veins are composed of magnesite, dolomite, calcite, and aragonite in variable proportions. At least two generations of carbonates occur in half of the samples and often coexist in the same vein. The samples 76B_8R_1_112-117 and 76B_9R_1_5-9 show at least four different generations (Figures 3h, 3i, 4b and 4c). The
altered mafic rocks are devoid of carbonate veins, except for sample 76B_5R_1_77-80 that contains a narrow (<50 μm) vein of calcite on the outside of the sampled rock piece (Figure 3b).

Serpentinites show the highest textural variability of carbonate veins at the AM. Magnesite, dolomite, and calcite veins are a few mm wide Figures (3a, 3h, 3i, and 4b–4i) and occur mainly at the central Sites M0072 and M0076, except for one calcite/dolomite vein found in a sedimentary breccia on top of Hole M0071A (71A_1R_1_45-48, Figure 3c). Magnesite veins are the least frequent and occur exclusively in association with dolomite. They are

Figure 3. Characteristic petrologic features of host lithologies and carbonates. (a) Thin section scan of fully serpentinized harzburgite with increased magnetite content, calcite I and II veins, and calcite II in serpentine after olivine. Two calcite I veins are cut by later serpentine veins (72B_8R_2_67-69.5). (b) Thin section scan of metagabbroic host rock with calcite I? vein and void-filling calcite I (76B_5R_1_77-80, close-up in Figure 3d. (c) Thin section scan of fully serpentinized dunite clast hosting fragments of dolomite and calcite II veins and aragonite with acicular growth within the fossiliferous carbonate sand (71A_1R_1_45-48). (d) Void-filling calcite in metagabbroic sample from Figure 3b (BSE image). (e) Calcite II replacing serpentinite after olivine (PPL, 72B_8R_CC_0-5, close-up in Figure 3f). (f) Calcite Ia and Iib in serpentine after olivine (BSE image). (g) A network of fibrous aragonite veins crosscut all other textures in the ultramafic host rock (PPL, 69A_10R_2_14-16). (h) Overview of a multigeneration dolomite vein in the ultramafic rock crosscut by later calcite III (PPL, 76B_9R_1_5-9, close-up in Figure 3i). (i) CL-image of the dolomite vein of Figure 3h containing multiple generations of dolomite, marked by Roman numerals, and other carbonate phases. Arg, aragonite; Cal, calcite; Chl, chlorite; Dol, dolomite; Amp, amphibole; Mgs, magnesite; Mag, magnetite; Srp, serpentine.
limited to the deeper sections of Hole M0076B (>12 mbsf) and show no distinct textural features to distinguish multiple magnesite generations (Table 2). Crosscutting relationships indicate that most magnesites pre-date all other carbonates (Figures 4d and 4e). Dolomite veins are more common at Site M0076; they are <5 mm wide and have precipitated at the rim of multiphase veins or are cut by serpentine and magnetite veins, suggesting an early stage of formation (Figures 3h, 3i, 4f, and 4g). One dolomite vein from sample 76B_9R_1_5-9 differed from other veins by the presence of multiple generations of dolomite with partly dissolved surfaces and pitted textures (Table 2, Figure 3i).

Calcite veins are most common at Site M0072, showing the highest textural variability, and often occur as two distinct generations within the same vein. At Site M0072, they constitute the earliest carbonate generation with intergrown or straight crystal contacts to the surrounding minerals (Figures 4b and 4c). Locally, calcite veins are intergrown with serpentine; these are highly deformed and exhibit secondary fluid inclusions, suggesting an earlier formation stage (Calcite I; Table 2, Figures 3a and 4h–4j). The fluid inclusions are composed of an aqueous liquid and a small vapor phase suggesting formation temperatures below 100°C. These veins are limited to the deeper section (>11 mbsf) at Hole M0072B. In two samples from Site M0076 (76B_5R_1_55-59, 76B_8R_1_112-117), calcite shows botryoidal textures (Calcite III; Table 2, Figures 4d and 4e), which is indicative of growth in open fractures.

The most frequent and volumetrically dominant carbonate phase is up to 1 cm wide aragonite veins that are observed throughout the ultramafic rocks at Hole M0069A, in the deeper section of Hole M0076B (>13 mbsf), in the top cores from Holes M0071A and M0068B, and on the altered surface of a deeper sample from Hole M0068B (Figure 2). The aragonite veins crosscut the primary rock fabric (Figures 3g and 4a), postdate all secondary fabrics, and are the last to be formed. The aragonite veins have three distinct crystal habits: fibrous crystals, acicular crystals, and microcrystalline aragonite (Figures 3c, 3g, and 4a). The long axes of acicular crystals grow into open spaces, and the fibrous crystals are predominantly oriented vertically to the borders of the veins. However, no distinct aragonite generations are observed; thus they will be no further discrimination of the aragonite, similar to magnesite (Table 2).

### 4.3. Carbonate Chemistry

Carbonates in the AM have pronounced variations in Mg, Fe, Sr, and Mn concentrations, which allow distinguishing different generations and indicate local variations in fluid chemistry, especially at Holes M0072B and M0076B (Table 2). Representative chemical analyses of all carbonate phases are presented in Table 3. Carbonates in the altered mafic rocks exhibit a limited range of major element compositions. Void-filling calcites have negligible MgO, and FeO and MnO concentrations that vary from 0 to 0.3 wt% and from 0 to 0.1 wt%, respectively.
The single identified calcite vein within the mafic rock samples only differs slightly by higher MgO concentrations of 2.4–3.2 wt% but also shows FeO and MnO concentrations <0.1 wt% (Table 3).

In contrast, the carbonates within the serpentinized ultramafic rocks show greater chemical variations. Magnesite veins have FeO concentrations ranging from 0.6 to 1.3 wt% and MnO concentrations from 2.4 to 4.1 wt% (Figures 5a and 5b). These FeO and MnO patterns are within the range of dolomite and calcite compositions.
## Table 3
Representative Analyses of Carbonates in Ultramafic and Mafic Rocks From the Southern Wall at the Atlantis Massif

| Sample       | Hole  | Host rock | Phase     | Occurrence | Vein | Void-filling | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Mesh core | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein | Vein |Vein

The analyses represent average values.
The dolomite veins have slightly more variable FeO and MnO concentrations compared to magnesite and lower FeO compared to calcites (FeO = 0 to 0.9 wt%, MnO = 0 to 6.0 wt%, Figures 5a and 5b). Sample 76B_9R_1_5-9, which contains a vein with multiple generations of dolomite (Figures 3h and 3i), has distinctly higher MnO contents at the rim than the center (MnO<sub>rim</sub> = 6.0 wt%, MnO<sub>center</sub> = 3.0 wt%; Figure 5b).

The calcite veins show pronounced variations and are classified into three groups. Calcite I is almost pure CaCO<sub>3</sub>; it constitutes the earliest carbonate phase and the only phase where we observed fluid inclusions (see Section 4.2.1). Calcite I has low SrO, MgO, FeO, and MnO concentrations of <0.2 wt%, similar to the void-filling calcites observed within the mafic rocks but distinct from other calcites within the ultramafic lithologies (Table 2, Figures 6a and 6c). Most calcite is classified as calcite II, whereby two subgroups can be distinguished based on the Mg concentration: Calcite IIa with MgO > 11.9 wt%, and calcite IIb with MgO < 8.3 wt% (Table 2). FeO and MnO contents vary from 0 to 3.3 wt% and from 0 to 7.3 wt%, respectively (Figures 6a–6d). Calcite IIa also shows higher Fe and Mn concentrations than calcite IIb in some samples at Site M0072, whereas such distinct variations between the subgroups were not observed at Site M0076. The calcite II subgroups show no unique chronology based on petrographic observations. Thus, depending on the location, either calcite IIa or calcite IIb represents the earlier phase.

Calcite III veins are defined by concentrations of MgO of 2.2–6.3 wt%, FeO of <0.2 wt%, and a wide range of MnO contents from 0 to 3.6 wt%. They are limited to Site M0076 and are the more common calcite type at this site where they represent the last or one of the latest carbonate phases, often occurring within open fractures as botryoidal crystals (Table 2, Figures 4d, 4e, 6b, and 6d). Aragonites throughout all cores have uniform compositions with MgO and MnO below the detection limit, only minor FeO concentrations <0.1 wt% and variable to high SrO concentrations (0.4–2.5 wt%, Table 3). In comparison, all other carbonate phases have SrO concentrations of less than 0.1 wt%.

The major element patterns of carbonates replacing serpentine mesh cores after olivine are indistinguishable from the vein carbonates. Calcite IIa and IIb in the serpentine mesh cores show FeO and MnO contents from 0.2 to 2.0 wt% and 0.8 to 7.3 wt%, respectively, with the highest MnO concentrations (>2.7 wt%) in calcite IIa. This is similar to calcite IIa in the veins; however, a distinct separation between the calcite subgroups based on their FeO content cannot be made (Figures 6e and 6f). No calcite I or calcite III, replacing serpentine after olivine, were identified. The dolomites replacing serpentine after olivine have uniform and much lower manganese concentrations than the calcites in the serpentine mesh cores (MnO = 1.7 to 2.3 wt%). FeO contents vary from 0.4 to 1.1 wt%, which is within the same range as dolomite in veins (Table 3).
Figure 6. FeO and MnO vs. MgO content of calcite veins hosted by serpentinized peridotites from the southern wall at the Atlantis Massif of Hole M0072B (a and c) and Hole M0076B (b and d), and of carbonates replacing serpentine mesh cores after olivine from Hole M0072B (e and f). Clumped isotope temperatures of carbonate formation from Ternieten et al. (2021). The dashed lines mark the difference between calcite IIa and IIb. Black symbols in Figures (e) and (f) represent the chemical composition of calcite veins from the same sample.
4.3.1. Carbonate Generations

Cathodoluminescence (CL) in carbonates results from trace-element substitution for Ca$^{2+}$ and Mg$^{2+}$ and is the most suitable method to identify different carbonate generations with only minor chemical variations. Although various trace elements are capable of influencing carbonate CL (Machel, 1985), Mn$^{2+}$ is the most critical “activator”, and Fe$^{2+}$, Fe$^{3+}$, Ni$^{2+}$, and Co$^{2+}$ are common “quenchers” (Götze, 2012). The intensity and color of luminescence are dependent on the relative proportions of Mn and Fe (Barnaby & Rimstidt, 1989). Additionally, the stimulated luminescence bands depend on the position in the crystal structure in which the Mn$^{2+}$ ion is integrated (Mg$^{2+}$ or Ca$^{2+}$ position) (Götze, 2012). Thus, the same MnO content can result in different CL. Identifying different carbonate generations based on CL analysis is particularly well illustrated in Figure 7, which shows simplified time logs of carbonate vein formation of three different samples.

Six carbonate generations with dolomite, magnesite, calcite III, and aragonite (Figure 7a) were identified in Sample 76B_9R_1_5-9 (Figures 3b and 3i). Multiple generations within a single vein characterize dolomite in this sample, and we identified three generations with partial dissolution and pitted textures between them and separated by magnesite precipitation. Dolomite veins from this sample showed significant variations in the MnO content (MnO$_{Dol}$ = 1.9 to 6.0 wt%) and only a limited range of FeO from 0.1 to 0.3 wt%, making variations in Mn$^{2+}$ concentration the most likely cause for the differences in luminescence (Figures 5a and 5b). However, no distinct geochemical signals can be assigned to the individual dolomite generations. Thus, in the upcoming discussion, dolomite will be discussed as a single generation. Sample 76B_8R_1_112-117 contains four different generations with magnesite, dolomite, calcite IIa and calcite III (Figure 7b). Similar to sample 76B_9R_1_5-9, dolomite is the volumetrically dominant carbonate phase in 76B_8R_1_112-117; however, the vein network is much finer, and no distinct generations within the dolomite can be identified. This coincides with the homogeneous MnO and FeO content of the dolomite (Figures 5a and 5b). However, both samples show a decrease in the FeO content with time from 0.6 to 1.0 wt% in the earliest generations of magnesite to <0.1 wt% in the latest aragonites and calcite III. Sample 72B_8R_2_67-70 (Figure 3a) is representative for Hole M0072B and contains three generations of calcite: calcite I, calcite IIa, calcite IIb (Figure 7c). The different generations observed by CL are also reflected by distinct major element compositions (Figures 6a and 6c). In addition, calcite

Figure 7. Simplified time log of carbonate vein formation of (a) 76B_9R_1_5-9, (b) 76B_8R_1_112-117, and (c) 72B_8R_2_67-70, based on optical and cold cathodoluminescence microscopy. The chronology is defined via crosscutting relationships and order within distinct, predominantly symmetrical veins. The Roman numerals (I to III) indicate different generations, letters (a) to (c) indicate a change in chemical composition or texture. Different generations occur intergrown, with erosive surface contacts or straight crystal contacts. The multiple dolomite generations in (a) are based on cold cathodoluminescence and only observed in one sample. $^{14}$C and clumped isotopes in (a) are average values of dolomite and magnesite, and calcite III and aragonite. The size of the veins did not allow sampling of distinct generations; clumped isotope temperatures of carbonate formation and radiocarbon age data from Ternieten et al. (2021).
IIa and IIb show a decreasing FeO content with time from FeO_{Cal IIa} = 1.2 wt% to FeO_{Cal IIb} = 0.3 wt%, similar to samples from Hole M0076B.

5. Discussion

Carbonation reactions in serpentinized peridotites are controlled by temperature, protolith composition, and activities of dissolved components and are particularly influenced by fluctuations in \( a\text{SiO}_2(aq) \) and \( a\text{CO}_2(aq) \). Figure 8 shows two activity-activity diagrams illustrating phase relations in the MgO-CaO-SiO\(_2\)-H\(_2\)O-CO\(_2\) system. It has long been recognized that serpentinized peridotites are highly reactive in the presence of dissolved CO\(_2\) (Johannes, 1967, 1969), and with increasing \( a\text{CO}_2(aq) \) the characteristic alteration assemblage of serpentine + brucite will successively be replaced by serpentine + magnesite, followed by talc + magnesite and finally quartz + magnesite (Figure 8a) (e.g., Boschi et al., 2009; Hansen et al., 2005). Even at low temperatures, moderate concentrations of CO\(_2\)\(_{aq}\) are sufficient to dissolve brucite in favor of carbonate. Thermodynamic models and experimental studies show that the \( a\text{SiO}_2(aq) \text{and} a\text{CO}_2(aq) \) are intimately linked and will ultimately affect silicate mineral stabilities, Mg/Ca ratios in the fluids, and thus the composition of the precipitating carbonate (e.g., Frost & Beard, 2007; Grozeva et al., 2017; Hövelmann et al., 2011). Low \( a\text{Mg}^{2+} \) will favor calcite precipitation, whereas high concentrations of CO\(_2\) and Mg\(^{2+}\) in the fluids will favor dolomite and magnesite precipitation (Figure 8) (Grozeva et al., 2017).

Serpentinized peridotites were recovered in all seven holes drilled across the AM southern wall. The exceptional core recovery preserved multiple carbonate generations that allow the reconstruction of the alteration history. One of the most remarkable features is the high variability in the composition and distribution of carbonate phases, including magnesite, dolomite, calcite, and aragonite. The diverse carbonate phases record progressive fluid infiltration and alteration during extension, unroofing, and uplift of the AM and the simultaneous evolution from pervasive- to localized serpentinization. Figure 9 shows an interpretative cross-section of the Atlantis Massif, highlighting the most critical processes affecting alteration and carbonate formation of the oceanic core complex. Previous studies (e.g., Boschi et al., 2006; Karson et al., 2006; Roumégoux, Früh-Green, et al., 2018) showed that the alteration of the AM is affected by seawater infiltration via grain boundary flow and a complex...
5.1. Alteration Heterogeneities Along the Southern Wall of the Atlantis Massif

Our detailed petrological, textural, and geochemical observations point to three main controlling factors that affect carbonate precipitation at the Atlantis Massif: (a) fluid composition and flux, (b) temperature of the system, and (c) presence of mafic intrusions. We focus on the three central sites (M0069, M0072, M0076), each providing distinct pieces of information about the complex history of fluid infiltration and alteration during exhumation of the massif.

5.1.1. Carbonate Formation in the Presence of Mafic Intrusions

Interlayers of mafic intrusions are most prevalent at Hole M0072B, which shows the highest degree of silica metasomatism, marked by talc-, amphibole- and chlorite-rich domains, and provides the best information about the influence of mafic intrusions on fluid and carbonate chemistry. This hole is characterized by a lack of
magnesite and dolomite in the ultramafic rocks, suggesting that Mg$^{2+}$ transport was strongly limited in the proximity of the metasomatic interlayered mafic intrusions and was locked into other silicate alteration assemblages (Figure 10, frame 1–3). At Hole M0072B, the initial carbonate phase is calcite I, which formed syn-serpentinization and is intergrown with serpentine in veins (Figure 10, frame 2). The lack of magnesite and dolomite suggests that the removal of Mg$^{2+}$ by silicates, resulting in low Mg/Ca ratios and favoring calcite precipitation (Figure 8b) (e.g., Hövelmann et al., 2011; Lumsdon et al., 1995; Peuble, Godard, et al., 2015). Alteration of mafic intrusions and an associated static replacement of serpentine by talc has been previously identified at the Atlantis Massif (Boschi et al., 2006; Rouméjon, Früh-Green, et al., 2018). Talc formation can result from either Si-addition or Mg-removal from the serpentinized peridotite. Previous studies (Bach et al., 2004; Boschi et al., 2006; Rouméjon, Früh-Green, et al., 2018) assumed that Si-addition rather than Mg-removal is responsible for talc formation, given that the alteration of mafic rocks releases significant amounts of silica compared with ultramafic rocks (Malvoisin, 2015; Wetzel & Shock, 2000). The metasomatism of gabbro and dolerites in the peridotites provide a source of Ca$^{2+}$ and SiO$_2$ from the dissolution of plagioclase and pyroxenes and removal of Mg$^{2+}$ through talc formation (Berndt et al., 1989; Bischoff & Dickson, 1975). In an experimental study at 40°C, Allen and Seyfried (2003) showed that high-temperature alteration could lead to a faster dissolution of clinopyroxene than olivine, providing an additional input of SiO$_2$ and Ca$^{2+}$ and lower input of Mg$^{2+}$. Elevated temperatures in the subsurface of the Atlantis Massif are indicated by the oxygen isotope composition of serpentine ($\delta^{18}$O$_{\text{recrystallized Srp}} = 1.6$ to 2.9 ‰, Rouméjon, Williams, et al., 2018), which corresponds to formation temperatures between 320 and 360°C (assuming $\delta^{18}$O$_{\text{fluid}} = 2.4$ ‰, Rouméjon, Williams, et al., 2018) and would be sufficient for clinopyroxene dissolution. Calcite

![Figure 10. Conceptual model for the evolution of alteration features in domains of the footwall along the detachment fault plane at the Atlantis Massif illustrating (1–3) alteration characteristics in Hole M0072B and (4–6) representative alteration textures in Hole M0076B. The combination of anisotropic thermal contraction and tectonic stresses in the brittle domain of the mantle leads to pervasive micro-fracturing (Rouméjon & Cannat, 2014), providing a fine network of pathways for serpentinization and carbonate precipitation. Hypothetical location of the illustrated lithologies can be seen in Figure 9. Hole M0072B is characterized by fully serpentinized peridotites with high magnetite concentration and carbonate precipitation in proximity to significant silica metasomatism of gabbroic and peridotite lithologies. The alteration of gabbroic lithologies provides a higher concentration of Ca in the hydrothermal fluid and leads to calcite I, Ila, and Iib formation. Calcite I represents carbonate formation during early hydration, whereas calcite Ila and Iib represent later formation. Carbonate precipitation at Hole M0076B is controlled by a high Mg/Ca ratio and zones of reduced flow due to structural heterogeneities, which leads to the early formation of magnesite and dolomite along microfractures. As exhumation and serpentinization proceed, alteration of clinopyroxene provides enough Ca for calcite Ila formation, and less altered seawater leads later to calcite III precipitation inside fractures. See text for a detailed discussion.](image-url)
I veins yield one of the highest carbonate formation temperatures of all carbonates across the AM (T$_\text{cal} = 145$–184°C) and indicate that temperatures in the hydrothermal system remained high during later alteration. We propose that the hydrothermal fluid in proximity to silica metasomatism is depleted in Mg$^{2+}$ and in combination with moderate $\delta$CO$_2$ and the dissolution of clinopyroxene with slower olivine serpentinitization was sufficient to maintain low Mg/Ca ratios in the hydrothermal fluid and prevent magnesite and dolomite precipitation. Calcite I veins have MgO, FeO, and MnO concentrations below the detection limit, which distinguish them from other carbonates at the AM and suggests that all dissolved Mg$^{2+}$, Fe$^{2+}$, and Mn$^{2+}$ were taken up by secondary silicate minerals (serpentine, talc, chlorite).

Calcite II is the second carbonate phase at Hole M0072B and occurs post-serpentinization within veins or replacing serpentinite mesh cores after olivine (Figure 10, frame 3). Calcite II has higher MgO contents, which may be linked to an increase in the Mg/Ca ratio resulting from an addition of unmodified seawater in the evolved hydrothermal fluids. A higher influx of seawater increase CO$_2(\text{aq})$ and Mg$^{2+}(\text{aq})$ concentrations, shifting the equilibrium toward the Mg-carbonate stability fields (Figure 8a). This may also lead to faster dissolution rates of the primary minerals of the mafic and ultramafic rocks, which can add additional SiO$_2$, Mg$^{2+}$, and Ca$^{2+}$ to the fluids. Calcite II is separated into two subgroups, a Mg-rich calcite IIa and a Mg-poor calcite IIb (Figure 6, Table 2). Calcite IIb precipitates from a fluid with lower Mg$^{2+}(\text{aq})$ concentration in areas where silica alteration minerals consumed more Mg$^{2+}(\text{aq})$ from solution. The Mg-rich calcite IIa precipitated from a fluid that was more influenced by serpentinization resulting in higher Mg$^{2+}$ concentrations, probably in more peripheral areas of the mafic intrusions. Some veins contain both calcite II subgroups but without a clear chronology. This may reflect changes in the influence of the mafic intrusions on the composition of the fluids over time by either change in the fluid pathway or variations in dissolution and/or formation rates of primary and secondary minerals.

Calcite IIa is not only characterized by higher MgO but also by higher FeO contents (Figure 6a). We suggest that high temperatures and higher fluid fluxes may result in elevated Fe$^{2+}$ concentrations in the fluid and consequently in the calcites. McCollom and Bach (2009) showed that moderate temperatures ($T = 100^\circ$C) and high water/rock ratios ($W/R > 2$) result in lower concentrations of Fe in the secondary minerals brucite and serpentine, leading to the formation of a higher amount of magnetite and H$_2$ (see also Malvoisin, 2015). This may also result in higher Fe$^{2+}$ concentrations in solution. In fact, at the Atlantis Massif, higher magnetite concentrations (5–8 vol%) are observed at sites influenced by mafic intrusions. In addition, higher amounts of magnetite have been associated with more oxidizing conditions caused by higher fluid flow (Andreani et al., 2009; Ulrich et al., 2014), supporting the argument of higher water/rock ratios and influx of seawater. High alteration temperatures and increased mixing of fresh seawater in the hydrothermal fluid is also supported by clumped isotope investigations of the same samples ($\delta^{13}$C$_{\text{Cal}} = -2.4 \%_o$, T$_\text{cal} = 145$–184°C; $\delta^{14}$C$_{\text{Cal-II}} = -0.9 \%_o$, T$_\text{cal} = 144°C$). Calcite II in serpentine mesh cores after olivine occurs either dispersed in the rocks or is associated with calcite II veins. The similarity of the element patterns suggests that dispersed calcite II formed simultaneously from a similar hydrothermal fluid as calcite II in the veins.

5.1.2. Magnesite and Dolomite Formation Within Oceanic Peridotites

Hole M0076B consists mainly of peridotites and exhibits the highest diversity of carbonate phases, including magnesite, dolomite, calcite, and aragonite. The ultramafic rocks of this hole show no evidence of silica metasomatism and thus provide constraints on carbonate formation in the absence of mafic intrusions (Figure 10, frame 4–6). The first generation of carbonate is either magnesite or dolomite forming syn- to post- serpentinization in veins and/or replacing serpentinite after olivine (Figure 10, frame 5). Dolomite, and especially magnesite precipitation, requires high Mg/Ca ratios in the fluids (e.g., Hövelmann et al., 2011; Lumsden et al., 1995; Peuble, Godard, et al., 2015). Thermodynamic calculations predict that magnesite should precipitate during the reaction of serpentine and CO$_2$-bearing aqueous fluids (e.g., Grozeva et al., 2017; Klein & Garrido, 2011). However, only a small number of studies have described magnesite in oceanic serpentinites (Gablina et al., 2006), whereas it is more common in altered ultramafic rocks on land (soapstone and listvenite; e.g., Beinlich et al., 2012; Falk & Kelemen, 2015). Previous studies suggest that magnesite in oceanic settings may form in Ca-poor ultramafic rocks, such as dunite or completely serpentinized peridotite, or where a continuous influx of CO$_2$-rich fluids results in silica activities in the fluid ($\alpha$SiO$_2(\text{aq})$) exceeding the stability of serpentine and consequently leading to high Mg/Ca ratios in the fluid (Figure 8a) (Barnes & O’Neil, 1969; Grozeva et al., 2017; Klein & Garrido, 2011). However, magnesite veins at the Atlantis Massif are found in highly serpentinized harzburgites with no evidence for serpentine instability. Indeed thermodynamic studies predict that relatively low $\alpha$SiO$_2(\text{aq})$ imposed
by serpentine and brucite formation and lower dissolution rates of the primary minerals, leads to stabilization of clinopyroxene and thus to lower Ca\(^{2+}\)(aq) in the fluids (Figure 8a) (e.g., Klein et al., 2013). Based on thermodynamic models (Grozeva et al., 2017), we argue that high CO\(_2\){aq} and moderate SiO\(_2\){aq} concentrations in the hydrothermal fluid have led to serpentinization of orthopyroxene and olivine while preserving primary clinopyroxene, which should have been sufficient to maintain high Mg/Ca ratios and favor the precipitation of Mg-rich carbonates (Figure 8b).

Magnesite is absent in most samples, even though high MgO content in calcites indicates high Mg\(^{2+}\) in solution. This suggests that CO\(_2\){aq} concentration alone does not control magnesite precipitation at the Atlantis Massif. An additional factor, such as confined flow pathways with limited fluid fluxes, may be necessary for magnesite formation. Pokrovsky and Schott (2002) and Schott et al. (2009) showed that a high fluid flux kinetically favors the formation of Ca-rich rather than Mg-rich carbonates due to different properties of Mg\(^{2+}\) and Ca\(^{2+}\) ions. The limiting step for carbonate crystallization is the dehydration of cations at the crystal surface. The lower the rate of water molecule exchange in the first hydration sphere of a metal ion, the lower the crystallization rate of a metal carbonate. Studies showed that the rate of water molecule exchange from the fluid into the cation hydration sphere is faster for Ca\(^{2+}\) than Mg\(^{2+}\) (e.g., Pokrovsky & Schott, 2002), which explains the different formation rates of Ca-rich (fast) and Mg-rich (slow) carbonates. Thus, dolomite will precipitate under conditions of moderate fluid fluxes when the fluid contains Ca even in a Mg-rich system, and magnesite will form in rock-dominated systems in zones of more-reduced fluid flow (Peuble, Andreani, et al., 2015). Dolomite replacing serpentine mesh cores after olivine occurs either dispersed in the rocks or is associated with dolomite veins. The similarity of the element patterns suggests that replacive dolomite formed at the same time from a similar hydrothermal fluid as dolomite in the veins.

The formation of magnesite and dolomite is followed by precipitation of calcite II at Hole M0076B (Figure 10, frame 6). The continuous hydration of the basement can lead to an increase in the activity of SiO\(_2\), which would lead to a dissolution of clinopyroxene (Figure 8a) and consequently to lower Mg/Ca ratios in the fluids favoring calcite formation. In addition, a potential higher influx of seawater caused by the progressive uplift of the massif may also result in an increase in fluid fluxes through brittle deformation, which favors the formation of Ca-rich carbonates (see magnesite discussion). Both calcite II (calcite Iia, calcite Iib) subgroups are observed (Figures 6b and 6d), indicating that high Mg\(^{2+}\) concentrations are maintained as the system evolves. The compositional variability of the hydrothermal fluids can be caused by changes in precipitation and/or dissolution rates, variations in the influx of fresh seawater, and changing temperatures. The high variability is also observed in the variable and high FeO and MnO contents of calcite II; whereby, a systematic difference between the subgroups cannot be distinguished.

The last carbonate phases are calcite III and aragonite precipitating in veins (Figure 10, frame 6): They have low FeO contents, which points to precipitation in equilibrium with cold hydrothermal fluids and mixing with unaltered seawater. In some cases, the botryoidal texture of calcite III (Figures 4d and 4e; Figure 5) indicates precipitation from a percolating solution within an open space. These spaces are often fractures crosscutting all minerals and deformation textures of the ultramafic rock, indicating that they formed after tectonic deformation of the serpentinite (Klein et al., 2015). Precipitation from cold, unaltered seawater mixing with evolved hydrothermal fluid is confirmed by clumped isotope temperatures of calcite III of Tserpentinite (Klein et al., 2018) precipitated from cold, unaltered seawater mixing with evolved hydrothermal fluid and consequent lower Mg/Ca ratios in the fluids favoring calcite formation. In addition, a potential higher influx of seawater caused by the progressive uplift of the massif may lead to an increase in fluid fluxes through brittle deformation, which favors the formation of Ca-rich carbonates (see magnesite discussion). Both calcite II (calcite Iia, calcite Iib) subgroups are observed (Figures 6b and 6d), indicating that high Mg\(^{2+}\) concentrations are maintained as the system evolves. The compositional variability of the hydrothermal fluids can be caused by changes in precipitation and/or dissolution rates, variations in the influx of fresh seawater, and changing temperatures. The high variability is also observed in the variable and high FeO and MnO contents of calcite II; whereby, a systematic difference between the subgroups cannot be distinguished.

5.1.3. Carbonate Formation at the Periphery of the Hydrothermal System

Peridotites at Hole M0069A are covered by ~7 m of sediments and dolerites but do not show evidence for silica metasomatism (Figure 2). Rouméjon, Früh-Green, et al. (2018) showed that the basement at this site is characterized by pervasive serpentinization with late overprinting with chrysotile and antigorite veins and recrystallization of serpentine. Thus, this site provides information on the alteration at the periphery of an active hydrothermal system, which is only affected by the first stage of serpentinization.
The only carbonate occurrence is aragonite veins, which are the volumetrically dominant carbonate phase at the AM. Geochemical and textural features are comparable to aragonite from sites in the vicinity of the hydrothermal system. Aragonite formed at low temperatures (T_{\text{AM}} = 4–7^\circ\text{C}) from fairly unmodified seawater (\(^{\delta^{13}}\text{C}_\text{Arq} = -0.2\) to +1.6 ‰; \(^{\delta^{13}}\text{C}_{\text{seawater}} = \sim 0\) ‰, Zeebe & Wolf-Gladrow, 2001). At such conditions, aragonite precipitation is favored over calcite due to high Mg\(^{2+}\) and sulfate concentrations (Burton, 1993; Eickmann, Bach, & Peckmann, 2009). The lack of carbonate veins within the dolerite, and the relative high vein density in the peridotites below it, strengthens previous interpretations that dolerite domains constitute impermeable barriers and fluids are channelled along the walls of the intrusion (Rouméjon & Cannat, 2014).

5.2. Manganese Concentration in Carbonates From the Atlantis Massif

A unique feature of the carbonates within the basement of the Atlantis Massif is the particularly high MnO concentrations. Carbonates are, in general, a favorable sink for Mn (Deer et al., 1992), and hydrothermal systems are important sources of manganese (e.g., Charlou et al., 2002; Klinkhammer et al., 1977). Laboratory studies have shown that seawater tends to leach manganese from mid-ocean ridge basalts at elevated temperatures (T > 200°C) and pressures (P > 500 bar), causing an increase of Mn\(^{2+}\) in solution (Bischoff & Dickson, 1975; Mottl et al., 1979). Previous studies of calcite cements formed in shallow fresh-water aquifers provide evidence that the uptake of Mn\(^{2+}\) in carbonates is mainly dependent on Eh, which controls Mn solubility (Barnaby & Rimstidt, 1989; Drigoole & Walter, 1990). Most marine low-temperature carbonates have MnO contents of \sim 1 wt% (e.g., Eickmann, Bach, Rosner, & Peckmann, 2009; Picazo et al., 2020). At the Atlantis Massif, carbonates exhibit remarkably high and variable MnO concentrations, and we argue that this is caused by a change in the redox state of the hydrothermal fluid. The high MnO contents would indicate a low redox state, typical of serpentinizing environments, and a shift to lower MnO would indicate a progressive increase in Eh. Most carbonate veins at the AM do not show a systematic change with their sequence of precipitation except for a thicker dolomite vein. In this vein, early dolomites have higher MnO (MnO_{\text{max, Dol}} = 6 wt%) than younger generations (Figure 5b), which may indicate a change to higher oxygen fugacity of the fluid over time. This change in fluid composition is also indicated by dissolved surfaces and pitted textures in early dolomites (Figure 3i).

Carbonates with extremely high MnO contents (max. 9.5 wt% Mn; Schroeder et al., 2015) have been previously reported at the Mid-Atlantic Ridge. Carbonates from the Ocean Drilling Program (ODP) Site 1275 (35 km north of the 15°20’ Fracture Zone and 30 km west of the MAR axis) have high MnO contents and were correlated to an early, high-temperature (T_{\text{max}} = 174°C) stage of formation. These carbonate veins occur within gabbros and troctolites, which most likely were the source of the elevated Mn. At the AM, carbonates are less common in the mafic rocks, and these have MnO contents below 0.1 wt%. However, we do observe higher MnO (MnO_{\text{max, M0072B}} = 7.3 wt%) in carbonates in proximity to mafic intrusions, compared to carbonates formed in the periphery to mafic intrusions. We propose that the high MnO concentrations are consistent with highly reducing serpentinizing fluids at the AM, leaching high amounts of Mn\(^{2+}\) from the interlayered gabbroic bodies, and transport to domains where cooling of hydrothermal fluids and or mixing with seawater leads to the formation of carbonates.

5.3. Carbonates as a CO\(_2\) Sink Within the Oceanic Lithosphere

Recently, carbonate formation in ultramafic rocks has received significant attention as a mechanism to sequester atmospheric CO\(_2\) (e.g., Kelemen & Matter, 2008), and studies of various ophicalcites showed that carbon is stored over millions of years (Bachu et al., 1994; Kelemen et al., 2011; Schwarzenbach, Früh-Green, et al., 2013; Seifritz, 1990). Studies of modern and ancient serpentinites estimate that up to 10 wt% total inorganic carbon is fixed as carbonates during serpentinization and carbonate precipitation and that approximately 10% of all carbon that is cycled through the ocean is incorporated into the oceanic lithosphere during alteration of oceanic crust (Alt et al., 2013; Schwarzenbach, Früh-Green, et al., 2013). To better evaluate the potential of ultramafic rocks as long-term storage for CO\(_2\), quantitative information on the sources and sinks of carbon is necessary. One critical factor contributing to the continuing uncertainty on this topic is the difficulty in obtaining samples directly from subsurface environments near mid-ocean ridges to analyze carbon compounds. Consequently, most estimations are based on information inferred from analysis of hydrothermal fluids discharged at the seafloor. Most of these hydrothermal vents have CO\(_2\) concentrations that are substantially higher than the amount of dissolved inorganic carbon (DIC: CO\(_2\)\(_{\text{aq}}\) + HCO\(_3^-\) + CO\(_3^{2-}\)) in seawater. However, \(^{13}\text{C}\) and \(^{14}\text{C}\) isotope analyses of CO\(_2\) in hydrothermal fluids from the Endeavor system (a high temperature basalt-dominated hydrothermal system) indicates that
most of the seawater DIC is removed from circulating fluids and stored as carbonates within the basement before discharge at hydrothermal vents on the seafloor (Proskurowski et al., 2004). It is assumed that this is true for most hydrothermal systems (McCollom, 2008; and references therein).

In contrast to basalt-hosted hydrothermal systems, the vent fluids from the Lost City hydrothermal system are highly depleted in DIC (Proskurowski et al., 2008), which leads to the assumption that the AM should contain significant carbonate deposits. Thus, one of the main goals of IODP Expedition 357 was to investigate the distribution of carbonate deposits within the basement of the Atlantis Massif. Our results show that the amount of inorganic carbon in the shallow basement holes is relatively low. Carbonates are almost exclusively observed close to the active venting LCHF and rarely at other locations along the southern wall. The distribution of carbonates throughout the cores from the central Sites (M0069, M0072, M0076) is heterogeneous, with an overall carbonate vein abundance estimated at ~2 vol%. However, because the maximum depth reached was 16.4 mbsf, we do not know how high the concentrations are at depth. We propose that most of the DIC removal occurred during deeper circulation of the fluid and that shallow parts of the AM record carbonate precipitation associated with present-day hydrothermal circulation and driven by localized shallow fluid circulation along fractures.

Additionally, the AM is a relatively young exhumed mantle sequence. Our study provides evidence that CO$_2$ uptake by low-temperature aragonite formation is generally low in young serpentinitized mantle but may increase if fracturing and veining continue for millions of years. A low abundance of low-temperature aragonite veins in young serpentinitized peridotite is also observed at ODP Site 895 at Hess Deep (Blusztajn & Hart, 1996), ODP Site 920 south of the Kane Fracture Zone (Alt & Shanks, 1998), and from ODP Site 175 at the Mid-Atlantic Ridge (Schroeder et al., 2015). The data presented here provide information on the fate of carbon in the oceanic lithosphere and will help to better evaluate whether hydration and serpentinization of oceanic crust is a globally significant sink for CO$_2$. Further investigations of deeper drill holes and older oceanic core complexes would be necessary.

6. Conclusions

We analyzed the hydration of mantle rocks in seven drill holes at five sites across the southern wall of the Atlantis Massif recovered during IODP Expedition 357. We developed a conceptual model for the genesis of carbonates in oceanic lithosphere. Detailed geochemical and petrographic analyses reveal variable alteration features between sites that represent different portions of the footwall exhumed in the vicinity of the detachment fault zone. Three characteristic types of carbonate occurrences, including the minerals magnesite, dolomite, calcite, and aragonite, could be identified. All occurrences formed close to the actively venting LCHF under different conditions and characteristic types of carbonate occurrences, including the minerals magnesite, dolomite, calcite, and aragonite, could be identified. All occurrences formed close to the actively venting LCHF under different conditions and were controlled by (a) fluid composition and flow, (b) temperature, and (c) influence of mafic intrusions. SiO$_2$$_{2(aq)}$, CO$_2$$_{2(aq)}$, and Mg$^{2+}$ concentrations of the hydrothermal fluid have particularly significant impacts on the carbonate compositions.

Even though carbonates within this oceanic hydrothermal system indicate ubiquitous high Mg$^{2+}$ concentrations, only limited amounts of magnesite and/or dolomite are formed, suggesting that kinetic factors impede the precipitation of Mg-rich carbonates relative to Ca-rich carbonates. This supports the hypothesis that both a high Mg/Ca ratio and confined flow are necessary to precipitate Mg-carbonates. In addition, our study indicates that lower Mg$^{2+}$ concentrations in solution characterize hydrothermal fluids in proximity to mafic intrusions. This lower Mg$^{2+}$ in solution in proximity to mafic intrusions in combination with the necessary lower fluid flow may limit magnesite formation in the oceanic lithosphere since we know that hydrothermal systems at mid-ocean ridges often contain mafic sequences and are, in general, characterized by high fluid flow (Bach et al., 2011). Therefore, magnesite precipitation in oceanic settings is likely limited to domains of ultramafic rocks without major mafic intrusions and less rapid upflow of hydrothermal fluids.

Our study shows that Ca-Mg carbonates, as a significant part of the process of hydration and alteration of oceanic peridotites, represent an important tool to reconstruct alteration histories of serpentinitized peridotites. Moreover, Ca-Mg carbonates potentially represent a significant sink of carbon from seawater and/or the mantle that would otherwise reenter/enter the oceans. Thus, carbonate-containing oceanic serpentinites entering subduction zones potentially represent an additional carbon pool that can be transported and recycled into the mantle. Radiocarbon analyses indicate that shallow carbonates in the AM basement precipitated over a short period of time from 37,000 to 23,000 years, corresponding to the early stages of the Lost City hydrothermal system (Früh-Green
et al., 2003). The active vent fluids at Lost City still have highly depleted DIC contents, which suggests that carbonate is continuing to be formed at deeper levels of this off-axis hydrothermal system.

The results of our study are of importance not only for the analyses of ancient and present-day carbonation systems, but they may also be relevant for the interpretation of artificial carbonation of serpentinite at the seafloor, on land, or in laboratory setups. We show that thermodynamically predicted magnesite precipitation occurs locally within oceanic serpentinitized and highlights the high heterogeneity of hydrothermal fluids circulating within the footwall exhumed at the Atlantis Massif and possibly in any detachment system.

Data Availability Statement

Results of individual analyses of the samples are available online on PANGAEA (https://doi.org/10.1594/PANGAEA.935262).

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