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1	Tracing of Cl input into the sub-arc mantle through the combined analysis of B, O and
2	Cl isotopes in melt inclusions.
3	
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11	Abstract
12	The effect that recycling crust and sediments have on the composition of the mantle
13	wedge, in particular in terms of volatiles, is still debated. Chlorine, an important fluid mobile
14	element that has stable isotopes with different concentrations in the terrestrial reservoirs, has
15	the potential to be used to trace slab-derived fluids.
16	Olivine-hosted melt inclusions (OHMIs) provide a first order constraint on the δ^{37} Cl of
17	primary magmas, since they are unaffected by near surface processes. In this study, $\delta^{37}Cl$
18	were coupled with $\delta^{11}B$ and $\delta^{18}O$ analyses in samples from the Lesser Antilles, Vanuatu,
19	Aeolian, NE Japan and Izu-Bonin arcs. This unique dataset is used to better understand the
20	large δ^{37} Cl variation in melt inclusions from a single sample. OHMIs from the Vulcano

22 2.6±0.8‰, respectively). These are different from δ^{37} Cl in OHMIs from the other three

(Aeolian arc) and Sukumoyama (Izu-Bonin arc) samples have similar δ^{37} Cl (-2.5±0.5‰ and -

23 localities (δ^{37} Cl of -0.7±0.6‰ for Aoba (Vanuatu arc) and St. Vincent (Lesser Antilles arc), -

1±0.9‰ for Iwate (NE Japan)). Vulcano OHMIs also have statistically different B and O 24 isotope compositions compared to those from the other locations: average $\delta^{11}B$ of -5.1±2.9‰ 25 for Vulcano OHMIs, compared to 2.5±3.7‰, 5.2±1.4‰, 7.0±2.2‰, 3.8±7.5‰ for 26 Sukumoyama, Iwate, Aoba and St. Vincent OHMIs, respectively. All OHMIs have δ^{18} O 27 between 4.0 and 7.4‰, except for those from Vulcano, which are significantly different, with 28 δ^{18} O from 7.2 to 9.1‰. Combining these three stable isotope systems suggests that the large 29 variation (>2‰) of δ^{37} Cl in OHMIs from a sample reflects inputs from different sources of Cl 30 31 rather than heterogeneities in a single main source. Variability between arcs might reflect different major sources of Cl. 32

Comparing OHMIs Cl isotope data from the Aeolian and Izu-Bonin arcs with existing bulk rock Cl isotope data suggest that OHMIs preserve the source signature of Cl input whereas this signal can be lost in whole rocks as a result of Cl isotope diffusive fractionation during Cl degassing. SIMS measurements of Cl isotopes in OHMIs could thus help refine models of Cl cycles in the mantle.

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Keywords: melt inclusions; chlorine isotopes; oxygen isotopes; boron isotopes; subduction
zones; SIMS

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42 **1. Introduction**

Chlorine has been widely studied as an element in volcanic systems (e.g., Aiuppa et al., 2009; Mather et al., 2012), but to date only a few studies have reported Cl isotopes (noted δ^{37} Cl) in rocks and minerals and used them to make petrogenetic interpretations (~35 papers since 1995 and a few pioneering works prior to that). After some debate about the Cl isotope values of the terrestrial reservoirs, refinement of bulk rock analytical techniques has brought

scientists to a consensus on the relatively narrow range of δ^{37} Cl found in terrestrial rocks 48 (Figure 1 and Supplementary material 1). In subduction zones, Cl, a soluble element, could 49 serve as a tracer of different Cl sources that contaminate the mantle wedge. Based on 50 theoretical calculations (Schauble et al., 2003) and experimental work (Liebscher et al., 2006), 51 fractionation of Cl isotopes between silicates and NaCl at T >300°C should be limited. This 52 implies that δ^{37} Cl should not be (or only slightly) fractionated during slab dehydration 53 (Barnes et al., 2006; Bonifacie et al., 2008; John et al., 2011). The isotopic composition of a 54 metasomatized mantle should thus reflect simple binary mixing between the host rock and the 55 metasomatizing agent. The δ^{37} Cl of arcs lavas vary from -2.6 to +3.0‰ (e.g., Barnes et al., 56 2009; Barnes and Straub, 2010). These studies have been used to estimate the Cl reservoirs 57 and suggested that Cl could be derived from altered oceanic crust (AOC), serpentinites or 58 sediments, depending on the depth of fluid extraction, which leads to a change in δ^{37} Cl across 59 an arc (Barnes et al., 2008). Variation along the Central America volcanic arc has also been 60 observed and has been used to suggest different sources of Cl along the arc (Barnes et al., 61 2009). 62

Recently, Manzini et al. (2017) reported δ^{37} Cl in olivine-hosted melt inclusions (OHMIs) 63 from three different arc settings (the Lesser Antilles, Aeolian and Vanuatu arcs). Their study 64 reported large intra-sample δ^{37} Cl variations (up to 2.5%), Figure 1) that are half of the 65 variation exhibited by all the bulk rocks from arc settings measured to date. Two arcs display 66 similar δ^{37} Cl (-1.9 to +0.6‰), whereas the third one has significantly different δ^{37} Cl 67 compositions (-3.4 to -2.0%). These authors suggested that for the Lesser Antilles and 68 Vanuatu arcs, δ^{37} Cl in the OHMIs reflect the imprint of serpentinite (and/or AOC) whereas 69 for the Aeolian arc, δ^{37} Cl is influenced by sediments. The difference between the two 70 different sets of compositions might reflect the different slab geometries that the arcs have. 71

72	As pointed out by Manzini et al. (2017), further work was required to better understand
73	the different δ^{37} Cl compositions between the three different arcs, as well as the relatively
74	large variation observed within a sample. Also, new data published for bulk rocks in the
75	Aeolian arc show large δ^{37} Cl difference between Stromboli lavas (-0.96 to +0.69‰; Liotta et
76	al., 2017) and Stromboli OHMIs (-3.2 to -1.75‰; Manzini et al., 2017). In order to
77	understand the large variations and assess how representative the δ^{37} Cl values recorded in
78	OHMIs are, this study couples δ^{37} Cl data obtained by Manzini et al. (2017) with δ^{11} B and
79	δ^{18} O obtained in the same OHMIs. In addition, new Cl, B and O data from two additional arcs
80	(Izu-Bonin and NE Japan) have been acquired. Oxygen isotopes do not largely fractionate
81	during dehydration at high temperature (e.g., Liebscher et al., 2006; Schauble et al., 2003;
82	Zheng, 1993). Comparison of these two isotopic systems allows sediment vs.
83	AOC/serpentinite influences to be identified. In contrast to O and Cl isotopes, B isotopes
84	largely fractionate during dehydration (e.g., Ishikawa & Nakamura, 1994; Peacok & Hervig,
85	1999), complicating the data interpretation due to the lack of B isotope partitioning data in the
86	different phases involved. However, B isotopes have the potential to distinguish between
87	AOC and serpentinites end-members as their respective bulk rocks are different (Boschi et al.,
88	2013; Chaussidon & Jambon, 1994; Smith et al., 1995; Vils et al., 2009). This unique dataset
89	combining Cl isotopes with two other stable isotope systems in melt inclusions is used to
90	assess if variations of δ^{37} Cl in melt inclusions within a sample represent heterogeneities in a
91	single, major, Cl reservoir, or alternatively reflect that Cl is added to the mantle wedge from
92	different sources. Disparities between the whole rocks and OHMIs are also discussed in terms
93	of analytical artefacts and/or fractionation due to diffusion.

2. Geological context and sample descriptions

All samples were previously studied and their geological settings are well-known. They were chosen as they contain glassy OHMIs with basaltic compositions, potentially providing access to Cl-undegassed melts. The different arcs have been chosen to represent various amounts of fluids involved in their magma genesis (see section 2.1). They are thus suitable samples to test whether δ^{37} Cl, coupled with other isotopic systems, could be a useful tracer of Cl sources and the fate of Cl in subduction zones.

102 2.1. Melt inclusions from previous studies

OHMIs from St. Vincent (Lesser Antilles arc), Vulcano (Aeolian arc) and Aoba (Vanuatu
arc) are the same as those reported by Manzini et al. (2017). OHMIs from Iwate Volcano (NE
Japan arc) were analyzed for volatiles by Rose-Koga et al. (2014). Nomenclature of each
OHMI is identical to their original papers.

The intra-oceanic volcanic arc of the Lesser Antilles arc is generated by the subduction of 107 the Atlantic Plate beneath the Caribbean Plate (see Macdonald et al., 2000 for a review). 108 Analyzed OHMIs were separated from a lapilli deposit collected near Troumaka Bay, in St. 109 Vincent, where high-MgO basalts occur. Based on B isotopes, Bouvier et al. (2008) suggested 110 that fluids involved in the magma genesis are derived from serpentinites, sediments and AOC. 111 The Aeolian arc is related to the subduction of the Ionian-Adriatic lithosphere underneath the 112 Calabrian arc. Mantle beneath the Aeolian arc volcanoes has been metasomatised by fluids 113 114 from the Ionian oceanic crust and the sediments from the Ionian basin in the eastern part of the arc (Peccerillo et al., 2013). The selected OHMIs come from La Sommata scoria cone on 115 Vulcano Island. OHMIs from Stromboli, analyzed for δ^{37} Cl by Manzini et al. (2017), were 116 not included in this study as they were too small to undergo further analysis. 117

118 The Vanuatu arc is generated by the northeastward subduction of the Australian Plate
119 beneath the Pacific Plate. The selected olivine crystals come from three different lapilli layers

of Aoba Island (Ao3 from an eruptive fissure on the volcano north flank; Ao15 from Torgil
tuff ring; and Ao17 from Red Cliff pyroclastic sequence; Sorbadere et al., 2011). Slab fluids
beneath Aoba involved fluids derived from serpentinites (Métrich and Deloule, 2014).

123 NE Japan arc is generated by the subduction of the Pacific Plate beneath the Okhotsk 124 Plate. The sample, coming from arc front Iwate Volcano, was studied by Rose-Koga et al. 125 (2014). These authors showed that fluids from the slab beneath Iwate Volcano are derived 126 from AOC and sediments. The selected OHMIs have already been analyzed for major and 127 volatiles elements, and Pb isotopes, and have been chosen based on their size – they have to 128 be large enough to place a δ^{37} Cl SIMS spots in between previous SIMS spots.

129

130 2.2. New melt inclusions from Izu-Bonin arc

New data for OHMIs from a sample collected from the Sukumoyama scoria cone in the 131 132 Higashi-Izu Monogenetic Volcano Field (HIMVF), on the northeastern Izu Peninsula, were used in this study. The Izu-Bonin arc is an intraoceanic arc generated by the subduction of the 133 Pacific Plate beneath the Philippine Sea Plate. Since 15 Ma, the Izu-Bonin arc has been 134 colliding with the Honshu Arc in central Japan to form the Izu collision zone. OHMIs from 135 sample IZ27-15 on Sukumoyama were analyzed previously for major, volatile and trace 136 elements (Nichols et al., 2012). Compositional variations were suggested to reflect interstitial 137 melts within the crystal mush of the magma chamber experiencing distinct crystallization 138 histories and variability between contributions from fluid and sediment melt components in 139 140 the source. For this study, new OHMIs were polished and analyzed from the same sample used by Nichols et al. (2012). 141

142 **3. Methods**

Olivine crystals from the 0.5-1 mm grain size fraction of crushed scoria were hand-picked under a binocular microscope and embedded in epoxy. Olivine crystals containing naturally glassy OHMIs were polished individually in order to expose the OHMI on the surface. Most OHMIs contain a shrinkage bubble, but none contain daughter minerals. Each polished olivine was removed from epoxy and pressed into indium, along with various reference materials for secondary ion mass spectrometry (see below). Each analytical method applies for all OHMIs.

150 *3.1.Electron microprobe (EMPA)*

Major element compositions of the MIs and their host olivine were acquired by electron 151 microprobe (EMPA) using a JEOL 8200 Superprobe. For the MIs, analytical conditions were 152 153 15kV, 10nA current and a 5 µm beam while conditions were 15kV, 15 nA, 1 µm, for the olivine host. Peak counting times were 30 s on all elements, except for K, Na and Ca (20 s). 154 Backgrounds were counted for 15 or 10 s (Si, Fe, Na, Ca, K). Cl was determined by EMPA 155 only in OHMIs analyzed by Manzini et al. (2017); for other OHMIs, Cl content was measured 156 using SIMS. For the MIs, the KL2-G glass standard (Jochum et al., 2006) was used to 157 calibrate SiO₂ and Al₂O₃. Other elements were calibrated on different minerals. ML3B-G 158 (Jochum et al., 2006) and San Carlos olivine were used as internal standards to check the 159 calibration. 160

161 *3.2.Secondary Ion Mass Spectrometry (SIMS)*

Boron, Cl and O isotope ratios were measured by secondary ion mass spectrometry (SIMS) using the CAMECA IMS 1280-HR at the SwissSIMS laboratory (University of Lausanne, Switzerland). A 1.5-2 nA, 10 kV Cs⁺ primary beam was used to analyze Cl and O isotope ratios, over two different sessions. The electron flood gun, with normal incidence, was used to compensate charges. For δ^{37} Cl, the analytical method was the same as described in

Manzini et al. (2017). To summarize, a pre-sputtering of 240 s with a 25 µm raster size was 167 used in order to remove all the possible Cl contamination. Analysis time was 240 s with a 10 168 µm raster. ³⁵Cl⁻ and ³⁷Cl⁻ were measured simultaneously on two Faraday cups (FCs), using 169 $10^{11} \Omega$ resistors, with a mass resolving power (MRP) set to ~3000. The internal uncertainty 170 (single measurement) for the samples was 0.15-0.5% (two standard errors, 2 s.e.), depending 171 on the Cl content. Reproducibility (point to point) was checked using an in-house Cl-enriched 172 173 standard UNIL_Gl-B7. Reproducibility of UNIL_Gl-B7 over eight points at the beginning of each session was <0.2‰ (two standard deviations, noted 2 SD hereafter), but up to 0.35‰ 174 2SD over a session (24-36h; two measurements of UNIL_Gl-B7 every six unknowns). 175 Instrumental mass fractionation (IMF), depending on Si, K, and Al mole content of the 176 glasses was calibrated using six glass standards (Manzini et al., 2017), including UNIL-GI-B7 177 at the beginning of each session (SM 2A). To that purpose, UNIL_Gl-B4, UNIL_Gl-B6, 178 RMR, PR2 and IB94 were measured four times at the beginning of each session, with two 179 measurements of UNIL GI-B7 in between each different glasses. To ensure the validity of the 180 calibration over the session and determine the uncertainty of the calibration, UNIL_Gl-B6 and 181 UNIL_Gl-B4 were measured repeatedly over the session. A reproducibility of 0.3‰ and a 182 maximum uncertainty of 0.4‰ were calculated. The IMF applied to each OHMIs was 183 calculated based on their major element composition, measured by EMPA before the SIMS 184 sessions. δ^{37} Cl is referenced to the standard mean ocean chloride (SMOC) 37 Cl/ 35 Cl ratio of 185 0.319592. 186

The Cl content of each OHMI could be estimated based on the ³⁵Cl count per second normalized to the primary beam intensity and compared to the standards. As the estimation by SIMS gives satisfactory results for Cl content <3000 ppm (SM 2B), SIMS was used to estimate the Cl contents of the new Sukumoyama OHMIs, which should have relatively low Cl contents based on those previously analyzed (Nichols et al., 2012).

¹⁶O and ¹⁸O secondary ions were analyzed at 2400 MRP and collected on two FCs in 192 multi-collection mode using 10^{10} and $10^{11} \Omega$ resistors, respectively. The FCs were calibrated 193 at the beginning of the session, using the calibration routine. Mass calibration was performed 194 at the beginning of each session (<12h). Each analysis takes ~4 minutes, including pre-195 sputtering (30 s) and automated centering of secondary electrons. Ion mass fractionation was 196 calibrated using several standards with different major element compositions (SM 2C). 197 Reproducibility of the standards is better than 0.3‰. δ^{18} O is referenced to the standard mean 198 ocean water (SMOW) 18 O/ 16 O ratio of 2005.2×10 $^{-6}$. 199

To determine δ^{11} B, 10 B and 11 B were measured in multicollection mode, using two electron 200 multipliers. The O⁻ primary beam intensity was 8 nA and mass resolution was set at 2400. A 201 single analysis (60 cycles) took 16 minutes. The instrumental isotopic fractionation was 202 determined by analyzing five standards with different major element compositions (SM 2D). 203 As already reported in the literature, instrumental isotopic fractionation was observed to be 204 independent of major element composition (Rosner et al., 2008). Reproducibility on the 205 standards varies from 2 to 4‰ (2SD), depending on the B content of individual melt 206 inclusions. δ^{11} B is referenced to the NBS 951 11 B/ 10 B ratio of 4.044. 207

Amongst the standards used for δ^{11} B and δ^{18} O calibrations, BHVO-2G and BCR-2G, with compositions encompassing those of our OHMIs, were also included in the sample mounts, as well as in the standard mount. During the session measuring δ^{18} O and δ^{11} B, several analyses of BHVO-2G and BCR-2G were undertaken before and after a maximum of 18 unknowns to monitor instrument stability (i.e., magnetic field stability). These standards were also treated as unknowns to check the accuracy of the calibration. Accuracies were similar to reproducibility: 0.3‰ for δ^{18} O and 3‰ for δ^{11} B.

215

4. Results

217 *4.1.Elemental geochemistry of melt inclusions*

All major elements, as well as details of the post-entrapment crystallization (PEC) corrections are reported in SMs 3-4. Compared to whole rocks from each locality, OHMIs plot at the primitive end of the differentiation trends (SM 5), and have basanite to basaltic compositions (Figure 2). The OHMIs from the different settings define distinct trends that show variable alkali enrichment, with some of St. Vincent OHMIs being the most alkali-rich, and those from Iwate Volcano, NE Japan arc, the most alkali-poor.

Manzini et al. (2017) showed that the highest and most variable Cl contents are in OHMIs from Aoba (0.06 to 0.39 wt%) and Vulcano (0.14 to 0.44wt%) islands, whereas in St. Vincent samples the range is more restricted (0.08 to 0.15 wt%). Rose-Koga et al. (2014) reported 0.03 to 0.05 wt% for Iwate OHMIs. The new data for OHMIs from Sukumoyama provide Cl contents from 0.04 to 0.09 wt%, similar to the values published by Nichols et al. (2012).

4.2.Stable isotope compositions

230 4.2.1. <u>Chlorine isotopes</u>

231 All stable isotope compositions are reported in Table 1. The six new data from Iwate Volcano range from $-2.1\pm0.8\%$ (2 s.e.) to $0.5\pm0.8\%$ (2 s.e.) (average of -1%), similar to St. 232 Vincent and Aoba OHMIs. The nine data from Sukumoyama OHMIs show distinct values 233 compare to Iwate Volcano OHMIs, ranging from -4.4±0.5‰ (2 s.e.) to -1.6±0.4‰ (2 s.e.) 234 (average of -2.3‰), more similar to Vulcano. As for Vulcano OHMIs, Cl isotopes for 235 Sukumoyama OHMIs can be compared to $\delta^{37}\text{Cl}$ bulk rock values. Indeed, lavas from two 236 volcanoes, Oshima and Niijima, located relatively close to Sukumoyama, at the northern end 237 of Izu-Bonin arc have been measured for δ^{37} Cl (Barnes et al., 2008). Oshima and Niijima 238

lavas have significantly higher Cl isotope compositions (-0.28‰ and -0.77‰, respectively)
compared to Sukumoyama OHMIs.

The lowest δ^{37} Cl value measured for a Sukumoyama OHMI, -4.4±0.5‰ (2 s.e.), is the 241 lowest measured for any OHMI to date and also lower than bulk rock values for solid 242 geological samples (Figure 1). This value has been carefully checked to assess its validity: all 243 244 SIMS analytical conditions are similar to other MIs measurements (secondary deflectors, primary intensity, topography, etc). Despite the fact that this inclusion also has the lowest Cl 245 content of Sukumoyama MIs (400 ppm), analytical artefact due to low Cl content is 246 discarded: internal error (counting statistic of a single measurement; 0.5‰, 2 s.e.) is similar to 247 the internal error of other OHMIs from this sample. Additionally, some Iwate OHMIs have as 248 low as, or even lower, Cl contents (300-400 ppm) and yet δ^{37} Cl for these inclusions is in the 249 250 range of published values (-1.6 to +0.5%). Hence the value is considered to be reliable.

4.2.2. <u>Boron isotopes</u>

The lowest δ^{11} B values have been measured in the St. Vincent OHMIs, with values 252 ranging from -18.5 ± 1.9 to $+11.7\pm1.9\%$ (2 s.e.) (average: +3.6%). This range, similar to the 253 range previously reported for St. Vincent OHMIs (-25.6 to +11.8‰, average 0.9‰; Bouvier 254 et al., 2008) is by far the largest range of δ^{11} B values reported for a group of OHMIs in this 255 study. $\delta^{11}B$ range from -9.5±1.9 to -1.0±2.1‰ (2 s.e.) (average: -5‰) for the Vulcano 256 OHMIs. Aoba OHMIs have positive δ^{11} B values (+3.4 to +12.0, average of +7‰), higher than 257 the values reported for MIs from two other Aoba samples (-11.9 to +6.4%; Métrich and 258 Deloule, 2014). Sukumoyama OHMIs have variable δ^{11} B values, ranging from -4.8±4‰ to 259 260 +5.8±4‰ (2 s.e.) (average: 1.7‰). Iwate Volcano OHMIs have a restricted range of positive δ^{11} B values (+4.7±1.7‰ to +8.1±2‰; average: 6.3‰). 261

262 4.2.3. Oxygen isotopes

The δ^{18} O varies from 4.3±0.4 to 6.9±0.4‰ (2 s.e.), with an average of 5.8‰ for the St. 263 Vincent OHMIs, similar to the average (6.0%) δ^{18} O reported by Bouvier et al. (2008). Aoba 264 OHMIs have a similar range of δ^{18} O composition, from 4.4±0.2 to 7.0±0.2‰ (2 s.e.) 265 (average: 6.3‰); and Sukumoyama OHMIs also have a similar range of δ^{18} O (4.8±0.3 to 266 267 7.4±0.3‰, 2 s.e.), but with slightly higher average (6.6‰). In contrast, Iwate Volcano OHMIs have slightly lower δ^{18} O values of 3.9±0.3 to 6.6±0.3‰ (2 s.e.), with an average of 5.4‰. 268 Vulcano OHMIs have the highest δ^{18} O measured in this study. They vary from 6.1±0.2 to 269 $9.1\pm0.2\%$ (2 s.e.), with an average value of 7.9‰. 270

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5. Discussion

273 5.1. Significance of δ^{37} Cl variation in OHMIs within a sample

Manzini et al. (2017) reported the first data of δ^{37} Cl in OHMIs from arc settings and 274 showed that Cl isotopes can vary by up to 2.5% within a single sample (St. Vincent). Here, 275 276 the OHMIs from two new samples from Iwate Volcano and Sukumoyama have variations in δ^{37} Cl of 2.6 and 2.8‰, respectively, further confirming that relatively large variations (>2‰) 277 in Cl isotopes within OHMIs from a single sample might be common. Given the large range 278 of δ^{37} Cl in terrestrial Cl reservoirs (Figure 1), the measured variation in OHMIs within a 279 sample could reflect either heterogeneity of a single Cl reservoir, or contributions from 280 different Cl reservoirs. To better understand the variability of δ^{37} Cl within a sample, O and B 281 isotopes were measured in the same OHMIs. As mentioned in the introduction, comparison 282 with O isotopes could help to distinguish between the influence of fluids from sediments and 283 the influence of fluids from AOC/serpentinites, whereas B isotopes may differentiate between 284 fluids from AOC and serpentinites. 285

286 5.1.1. Are the measured δ^{37} Cl representative of the source?

Before any interpretation of source process(es), it is important to evaluate several potential processes that can disturb the measured δ^{37} Cl. First, we ensured that the selected OHMIs have not degassed Cl. In a plot of SiO₂ vs. Cl content of the OHMIs (Figure 3), within each sample Cl remains constant over the SiO₂ variation. This suggests that Cl is not degassing from the melts as they were trapped.

292 Analytical artefacts were also investigated. Uncertainty in the calibration for Cl isotopes is 0.4% (2SD), smaller than the observed variation within a sample (> 2‰) or between samples. 293 The absence of correlation between δ^{37} Cl and SiO₂ (Figure 4A) and Na₂O + K₂O (Figure 4B) 294 illustrate that the different δ^{37} Cl measured in Vulcano and Sukumoyama OHMIs compared to 295 other OHMIs (Fig. 1) is not an analytical artefact. Indeed, OHMIs from these two samples 296 have different Cl contents (>3000 ppm and <1000 ppm, respectively) and major element 297 compositions (Figure 2, SMs 3 to 5). Since calibrations were made with the same standards 298 299 for all samples, and that OHMIs are corrected for IMF using their own major element compositions, it cannot be argued that the low δ^{37} Cl values in the OHMIs are induced by 300 analytical bias. The absence of correlation between δ^{37} Cl and SiO₂ and Na₂O + K₂O within a 301 sample (Figure 4) also suggests no fractionation of δ^{37} Cl during magmatic evolution or due to 302 303 variable degrees of melting.

We have also evaluated the effects of secondary processes (e.g, PEC, shrinkage bubble). Given that Cl is an ultra-trace element (<1ppm) in magmatic olivine (e.g., Urann et al., 2017), late crystallization of olivine on the wall of OHMIs should not strongly affect the Cl composition and Cl isotopes of OHMIs. This is verified by the absence of any relationship between δ^{37} Cl and the extent of PEC. Shrinkage bubbles can contain volatile elements, mostly CO₂ (e.g., Moore et al., 2015). So far, Cl content in such bubbles have not been determined. Most of the OHMIs analyzed here have a shrinkage bubble inside. Several olivines contain

two or three OHMIS, but only one olivine contains two OHMIs with one containing a bubble 311 and the other not (Aoba 3-1a and b). The one without bubbles contain 300 ppm less Cl than 312 the one with a bubble, but the two OHMIs have similar δ^{37} Cl, within error (-0.5±0.2 and -313 0.7±0.2%; SMs 3-4). Also, within a sample, the range of Cl content and δ^{37} Cl are similar for 314 OHMIs with or without a shrinkage bubble. We thus conclude that the presence of a 315 shrinkage bubble does not affect Cl isotopes of OHMIs. This is further confirmed by the 316 absence of a correlation between δ^{37} Cl and Cl/K₂O (Figure 5), suggesting no Cl degassing 317 syn- or post-entrapment. Also, Cl is not affected by diffusive re-equilibration through the host 318 olivine (Bucholz et al., 2013). All this information suggests that the measured δ^{37} Cl are 319 representative of the melt in which the olivine grew. 320

321

322

5.1.2. <u>Cl and O isotope systematics</u>

In a plot of Cl isotopes compared to O isotopes, AOC and serpentinite fields are distinct 323 from the field for sediments (Figure 6 and references in the caption). Oxygen isotope 324 fractionation between hydrous minerals and fluids at high temperature is relatively small (for 325 example, at 500°C, ~-1.5‰ for serpentinite, amphibole; Zheng, 1993). Also, Cl isotopes in 326 metamorphosed serpentinites and AOC are similar to those in fresh corresponding lithologies, 327 suggesting that δ^{37} Cl do not fractionate in fluids during the dehydration of mafic lithologies 328 (Barnes et al., 2006; Bonifacie et al., 2008; John et al., 2011). Thus, the Cl-O isotope 329 systematics in the OHMIs from each location have been compared to these end-members 330 (Figure 6). In all the data taken together, Cl isotopes are negatively correlated with O 331 isotopes, with the lowest δ^{37} Cl observed in OHMIs with highest δ^{18} O and vice versa. In detail, 332 the OHMIs from St. Vincent and Iwate Volcano display a negative relationship, whereas no 333 systematic could be described for those from Vulcano, Aoba and Sukumoyama. There is 334

greater variation in δ^{18} O data at low δ^{37} Cl (< -1.6‰). The lowest bulk δ^{37} Cl (down to -3‰) and the highest bulk δ^{18} O (> 9‰) have been observed in fresh sedimentary rocks, suggesting that the low δ^{37} Cl (<-2‰) associated with high δ^{18} O (>7‰) in the OHMIs do probably reflect the influence of fluids derived from sediments. The increased scatter of the OHMIs toward this end-member could be explained by the large range of subducted sediment compositions (carbonates, silicic sediments, shales, e.g., Bindeman, 2008 for a review).

High δ^{37} Cl (>0.5‰) associated with low δ^{18} O (<5.5‰) point toward an AOC or 341 serpentinite end-member. In detail, the coupled δ^{37} Cl and δ^{18} O values reported for fresh 342 serpentinite rocks formed at more than 100°C suggest that δ^{18} O increases with δ^{37} Cl in 343 serpentinites (Bonifacie et al., 2008; Boschi et al., 2013). A fluid derived from fresh oceanic 344 serpentinites would have a δ^{18} O value 1.8‰ higher than the bulk rock it is derived from 345 (Zheng, 1993) considering serpentinite dehydration at ~680°C (antigorite breakdown, e.g., 346 Trommsdorff et al., 1998). Thus, a fluid escaping a serpentinite with δ^{37} Cl of ~1‰ and δ^{18} O 347 from 4.5 to 7‰, based on the literature, will have a δ^{18} O composition >6.3‰. Such δ^{18} O is 348 too high to explain the highest δ^{37} Cl end-member influencing the compositions of three out of 349 four of the OHMIs with δ^{37} Cl > 0‰. Subducted serpentinites with higher δ^{37} Cl have been 350 reported (up to 2.4‰, Barnes et al., 2006, 2013, 2014). In subducted serpentinite samples for 351 which δ^{37} Cl and δ^{18} O have been determined, the same tendency of δ^{37} Cl to increase with δ^{18} O 352 is observed (Barnes et al., 2014), leading to a similar conclusion as for the fresh oceanic 353 serpentinites: a fluid escaping high δ^{37} Cl subducted serpentinites might have a δ^{18} O too high 354 to explain the composition of OHMIs with $\delta^{37}Cl > 0$ %. AOC, especially high temperature 355 altered portions, have high δ^{37} Cl (up to +1.6%; Barnes and Cisneros, 2012) and could have 356 low δ^{18} O (down to 3‰ for the bulk rock (Alt & Bach, 2006); even as low as 0‰ for 357 amphibole veins (Muehlenbachs, 1986)). If we assume a temperature of dehydration for 358

amphibole of ~500°C (blueschist – eclogite transition), a dehydration fluid from amphibole veins would have δ^{18} O ~1.7‰ higher than the dehydrating lithology/mineral (Zheng, 1993). Thus, local dehydration fluids from lower AOC could have δ^{18} O values as low as 1.7‰, making amphibole-bearing AOC a more suitable candidate for the high δ^{37} Cl - low δ^{18} O endmember of the correlation. However, in detail, the Aoba OHMIs do not follow the general trend and might also point toward a serpentinite fluid end-member.

Following this reasoning, Vulcano and Sukumoyama OHMIs, with lower δ^{37} Cl and higher δ^{18} O than MORB (Sharp et al., 2007), possibly record the strongest influence from sediment derived fluids. St. Vincent and Iwate Volcano OHMIs, which have δ^{37} Cl and δ^{18} O both higher and lower compared to DMM, suggest that melts were influenced by a fluid possibly derived from the lower AOC and also some sediments. Aoba OHMIs also have both higher and lower δ^{37} Cl and δ^{18} O compared to DMM, suggesting influence from sediment fluids and AOC or serpentinite fluids.

372 It is important to note that contrary to the possible AOC and/or serpentinite end-member, which does not seem to have suffered fractionation of Cl isotopes during dehydration, as 373 previously suggested (Barnes et al., 2006; Bonifacie et al., 2008; John et al., 2011; Schauble 374 375 et al., 2003), the sediment end-member probably has fractionated at some point, as some OHMIs have δ^{37} Cl values lower than the lowest δ^{37} Cl reported for bulk sediments. This 376 observation agrees with John et al. (2010), who suggest preferential loss of ³⁵Cl from 377 sediments into pore fluids, even at high temperature, leads to enrichments in δ^{37} Cl in recycled 378 subducted marine sediments. 379

380 5.1.3. <u>Cl and B isotope systematics</u>

381 The distinction between AOC and serpentinite derived fluids is not unique due to their 382 similar δ^{18} O and δ^{37} Cl. Here, we investigate whether they can be distinguished in terms of their Cl-B isotope systematics (Figure 7). To date, Cl isotopes have not been coupled with B isotopes in the same rock samples. Fields for the different lithologies were thus assembled from different samples and different settings. Cl isotope values are given in SM 1. For B isotopes, sediment values are from Smith et al. (1997) and Ishikawa and Nakamura (1993); The AOC values are from Smith et al. (1995), and serpentinite values are from Boschi et al. (2013) and Vils et al. (2009). For reference, the MORB field is also plotted using the data for δ^{11} B from Chaussidon and Jambon (1994) and for δ^{37} Cl from Sharp et al. (2007).

Boron isotopes fractionate strongly during dehydration, with ¹¹B partitioning preferentially into the fluid phase (e.g., Peacok & Hervig, 1999; Ishikawa & Nakamura, 1994). Thus, B isotopes of the rock reservoirs cannot be directly compared to MI data. A simple open system Rayleigh distillation is used to estimate the theoretical composition of the dehydration fluids (Bouvier et al., 2010; Rose et al., 2001). For these calculations, B isotope fluid-rock partition coefficients of 3-5 ‰ for serpentinites (Tenthorey and Hermann, 2004); and 5‰ for AOC and sediments (You et al., 1996) were used.

397 There is no simple relationship between Cl and B isotopes for the OHMIs (Figure 7). Instead, for the whole dataset, different end-members are needed to explain the variations of 398 δ^{11} B and δ^{37} Cl in OHMIs. A first fluid end-member with δ^{11} B >12‰ and δ^{37} Cl between -2 399 and -0.5% could be described, most probably representing the influence of fluids derived 400 401 from serpentinite dehydration. Indeed, amongst the three major Cl reservoirs shown in Figure 7, serpentinite rocks form the reservoir with the highest δ^{11} B. Based on the Rayleigh 402 403 distillation calculation, a fluid derived from the dehydration of serpentinites with an initial average δ^{11} B composition of 20% (Figure 5) would have δ^{11} B down to ~15% after 75% of B 404 loss from serpentinite. 405

406 A second end-member can be described by higher δ^{37} Cl (>1‰) and lower δ^{11} B (<~4‰). 407 High δ^{37} Cl have been reported for AOC with a lithology containing amphiboles (Barnes and Cisneros, 2012) or for metamorphosed serpentinites (Barnes et al., 2014). Based on Figure 6, OHMIs with δ^{37} Cl > 0‰ most likely record the influence of fluids released from amphibole bearing- lower AOC. If an influence from an AOC dehydration fluid is assumed, with initial AOC average B isotopes composition of 3‰, δ^{11} B of ~4‰ would suggest that a fluid escaping from an AOC lithology would have lost ~50% of B.

Finally, the last end-members most probably represent fluids from the subducted 413 sediments, as suggested by Figure 6. Indeed, the lowest δ^{37} Cl are measured in sedimentary 414 rocks (Figure 1). Likewise, extremely low $\delta^{11}B$ (<-15‰) measured in OHMIs are generally 415 associated with the influence of fluids lost by dehydrated sediments - be they melts or fluids -416 derived from such a lithology (e.g., Bouvier et al., 2010; Rose et al., 2001). Interestingly, 417 Sukumoyama and Vulcano OHMIs seem to require two distinct sediment end-members, as in 418 Figure 6 (thus OHMIs with similar δ^{37} Cl, but different δ^{18} O and δ^{11} B). This distinction could 419 reflect different degrees of dehydration of the sediment lithology. However, due to the 420 relatively small fractionation factor for O isotopes in the fluid at high temperature (650°C, -421 0.4 and 1.8% for quartz and calcite; Zheng et al., 1993, 1999), different degrees of 422 dehydration of similar sediment lithologies would not explain the different δ^{18} O of Vulcano 423 and Sukumoyama OHMIs. When compared to sediment bulk rock data, continental sediments 424 have lower δ^{11} B than marine sediments (Ishikawa and Nakamura, 1993). Similarly, marine 425 sediments have higher δ^{18} O compared to continental sediments (see Bindeman et al., 2008 for 426 a review). Simple Rayleigh distillation of marine sediment with an average initial composition 427 428 of 5‰ suggest 85% B loss to reach a fluid end-member with an average composition of 0‰. Similarly, an average composition of -8% for continental sediments would lead to 429 dehydration fluids with a composition of -14‰ after 90% B loss. Whereas sediments being 430 subducted in the Aeolian arc are mostly continental (dust from Sahara and clastic sediments 431 from river Nile; Klaver et al., 2015), sediments subducted in the Izu-Bonin arc, where 432

Sukumoyama is located, are mainly composed of marine sediments (ODP leg 1149; Plank et 433 al., 2007) or clays (ODP leg 808; Plank and Langmuir, 1998), both of which have higher δ^{18} O 434 than continental sediments. Oxygen, B and Cl isotopes all point toward two distinct sediment 435 end-members that are coherent with the tectonic settings of Vulcano and Sukumoyama. Boron 436 isotopic variability within each sediment end-member could represent variable degrees of 437 438 dehydration or variable compositions of the subducted sediments. John et al. (2010) suggested that dehydrated subducted marine sediments could have δ^{37} Cl up to +4‰ (2‰ higher than 439 highest measured values for fresh sediments, SM 1) due to Cl isotope fractionation during 440 dehydration. Thus, a fluid derived from the dehydration of sediment with δ^{37} Cl as low as -5% 441 would suggest a similar, but opposite as in fluid and not in solid, fractionation than suggested 442 by John et al. (2010) (2.5‰ lower than lowest measured values for fresh sediments). 443

Combining B, O and Cl isotopes might thus allow better identification of different Cl reservoirs (serpentinites vs. AOC; marine vs continental sediments) involved in magma genesis beneath arcs. Whereas the serpentinite reservoir cannot be easily identified in the O and Cl isotopes system, it can be distinguished from AOC in the B and Cl isotopes system. Similarly, influence of marine vs. terrigenous sediments seems to be more easily tracked in B and Cl isotope systems.

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5.1.4. <u>Implications for the significance of δ^{37} Cl variability</u>

Based on Figures 6 and 7, St. Vincent and Aoba mostly vary between AOC and serpentinite dehydration fluid end-members, suggesting a more pronounced influence from AOC and serpentinite fluids, compared to sediment fluids. Sediment fluids for St. Vincent can be marine and/or continental. The influence of three fluid end-members agrees with previous studies (Bouvier et al., 2008, 2010), and the suggested marine and continental sediments agrees with the sediment composition of the Barbados prism (Carpentier et al., 2008). Aoba OHMIs also seem to reflect the mixed influence of AOC and serpentinite fluids, and a marine fluid sediment component. Vulcano and Sukumoyama OHMIs record the strongest sediment influence, with a possible influence from serpentinite fluids to explain the scatter. Iwate Volcano OHMIs follow the Cl-O isotopes trend (Figure 4) suggesting the influence of both AOC and sediment fluids, in agreement with Pb isotope measurements on Iwate Volcano OHMIs (Rose-Koga et al., 2014). This is confirmed in Figure 7, with δ^{11} B suggesting an influence from marine sediments.

464 Overall, these observations, based on the Cl-O-B isotope systematics, are consistent with 465 the geological settings of the volcanoes that have been sampled and suggest that variability 466 within a sample reflects multiple reservoirs adding Cl to the mantle wedge. Inter-arc 467 variations in δ^{37} Cl most probably reflect global differences in the values for the main Cl 468 reservoir (e.g., sediments vs. AOC and/or serpentinites).

469 5.2. Bulk rock vs. OHMI Cl isotope differences

New low δ^{37} Cl (< -1.5‰) are reported here, with Sukumoyama OHMIs having similar 470 δ^{37} Cl compositions compared to Vulcano OHMIs. Interestingly, bulk rock values of samples 471 from the same arcs exist and can be compared with OHMI data. Even if bulk rock data are not 472 from the same sample, nor from the same volcanoes, as the OHMIs in this study, nearby 473 volcanoes from the same arc should have similar source(s) of Cl input. For the Aeolian arc, 474 bulk rock data exist for Etna and Stromboli, both having similar δ^{37} Cl (-0.96 to 0.69‰ for 475 Stromboli and -0.7 to 0.1 for Etna; Liotta et al., 2017; Rizzo et al., 2013) despite their 476 different magmatic histories (e.g., Tommasini et al., 2007). Also, as described in the 477 introduction, Manzini et al. (2017) reported δ^{37} Cl in OHMIs from Stromboli, with values 478 significantly different (-3.2 to -1.75‰) to Stromboli lavas, but similar to Vulcano OHMIs. We 479

thus conclude that bulk rock data from nearby volcanoes (e.g., Stromboli for Vulcano, andOshima and Niijima for Sukumoyama) could indeed be used to compare to OHMIs.

Bulk rock data are clearly different to our OHMI data, with a shift of up to 2‰ and 3.6‰ toward lower values for Vulcano and Sukumoyama OHMIs, respectively. As discussed in section 5.1.1., this could not be induced by analytical bias. Rather, the combination of negative δ^{37} Cl and δ^{11} B associated with elevated δ^{18} O suggest that the main source of Cl is from a fluid issued from sediments, as discussed above.

Fortin et al. (2017) showed that Cl isotopes could be fractionated by up to 5% by Cl 487 diffusion during bubble growth, and potentially more in the case of diffusion not related to 488 bubble growth (depending on time scale). As mentioned by these authors, fractionation by Cl 489 490 diffusion during the formation of melt inclusions (i.e., via boundary layer entrapment) could not generate lower isotopic compositions, but instead should generate higher δ^{37} Cl compared 491 to the initial composition. In section 5.1.1., the migration of Cl into a shrinkage bubble was 492 ruled out. If fractionation due to diffusion of Cl isotopes during shrinkage bubble growth 493 within the OHMIs does occur, it will cause ³⁵Cl to preferentially enter the bubble, leading to 494 δ^{37} Cl progressively increasing in the melt (Fortin et al., 2017). In this case, OHMIs should 495 have δ^{37} Cl compositions higher than bulk rock, whereas we have observed the opposite. As 496 no correlation has been observed between Cl or Cl/K₂O and δ^{37} Cl (Figure 5), the lower δ^{37} Cl 497 498 values measured in Vulcano and Sumukoyama OHMIs compared to respective bulk rocks probably do not represent degassing melts. On the contrary, bulk rocks have often suffered 499 some volatile degassing. For example, the Stromboli OHMIs measured for δ^{37} Cl have Cl 500 contents from 0.14 to 0.46 wt% (Manzini et al., 2017), whereas bulk rocks only contain 0.03 501 to 0.08 wt% (Liotta et al., 2017). Assuming fractionation due to diffusion of Cl isotopes 502 during bubble growth in the magma (as opposed to shrinkage bubble growth in OHMIs), 503 gases enriched in ³⁵Cl could be created. This will thus create low δ^{37} Cl gases and a magma 504

with δ^{37} Cl increasing with the increase of degassing. Interestingly, low δ^{37} Cl has been measured in Hakone (northern part of Izu-Bonin arc, down to -5.41‰; Barnes et al., 2008) and Stromboli gases (down to -2.2‰; Liotta et al., 2017). OHMIs from Sukumoyama and Vulcano could thus represent relatively primary undegassed melt, whereas the bulk rocks record Cl loss during degassing, enriching the melt in ³⁷Cl due to diffusive fractionation during bubble growth. In that case, the bulk rocks lose the original δ^{37} Cl signature of the source of the Cl input.

512 **6.** Conclusions

This study expands the dataset of Cl isotopes in OHMIs from arc settings and is the first coupling δ^{37} Cl with two other isotopic systems in OHMIs. This dataset suggests that Cl isotopes variation in OHMIs within a single sample most probably record Cl input from different Cl reservoirs, rather than heterogeneity from a single reservoir. Inter-arc variability in δ^{37} Cl might reflect differences in the main Cl reservoirs (e.g., larger influences of sediments in some arcs).

Indirect comparison of OHMI data from Vulcano and Sukumoyama and bulk rock data from the same arc reveal large (2 to 4‰) differences, with OHMIs having lower δ^{37} Cl than bulk rocks. This difference most probably reflects diffusive fractionation of Cl isotopes during degassing of the lavas. OHMIs have relatively primary compositions whereas the bulk lavas have lost the original δ^{37} Cl signature during degassing. This conclusion is supported by some low (<0‰) δ^{37} Cl gases measured at volcanoes in both arcs.

525 Chlorine isotope measurements by SIMS in MIs could thus provide valuable and unique 526 access to relatively primary (undegassed) δ^{37} Cl compositions. Better constraints on the source 527 of Cl is important to better understand the behavior or Cl in different contexts and ultimately, 528 refine the Cl budget. 529

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686 **Captions:**

Figure 1: Compilation of published δ^{37} Cl for terrestrial bulk rocks and olivine-hosted melt inclusions (OHMIs). Sukumoyama and Iwate Volcano OHMIs are new data, the remaining are published data. A summary of the database can be found in SM 1, with all the associated references. OHMIs are plotted using symbols, whereas bulk rocks are represented with bars. Sediments and serpentinite bulk rocks are plotted with dark colors representing fresh rocks, and light colors metamorphic rocks.

- Figure 3: SiO₂ compared to Cl content in the studied OHMIs. Symbols as in Figure 2. No variation of Cl content
 with increasing SiO₂ is observed, suggesting Cl is not degassing during OHMIs entrapment.
- 697 <u>Figure 4:</u> Chlorine isotopes compared to SiO₂ (A) and Na₂O + K₂O (B) in the OHMIs. No correlation between 698 δ^{37} Cl and the major elements are seen, proving that the matrix effect has been corrected. Also, the absence of any

699 correlation suggests that Cl isotopes are not affected by extent of magmatic evolution.

Figure 5: Chlorine isotopes compared to Cl/K₂O ratios. The absence of correlation between δ^{37} Cl and Cl/K₂O supports the absence of Cl degassing at the time of entrapment. Thus, the variation of δ^{37} Cl in the OHMIs cannot be linked to degassing.

703 Figure 6: O isotopes compared to Cl isotopes in St. Vincent (squares), Aeolian islands (triangles), Aoba 704 (diamonds), Sukumoyama (yellow circles) and Iwate (brown circles) OHMIs. All OHMIs define a negative 705 relationship between δ^{37} Cl and δ^{18} O. Sukumoyama and Aeolian OHMIs record the strongest influence from 706 sediment fluids, whereas St. Vincent and Iwate Volcano record fluids derived from the dehydration of both AOC 707 and sediments. Aoba also reflect influences from sediment fluids, and either AOC or serpentinite fluids. Fields 708 for sediments, AOC and serpentinites are based on bulk rock data. Bulk rock O and Cl isotope data were 709 obtained on the same samples only for oceanic serpentinites. Cl isotope data are from the same references as in 710 Figure 1 and SM 1. Oxygen isotope data for AOC are from Alt (2003) and Alt & Bach (2006). Only fresh oceanic serpentinites have been reported here (data for δ^{18} O are from Bonifacie et al. (2008) and Boschi et al. 711 (2013)). Fresh sedimentary rock δ^{18} O data are from the compilation of Bindeman (2008). 712

713	Figure 7: B isotopes compared to Cl isotopes. Symbols are as in Figure 5. For comparison, different reservoirs
714	have been plotted (see text for references). As for Figure 4, only fresh sediments and serpentinites have been
715	reported. Reservoirs defined by dashed lines are a compilation of data in which none of the samples have
716	combined δ^{11} B and δ^{37} Cl measurements. OHMIs suggest the influence of four different fluid end-members, one
717	each for AOC and serpentinites, and two distinct end-members for sediments, one for marine and one for
718	continental sediments.
719	
720	Table 1: B, O and Cl isotopes in OHMIs
721	
722	
772	Supplementary material
/23	Supplementary material
723	SM1: Database for Cl isotopes in terrestrial bulk rocks and OHMIs
723 724 725	SM1: Database for Cl isotopes in terrestrial bulk rocks and OHMIs SM2: SIMS calibrations
723 724 725 726	SM1: Database for Cl isotopes in terrestrial bulk rocks and OHMIs SM2: SIMS calibrations SM3: Summary of major element and stable isotopes of OHMIs used in Manzini et al. (2017)
723 724 725 726 727	SM1: Database for Cl isotopes in terrestrial bulk rocks and OHMIs SM2: SIMS calibrations SM3: Summary of major element and stable isotopes of OHMIs used in Manzini et al. (2017) SM4: Summary of major elements and stable isotopes of OHMIs from Iwate Volcano and Sukumoyama
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Figure 2 Click here to download high resolution image



Figure 3 Click here to download high resolution image







Figure 6 Click here to download high resolution image





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Table 1: B, O and Cl isotopes in OHMIs

	δ ³⁷ Cl	2σ	δ ¹⁸ 0	2σ	δ ¹¹ B	2σ				
OHMIs from St. Vincent (Lesser Antilles arc),										
svn4b-24	-0.5	0.4	4.3	0.4	-7.0	2.0				
svn4b-36	-0.4	0.3	6.0	0.2	3.9	2.2				
svn4b-60	-0.5	0.5	6.0	0.4	3.2	2.9				
svn4b-90a	-0.2	0.3	5.9	0.2	n.d.	n.d.				
svn4b-90b	-1.1	0.3	7.0	0.3	8.4	2.5				
svn4b-90v	-0.7	0.3	6.1	0.3	6.0	n.d.				
svn4b-164a	0.2	0.2	5.1	0.2	6.4	2.0				
svn4b-171a	-1.2	0.3	6.9	0.4	-18.5	1.9				
svn4b-176a	-0.4	0.3	4.7	0.3	7.1	2.0				
svn4b-181a	0.6	0.3	4.4	0.3	4.5	1.9				
svn4b-181b	-1.7	0.2	6.9	0.4	2.0	2.2				
svn4b-184	-0.5	0.3	6.3	0.4	-3.2	3.0				
svn4b-186	-0.8	0.4	6.3	0.3	n.d.	n.d.				
svn4b-MM1a	-1.3	0.3	6.3	0.1	9.1	1.6				
svn4b-MM2	-1.1	0.4	6.2	0.4	10.8	1.6				
svn4b-MM3a	-0.1	0.3	6.1	0.7	5.9	2.0				
svn4b-MM7	-1.9	0.5	5.3	0.2	10.3	2.4				
svn4b-MM15a	-1.2	0.3	6.1	0.4	3.7	2.4				
svn4b-MM22	-0.8	0.3	5.4	0.3	11.7	1.9				
OHMIs from Ac	oba (Va	nuat	u arc)							
17–10-ь	-0.4	0.2	n.d.	n.d.	7.9	3.2				
17–10-с	-0.3	0.1	6.6	0.2	7.9	3.2				
17–16-a	-1.2	0.1	6.6	0.2	4.8	3.2				
17–16-b	-0.7	0.2	6.0	0.2	5.8	3.2				
3–9	-1.4	0.7	7.0	0.2	n.d.	n.d.				
3–10 h3–1-a	-0.9	0.2	4.4	0.2	5.6	3.2				
3–1-a	-0.5	0.2	6.7	0.2	7.7	3.6				
3–1-b	-0.7	0.2	6.3	0.2	8.2	3.3				
3-h1-a	-0.5	0.1	6.4	0.2	8.1	3.7				
3-4	-0.82	0.2	6.2	0.2	12.0	3.2				
3-5	0.4	0.1	6.4	0.2	4.7	3.2				
3-0	-1.8	0.2	6.1	0.2	1.3	2.1				
15–1-a	-0.3	0.2	6.5	0.2	⊃.⊃ 2.4	3.2				
15–10-a	-1.2	0.3	0.1	0.2	3.4	3.2				
15–10-D	-0.4 Jaama (0.2	0.3	0.2	8.9	3.2				
CHIMIS IFOM VU	псапо (э э		an ar	() 0.2	4.0	าา				
51 S2	-2.3 2 0	0.2	0.3 8 7	0.2	-4.9 5 1	∠.∠ 1 0				
54	-2.0 2.5	0.1	0.2 8 0	0.2	-J.I 2 1	1.0				
55 \$7_9	-2.3 2 2	0.1	0.U Q 1	0.2	-3.1	1.0 1.0				
57-a \$7-b	-2.2 _2.0	0.1	0.1 Q 1	0.2	-9.5	1.7 7 1				
S0	-2.0 2.2	0.1	9.1 8 5	0.2	-1.0	$\frac{2.1}{1.7}$				
	-2.2	0.1	0.5	0.2	-3.7	1./				

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