Evolution of fluid-rock interaction in the Reykjanes geothermal system, Iceland:
Evidence from Iceland Deep Drilling Project core RN-17B

Andrew P.G. Fowler^{a*}, Robert A. Zierenberg^a, Peter Schiffman^a, Naomi Marks^b,
Guðmundur Ómar Friðleifsson^c

^a Department of Earth and Planetary Sciences, University of California, Davis,
California, 95616, USA

^b Lawrence Livermore National Laboratory, 7000 East Avenue, Livermore, California
94550, USA

^c HS Orka, hf, Reykjanesbaer, Iceland

*Corresponding author: Department of Geology, University of California, Davis,
California, 95616, USA. email: apfowler@ucdavis.edu

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1 **Abstract**

2 We describe the lithology and present spatially resolved geochemical analyses of samples
3 from the hydrothermally altered Iceland Deep Drilling Project (IDDP) drill core RN-17B.
4 The 9.3 m long RN-17B core was collected from the seawater-dominated Reykjanes
5 geothermal system, located on the Reykjanes Peninsula, Iceland. The nature of fluids and
6 the location of the Reykjanes geothermal system make it a useful analog for seafloor
7 hydrothermal processes, although there are important differences. The recovery of drill
8 core from the Reykjanes geothermal system, as opposed to drill cuttings, has provided the
9 opportunity to investigate evolving geothermal conditions by utilizing *in-situ*
10 geochemical techniques in the context of observed paragenetic and spatial relationships
11 of alteration minerals. The RN-17B core was returned from a vertical depth of ~ 2,560 m
12 and an *in-situ* temperature of ~ 345°C. The primary lithologies are basaltic in
13 composition and include hyaloclastite breccia, fine-grained volcanic sandstone, lithic
14 breccia, and crystalline basalt. Primary igneous phases have been entirely
15 pseudomorphed by calcic plagioclase + magnesium hornblende + chlorite + titanite +
16 albitized plagioclase + vein epidote and sulfides. Despite the extensive hydrothermal
17 metasomatism, original textures including hyaloclastite glass shards, lithic clasts, chilled
18 margins, and shell-fragment molds are superbly preserved. Multi-collector LA-ICP-MS
19 strontium isotope ratio ($^{87}\text{Sr}/^{86}\text{Sr}$) measurements of vein epidote from the core are
20 consistent with seawater as the dominant recharge fluid. Epidote-hosted fluid inclusion
21 homogenization temperature and freezing point depression measurements suggest the
22 RN-17B core records cooling through the two-phase boundary for seawater over time to
23 current *in-situ* measured temperatures. Electron microprobe analyses of hydrothermal

24 hornblende and hydrothermal plagioclase confirm that while alteration is of amphibolite-
25 grade, it is in disequilibrium and the extent of alteration is dependent upon protolith type
26 and water/rock ratio. Alteration in the RN-17B core bares many similarities to that of
27 Type II basalts observed in Mid Atlantic Ridge samples.

28

29 **Introduction**

30 Due to the technical challenges and great expense of seafloor drilling, only four drill
31 holes that exceed 1000 m into oceanic basement rock have been completed (Ildefonse et
32 al., 2007). The location of the Reykjanes geothermal system on the immediate subaerial
33 extension of the Mid Atlantic Ridge coupled with seawater recharge fluids make it a
34 useful analog for seafloor hydrothermal alteration (Spooner and Fyfe, 1973; Spooner,
35 1974). In contrast to seafloor systems, drill cutting samples are available to depths
36 exceeding 3000 m from the Reykjanes geothermal system thanks to the 38 geothermal
37 wells associated with a 100-megawatt geothermal power plant.

38

39 Important differences between Reykjanes and seafloor systems include evidence for a
40 period of meteoric fluid recharge during the Pleistocene, when a substantial icesheet was
41 present (Sveinbjörnsdóttir et al., 1986; Franzson, 2002; Pope et al., 2009; Marks et al.,
42 2010a). The Reykjanes geothermal system also lacks the hydrostatic pressure of
43 overlying cold seawater; therefore much greater depths must be reached before the
44 venting temperatures observed in seafloor hydrothermal systems are achieved. Another
45 proposed difference is that lithologies at depth in the Reykjanes geothermal system were
46 deposited in a shallow marine setting and are at the current depth due to high rates of

47 subsidence (Friðleifsson and Richter, 2010). With the exception of Friðleifsson and
48 Richter (2010), previous studies relied on drill cutting samples.
49
50 Drill core has only recently been recovered from three locations in the Reykjanes
51 geothermal system: a relatively unaltered ~2 m long core from RN-19 recovered at an *in-*
52 *situ* temperature between 250 and 260°C (Friðleifsson and Richter, 2010), the 9.3 m RN-
53 17B core that is the subject of this study, recovered at an *in situ* temperature of 345 °C
54 (Friðleifsson and Richter, 2010, Friðleifsson et al., 2011), and three sequential cores
55 totaling 22.5 m from RN-30 (estimated temperature based on depth to the boiling curve
56 approximately 365° C, although an *in-situ* temperature measured after the hole has
57 thermally equilibrated following drilling was not available at the time of this writing).
58 These 4” diameter cores permit the study of basalt-seawater interaction at high *in-situ*
59 temperatures.
60
61 Drill core preserves the textural relationships between hydrothermal minerals, which can
62 be recognized by sequential deposition in voids and fractures, replacement textures,
63 compositional zoning of mineral grains, and cross-cutting veins. Conversely, drill
64 cuttings from the same depth intervals are << 1/4” in diameter (thus preserving limited
65 spatial mineralogical relationships), represent depth intervals that are potentially
66 intermixed during ascent of the drilling fluid, and are returned from uncertain depth
67 intervals because recovery time lags behind the actual drill depth. Drill core provides an
68 accurate *in situ* depth record and provides the opportunity to evaluate spatial relationships
69 of alteration mineralogy to a degree not possible from drill cuttings. The recovery of drill

70 core from deeper portions of the Reykjanes geothermal system provides an important
71 opportunity to make comparisons with drill core recovered from submarine hydrothermal
72 systems collected by the Ocean Drilling Program (ODP).

73

74 The focus of this study is the hydrothermally altered RN-17B core. Well RN-17 (Figure
75 1) was drilled in the Reykjanes geothermal system to a depth of 2,266 m when the drill
76 rod broke, requiring a sidetrack hole to reach the targeted depth of 2,700 m (Friðleifsson
77 et al., 2005). RN-17 was completed as a vertical well with a total depth of 3082.4 m in
78 February 2005. RN-17 subsequently suffered a formation collapse during a flow test.
79 The well was re-entered in 2009 and sidetrack hole RN-17B was directionally drilled
80 (SSW; 35° from vertical) from a kick off point at approximately 920 m depth, and
81 eventually reached a down hole depth of 3077 m (Friðleifsson et al., 2005; Helgadóttir et
82 al., 2009; Friðleifsson and Richter, 2010). Drill cuttings were collected during the
83 drilling of RN-17 and RN-17B at 2 m intervals, and larger samples of cuttings intended
84 for geochemical analysis were collected at 50 m intervals. In addition, a 4-inch diameter
85 drill core was taken from a down-hole depth of 2,798.5 m to 2,807.8 m (ca. 2,560 to
86 2,570 m vertical depth) in RN-17B, at an *in-situ* temperature of 345°C (Friðleifsson and
87 Richter, 2010; Friðleifsson et al., 2011).

88

89 We provide details of the lithology and variable geothermal conditions in the Reykjanes
90 geothermal system at a resolution not previously possible using drill cutting samples. We
91 suggest that the lithologies present in the RN-17B core resemble hyaloclastite deposits
92 formed during submarine volcanic and subsequent epiclastic processes described for deep

93 marine portions of the Reykjanes Ridge. Fluid inclusion measurements from epidote
94 veins in the RN-17B core record temperatures cooling through the two-phase boundary
95 for seawater to current temperatures. Cooling of the geothermal system was coupled with
96 a decrease in water/rock ratio. Epidote veins in the RN-17B core do not record an early
97 period of meteoric alteration. The results of this study provide insight into evolving
98 hydrothermal process in a setting analogous to shallow submarine hydrothermal systems.

99
100 We support the conclusions of this study using: 1) a detailed lithological log of the RN-
101 17B core; 2) electron microprobe analyses of vein epidote, hydrothermal hornblende, and
102 altered relict igneous plagioclase; 3) homogenization temperature and freezing point
103 depression measurements of fluid inclusions hosted in epidote veins; and 4) strontium
104 isotope ratio ($^{87}\text{Sr}/^{86}\text{Sr}$) measurements of vein epidote using laser ablation multi collector
105 ICP-MS (LA-ICP-MS). We present and discuss our results our results in the context of
106 seafloor hydrothermal systems, and highlight key similarities and differences.

107 **2.0 Geological Setting**

108 The Reykjanes volcanic system is the subaerial continuation of the submarine Reykjanes
109 ridge. The volcanic system is hosted in an oblique extensional tectonic setting, resulting
110 in episodic fissure eruptions (Clifton and Kattenhorn, 2006). Tectonic extension-
111 controlled structures and active volcanism have resulted in significant geothermal activity
112 along the Reykjanes Peninsula, notably in the Reykjanes and Svartsengi geothermal
113 systems (Jakobsson et al., 1978, Arnórsson, 1995). The Reykjanes geothermal system is
114 located on the western tip of the Reykjanes Peninsula, approximately 50 km southwest of
115 Reykjavík (Figure 1).

116

117 The Reykjanes geothermal fluid is seawater, chemically modified due to boiling and
118 fluid/rock interaction (Tómasson and Kristmannsdóttir, 1972; Arnórsson et al., 1978;
119 Ólafsson and Riley, 1978; Arnórsson, 1995; Bird and Spieler, 2004; Friðleifsson and
120 Elders, 2005; Marks et al., 2010a). Compared to seawater, hydrothermal fluids in the
121 Reykjanes geothermal system have similar concentrations of Na and Cl; increased
122 concentrations of K, Ca, SiO₂, Fe, and Al; and decreased concentrations of SO₄ and Mg
123 indicating that mineral precipitation, and significant water rock interaction has occurred
124 (Lonker et al., 1993; Arnórsson, 1995; Pope et al., 2009).

125

126 Alteration zones based on the appearance of index minerals have been established for the
127 Reykjanes geothermal system (i.e., Tómasson and Kristmannsdóttir, 1972;
128 Kristmannsdóttir, 1975; Lonker et al., 1993; Franzson et al., 2002; Friðleifsson et al.,
129 2005; Marks et al., 2010a; Marks et al., 2011). The progressive alteration zones
130 determined from RN-17 drill cuttings are: smectite-zeolite (0-312 m), epidote-mixed
131 layer clay (MLC) (312-520 m), chlorite-epidote (520-1,200 m), epidote-actinolite (1,200-
132 2,350 m), and amphibole (2,350-3,082 m). Sidetrack hole RN-17B shows a similar
133 progression in alteration with depth. Localized high-grade granoblastic hornblende
134 hornfels and pyroxene hornfels zones resulting from contact metamorphism were
135 identified in drill cuttings at depths of 2,150 m and intervals in the depth range 2,600-
136 2,850 m from RN-17 (Marks et al., 2011).

137

138 Epidote-MLC, chlorite-epidote, and epidote-actinolite alteration zones in RN-17 and RN-

139 17B fall within the greenschist metamorphic facies, while the amphibole alteration zone,
140 defined by the presence of hornblende amphibole and calcic plagioclase, falls within the
141 amphibolite metamorphic facies (Marks et al., 2011). The transition from greenschist
142 assemblages to amphibolite assemblages coincides with a change in amphibole
143 composition from actinolite to hornblende accompanied with the presence of calcic
144 plagioclase, while epidote and chlorite remain stable (Marks et al., 2011). The RN-17B
145 core falls into the amphibole alteration zone. The alteration to calcic plagioclase +
146 magnesium hornblende + chlorite + titanite + albitized plagioclase + vein epidote and
147 sulfides is similar to that of Type II basalts observed in Mid Atlantic Ridge samples,
148 where it is proposed that a reaction zone is present where the composition of
149 hydrothermal fluids actively venting on the seafloor acquire their geochemical
150 characteristics (Gillis and Thompson, 1993).

151

152 The shallow depth at which greenschist-equivalent alteration assemblages are
153 encountered, relative to the present day temperature-depth relation controlled by a boiling
154 hydrostatic curve, indicates that the Reykjanes geothermal system experienced hotter
155 temperatures and higher pressures in the past (Sveinbjörnsdóttir et al., 1986; Franzson et
156 al., 2002; Friðleifsson et al., 2005; Marks et al., 2010a). The salinity of fluid inclusions
157 in quartz, plagioclase, and calcite in wells RN-9 and RN-10 (Franzson et al., 2002), along
158 with deuterium and oxygen isotope studies of fluids and alteration minerals, have been
159 used to suggest an evolution of hydrothermal fluids from meteoric to seawater (Ólafsson
160 and Riley, 1978; Sveinbjörnsdóttir et al., 1986; Lonker et al., 1993; Pope et al., 2009,
161 Marks et al., 2010b). The postulated evolving temperature conditions and fluid

162 compositions in the Reykjanes geothermal system have been attributed to the presence of
163 a substantial ice sheet overlying the Reykjanes Peninsula during the Pleistocene,
164 providing a mechanism for fresh water recharge and increased pressure, which raised the
165 boiling point curve and resulted in the formation of anomalously shallow, high-
166 temperature alteration minerals (Sveinbjörnsdóttir et al., 1986; Friðleifsson et al., 2005).

167 **3.0 Methods**

168 **3.1 Fluid Inclusions**

169 Fluid inclusion melting point (T_m) and homogenization temperature (T_h) measurements
170 were performed on fluid inclusions hosted by epidote crystals in veins at depths of
171 2,798.15 and 2,800.05 m in the RN-17B core. Salinity, as weight sodium chloride (wt. %
172 NaCl), was calculated from freezing point depression (FPD) measurements using the
173 equation of Bodnar (1993).

174
175 Fluid inclusion samples were prepared from doubly polished thin sections cut to a
176 thickness of 100 μm and mounted on glass slides at room temperature with acetone-
177 soluble epoxy. Fluid inclusions were identified and mapped on the prepared sections
178 prior to floating the samples from the glass slides by dissolving the epoxy with acetone.
179 T_h and T_m were measured using a Fluid INC. adapted USGS gas-flow heating/freezing
180 system equipped with a petrographic microscope, calibrated using synthetic H_2O fluid
181 inclusions trapped at the critical point of water and synthetic CO_2 fluid inclusions.
182 Homogenization of fluid inclusions was verified by cycling the temperature to ensure the
183 vapor bubble did not readily nucleate. Care was taken to keep measured fluid inclusions
184 within 1 mm of the thermocouple tip to minimize irreproducibility due to thermal
185 gradient effects. Repeated measurements of the standards reproduced the known T_h by
186 $\pm 0.2^\circ\text{C}$ (which equates to ± 0.4 wt. % NaCl), and T_h by $\pm 0.3^\circ\text{C}$, a similar range to that
187 reported by Kelley and Delaney (1987). The precision of our T_m measurements is
188 consistent with that reported in previous studies (Kelley and Delaney, 1987; Vanko,
189 1988). Results on Figure 5 are presented using a more conservative estimate for T_h error

190 of $\pm 5^{\circ}\text{C}$, which accounts for $\pm 2^{\circ}\text{C}$ thermal gradient effects due to sample distance from
191 the thermocouple, $\pm 1^{\circ}\text{C}$ for thermocouple accuracy, and $\pm 2^{\circ}\text{C}$ for bracketing of the
192 homogenization temperature (Vanko, 1988). Upon completion of T_m and T_h
193 measurements, fluid inclusion chips were mounted on 1-inch diameter Plexiglas™ rounds
194 with double-sided sticky tape for major element and strontium isotope analyses.

195

196 **3.2 Major Element Mineral Chemistry**

197 Mineral identification and major element analyses were performed at UC Davis using a
198 Cameca SX-100 electron microprobe (EMP). Mineral identification was completed
199 qualitatively on carbon-coated thin sections using the energy-dispersive spectrometry
200 (EDS) capabilities of the EMP. Quantitative EMP analyses were performed on epidote
201 crystals hosting fluid inclusions, vein epidote, and zoned epidote crystals. EMP analyses
202 were performed on amphibole in altered hyaloclastite shards, the altered quenched
203 margin of a crystalline clast, amygdales, vein selvages, and altered crystalline clasts.
204 EMP analyses were also performed on relict igneous plagioclase phenocrysts in altered
205 crystalline clasts.

206

207 Epidote analyses were performed using an accelerating voltage of 15 kV, a beam current
208 of 15 nA (100 nA for Sr), a nominal beam diameter of 1 μm , and a peak counting time of
209 10 seconds (60 seconds for Sr). The calibration standards included Ti (Taylor rutile), Fe
210 (Taylor hematite), Ca, Al, and Si (Smithsonian anorthite), Mg (tremolite from
211 Campolungo, Switzerland), Mn (rhodonite; Harvard collection #104791), and Sr
212 (SrAl_2Si_2 synthesized by Naohito Tsujii). The mole fraction of Fe (X_{Fe}) was calculated

213 using the equation $X_{\text{Fe}} = \text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Al}^{3+})$. The nomenclature X_{Fe} in lieu of pistacite
214 component (X_{Ps}) is used in this study, after Armbruster et al. (2006).

215

216 Amphibole analyses were performed using an accelerating voltage of 15 kV, beam
217 current of 10 nA, a nominal beam diameter of 5 μm , and a peak counting time of 10
218 seconds. The calibration standards included Ti (Taylor rutile), Ca (Taylor wollastonite),
219 Fe (Taylor hematite), Al (Smithsonian anorthite), Mg and Si (tremolite from
220 Campolungo, Switzerland), Na (jadeite from Clear Creek, California), Mn (rhodonite;
221 Harvard collection #104791), and K (orthoclase).

222

223 Plagioclase analyses were performed using an accelerating voltage of 15 kV, beam
224 current of 10 nA, a nominal beam diameter of 5 μm , and a peak counting time of 10
225 seconds. The calibration standards included Na and Si (Taylor albite), Ca (Taylor
226 wollastonite), Fe (Taylor hematite), Al (Smithsonian anorthite), and K (orthoclase).

227

228 **3.3 Strontium Isotope Chemistry**

229 ^{88}Sr , ^{87}Sr , ^{86}Sr , ^{85}Rb , and ^{84}Sr were measured simultaneously in epidote crystals hosting
230 measured fluid inclusions, and in transects across a vein from 2800.05 m using laser
231 ablation multi-collector inductively coupled plasma mass spectrometry (LA-MC-ICP-
232 MS). Analyses were performed at UC Davis using a Thermo Scientific Neptune
233 multicollector coupled to a Photon Instruments laser ablation system equipped with a 193
234 nm excimer laser and a dual volume sample cell. Helium was used as the aerosol carrier
235 gas and mixed with argon prior to introducing ablated aerosol into the plasma torch. A

236 flow rate of 1.3 liters/minute was obtained for argon, the outer containment of the dual
237 cell had a helium flow rate of 0.75 liters/minute, and the inner cup had a helium flow rate
238 of 0.13 liters/minute. Laser conditions consisted of a laser spot size of 60 μm , 8 Hz
239 repetition rate and 1.8 J/cm² fluence.

240

241 An internal deep-sea coral standard (DWC; modern coral with homogenous $^{87}\text{Sr}/^{86}\text{Sr}$
242 composition verified previously by standardization to marine fish otoliths) was used to
243 calibrate the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of seawater. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of fresh volcanic rock (not
244 exposed to seawater) was calibrated using plagioclase phenocrysts in rhyolite magma
245 quenched to glass, obtained from the IDDP-01 well in the meteoric water-dominated
246 Krafla geothermal field, Iceland (Zierenberg al., 2013). Plagioclase in the Krafla sample
247 was chosen as a standard to replicate as closely as possible the silicate laser ablation
248 characteristics expected from epidote. Standards were analyzed at the start of the sample
249 run and after every 10 samples.

250

251 Background concentrations were measured for a period of 1 minute prior to sample
252 ablation. Samples were ablated for a minimum of 200 cycles. Individual sample counts
253 were excluded from the analysis if they fell outside of a 2 standard error range. The
254 background signal and ^{87}Rb interference were automatically subtracted and corrected for
255 by Neptune software.

256 **4.0 Results**

257 **4.1 Lithology**

258 Preliminary lithological observations for the RN-17B core are described in Friðleifsson
259 and Richter (2010) and Marks et al. (2010a, 2010b). A detailed core log is included as an
260 electronic appendix. Four distinct lithologies are present in the core: pillow basalt
261 breccia, hyaloclastite breccia, lithic breccia, and volcanic sandstone. Two types of basalt
262 are also present in the core: plagioclase-phyric basalt with 2-3 mm long altered igneous
263 plagioclase phenocrysts; and olivine-phyric basalt with microcrystalline (<0.05 mm long)
264 plagioclase laths and sparse 0.5 -1 mm chloritized olivine phenocrysts. For ease of
265 description, the core has been divided into five units. The stratigraphic top of each unit is
266 defined by the presence of poorly laminated volcanic sandstone that exhibits
267 characteristics of transport from the sediment source area (Figure 2).

268

269 Unit 1 is a clast-supported homolithitic hyaloclastite breccia composed of vesicular and
270 non-vesicular 20 to 80 mm plagioclase-phyric basalt clasts. Clasts are predominantly
271 composed of felted or feathery green hornblende with fine titanite, and dusty, albitized
272 igneous plagioclase laths. Vesicles in clasts are 1 to 4 mm in diameter, rimmed by green
273 acicular hornblende and dominantly filled with dusty white albite and minor anhedral
274 epidote. Basalt clast partly surrounded by quenched margins and are in a matrix of
275 abundant 0.1 to 1.5 mm blocky hyaloclastite glass shards altered to coarse, felted to radial
276 brown hornblende with fine titanite. Some sand-sized grains of crystalline basalt are also
277 present in the hyaloclastite, identifiable by the presence of large, altered igneous
278 plagioclase laths. The hyaloclastite matrix is sealed with dusty white albite and sparse,

279 anhedral epidote and acicular green hornblende. This unit is interpreted as a pillow basalt
280 breccia with limited down-slope transport from the site of eruption. Representative
281 textures described in the text are indicated on Figure 3.

282

283 Unit 2 is a clast-supported homolithic hyaloclastite breccia with plagioclase-phyric basalt
284 clasts ranging in size from 3 mm to greater than the core diameter. A plagioclase-phyric
285 basalt pillow is present within the hyaloclastite breccia at ~2800.4 m. Alteration in the
286 quenched margins of plagioclase-phyric basalt clasts consists of coarse felted green
287 hornblende + titanite. Calcic plagioclase occurs in <1 mm veins in the basalt pillow.
288 Alteration in the crystalline centers of plagioclase-phyric clasts consists of green
289 hornblende in a chloritic matrix intergrown with sparse, microcrystalline calcic
290 plagioclase. Vesicles, which are uncommon, measure 0.1 to 0.2 mm diameter and are
291 filled with fine white acicular actinolite and epidote: some retain open space. Basalt clast
292 margins are chilled against a matrix of abundant blocky 0.1 to 1.5 mm hyaloclastite glass
293 shards. Shard edges are altered to coarse, felted to radial brown hornblende + titanite
294 with sub-spherical epidote centers. Rare, uralitized pyroxene grains are distributed
295 throughout the hyaloclastite matrix. Unit 2 is capped by lithic volcanic sandstone with
296 crudely defined graded laminations. The sandstone is composed of glass shards and
297 crystalline grains ranging in size from 0.3 to 1 mm and rare fragmented vesicular basalt
298 clasts up to 20 mm wide with chilled margins on one side. This section of hyaloclastite,
299 pillow breccia, and the apparently intact pillow lobe may have erupted in-situ.

300

301 Unit 3 is a matrix supported heterolithic breccia with hyaloclastite matrix grading down

302 into a matrix of sand-sized crystalline basalt grains at 2802.8 m, and back into
303 hyaloclastite matrix at 2805.2 m. Breccia clasts are sand-sized to 40 mm and are
304 composed of both plagioclase-phyric and olivine-phyric basalt. Clasts appear
305 fragmented, generally with chilled margins on one side only. The matrix is composed of
306 detrital albitized plagioclase, chloritized olivine, hyaloclastite shards, and rounded sand-
307 sized grains of basaltic composition. Alteration in clasts, matrix, and amygdales is
308 similar to that in Units 1 and 2.

309

310 Unit 4 consists of crudely bedded volcanic sandstone overlaying a vesicular, plagioclase-
311 phyric basalt pillow at 2807.5 m. The volcanic sandstone contains bivalve shell
312 fragments that have been replaced by epidote (Friðleifsson and Richter, 2010). The
313 crudely laminated volcanic sandstone is in contact with the pillow at an angle 55° to the
314 core axis. Alteration in the center of the pillow consists of albitized relict igneous
315 plagioclase laths, feathery green hornblende + fine titanite, abundant pyrite intergrown
316 with magnetite, and rare patches of chlorite. Alteration on the edge of the pillow consists
317 of coarse, felted green hornblende + fine titanite, and altered relict igneous plagioclase
318 laths. Vesicles are similar to those in Unit 1. Below the pillow, at the base of the unit, is
319 a volcanic breccia composed of plagioclase-phyric basalt clasts. Poorly defined bedding
320 in the volcanic sandstone, the depositional contact of the hyaloclastite on the underlying
321 pillow basalt, and a geopetal fill in a void interpreted to be a replaced bivalve shell, all
322 dip at an angle of ~ 55° to the core axis, which suggests these features have not been
323 significantly tilted from horizontal given that the drill hole is inclined at 35°.

324

Unit 5 is fine-grained volcanic sandstone with occasional hyaloclastite shards. Unit 5 has crude laminations oriented 55° to the core axis at the top, grading down into coarser grained sedimentary breccia with no laminations and clasts up to 5 mm. Amygdales in clasts are filled with epidote. Sedimentary voids and matrix are filled with albite, acicular green hornblende and epidote. The base of the unit consists of large clasts with chilled margins.

The core is riddled with epidote-mineralized veins and fractures <1 to 20 mm wide, which are inclined $\sim 30^{\circ}$ or $\sim 50^{\circ}$ relative to the core axis. Although the core is not oriented, these vein orientations likely correspond approximately to vertical and horizontal veins, respectively. Thicker veins are more abundant in Units 1 and 2 and have an outer generation of clear epidote + pyrite and an inner generation of yellow epidote, as observed in thin section. Green acicular hornblende + epidote $\pm$ chlorite $\pm$ albite occur in vein selvages. Fracture surfaces in Unit 2 are coated with fibrous white actinolite (identified by refractive index) + pyrite + chalcopyrite + rare sphalerite. Discontinuous veins of epidote + minor albite up to 6 mm wide with chlorite selvages are cut by steeper 1 mm wide epidote + pyrite veins with chlorite selvages that are oriented $\sim 60^{\circ}$ to the core axis.

4.2 Fluid Inclusions

Fluid inclusions were identified in epidote samples from 2799.15 m and 2800.05 m. Fluid inclusions present parallel to cleavage along the {001} crystallographic axis were elongate and frequently tapered, some were planar in shape (Figure 4A). Fluid inclusions concentrated in concentric growth rings perpendicular to the {001} crystallographic axis

348 were generally equant and planar, some were tapered (Figures 4B and 4C). Concentric
349 growth zoning was identified in transmitted light based on fluid inclusion densities and
350 slight color variations in the epidote. Fluid inclusions associated with crystal growth
351 features (concentric zoning and crystal elongation) were assumed to be primary in origin.
352 Primary fluid inclusions ranged in size from <1 to $10\text{ }\mu\text{m}$ in diameter. Secondary fluid
353 inclusions were present in healed, curvilinear and irregular fractures that crosscut crystal
354 growth features (Figures 4B and 4C). Secondary fluid inclusions were $<1\text{ }\mu\text{m}$ in diameter
355 and not measurable. Both liquid and vapor rich fluid inclusions were present. The
356 proportion of liquid to vapor varied according to the location of the epidote crystal
357 hosting the fluid inclusion within the vein. Primary fluid inclusions adjacent to the vein
358 wall typically had $\sim 60\%$ vapor, although a few were as low as $\sim 30\%$ vapor (Figure 4A).
359 Primary fluid inclusions inwards of the vein walls ranged from liquid dominated with
360 $\sim 30\%$ vapor, to vapor dominated with almost the entire cavity filled by vapor (Figure
361 4B). Vapor dominated fluid inclusions with $>80\%$ vapor volume could not be measured.
362 Primary fluid inclusions in the vein center were liquid dominated with ~ 15 to 20% vapor.
363
364 T_h and T_m measurements were performed for fifty-seven epidote vein-hosted primary
365 fluid inclusions (Table 1; Figure 5). In some cases epidote crystals darkened on heating
366 and T_h could not be observed, and in other cases fluid inclusions were too small to
367 observe T_m . Calculated salinities for individual fluid inclusions ranged from 0.5 to 7.6 wt.
368 $\%$ NaCl. T_h measurements for individual fluid inclusions ranged from 345 to $>480^\circ\text{C}$. A
369 pressure correction for T_h measurements was evaluated based on the pressure equal to the
370 hydrostatic head of a boiling water column at the depth of the samples, using

371 experimental results for the two-phase boundary of seawater (Bischoff and Rosenbaur,
372 1985). The pressure at the depth of the RN-17B samples (~2,560 m) was estimated to be
373 approximately 185 bars. The trapping temperature (T_t) corrections of Potter (1977) are
374 not well defined for the measured fluid inclusion T_h and salinity range, therefore fluid
375 inclusion T_h results are reported uncorrected.

376

377 Calculated salinity values and T_h measurements of fluid inclusions followed a spatial
378 pattern and fell into 3 categories that corresponded to fluid inclusion petrography: 1) a
379 mixture of vapor-rich and liquid dominated inclusions with low salinity and T_h of
380 383.6°C to 401.5°C adjacent to vein walls; 2) a mixture of vapor and liquid dominated
381 inclusions with low salinity and $T_h > 480^\circ\text{C}$ inwards of the vein wall; and 3) liquid
382 dominated fluid inclusions with T_h of 344.9°C to 378.3°C in vein centers with high
383 salinities (Figure 5). All three categories of fluid inclusions were present in the sample
384 from 2798.64 m; only category 2 and 3 fluid inclusions were present in the sample from
385 2800.05 m.

386 **4.3 Major Element Chemistry**

387 EMP analyses of epidote are presented in Table 2. Vein epidote crystals are euhedral and
388 generally compositionally zoned. Major element concentrations measured by EMP vary
389 within each zone. Epidote analyses in Table 2 represent the centers of each epidote
390 crystal where measured primary fluid inclusions occur. OH concentrations are not
391 reported in Table 2, but they are included in the microprobe totals, as the number of OH
392 ions is fixed at 2.0. The X_{Fe} of epidote hosting fluid inclusions ranged from 0.17 to 0.37.
393 The X_{Fe} component in compositionally zoned epidote crystals is both normally (high X_{Fe}

394 cores with low X_{Fe} rims) and reversely zoned, with no consistent spatial pattern.
395 However, vein epidote adjacent to the wall rock have a higher aluminum concentration
396 than the vein centers (Figure 6).
397
398 EMP analyses of amphibole are presented in Table 3. Back-scattered electron (BSE)
399 images reveal various morphologies of hydrothermal amphibole in the RN-17B core.
400 Fine felted amphibole is present in chilled margins and crystalline edges of basalt
401 fragments (Figure 7A); larger, isolated amphibole crystals occur in the centers of basalt
402 clast fragments (Figure 7B), acicular amphibole occurs in the edges of vesicles (Figure
403 7C) and veins, and coarse felted amphibole replaced hyaloclastite shards (Figure 7D).
404 Amphibole EMP analyses are corrected for ferric iron content according to the method of
405 Dale et al. (2005). Results confirm that the amphibole is calcic and spans a compositional
406 range from actinolitic hornblende to pargasitic hornblende (Figure 8). Despite the green
407 acicular, actinolitic appearance of amphibole in veins and vesicles, analyses confirm it is
408 hornblendic in composition (Figure 8). Amphibole analyses from the RN-17B core are
409 consistent with amphibole analyses from RN-17 and RN-17B drill cuttings published by
410 Marks et al. (2011), although RN-17B core samples do not span into actinolitic
411 compositions along the pargasite exchange vector. Amphibole compositions vary
412 spatially in a plagioclase-phyric clast with chilled margins from 2,800.05 m (Figures 7A
413 and 7B, Figure 8, and Figure 9). Hornblende in the crystalline center of a clast from
414 2,800.05 m is more actinolitic in composition than the outer crystalline edge of the clast,
415 which is more hornblendic to pargasitic in composition. Hornblende in the altered
416 quenched margin falls in between the compositions for the crystalline center and outer

417 crystalline clast edge (Figure 8). The Fe concentration is lower and the Mg concentration
418 is higher for the altered quenched margin of the clast and altered hyaloclastite shards in
419 the matrix than for the outermost crystalline portions of the clast, while the crystalline
420 center of the clast plots across a wide spectrum of FeO-MgO concentrations (Figure 9).

421

422 EMP analyses of plagioclase are presented in Table 4, and representative plagioclase
423 textures are presented on 10A through 10D. The majority of plagioclase analyses are
424 biased towards plagioclase adjacent to hornblende amphibole, which is predominantly
425 very calcic, although two analyses of more sodic plagioclase were obtained. EMP
426 analyses were performed on plagioclase in epidote vein selvages present at 2799.90 and
427 2800.35 m (samples B-plag 1 and 2; C-plag 1, 2, and 3), relict igneous plagioclase laths
428 present at 2804.30 m (samples F1-plag 1 and 2), the center of an altered volcanic clast
429 present at 2804.5 m (samples G2-plag 1, 2, and 3), and the matrix between altered
430 hyaloclastite shards present at 2804.80 m (samples H1-plag 1, 2, 3 and 4; H2-plag 1, 2, 3,
431 and 4).

432

433 EMP analyses indicate that plagioclase compositions fall into two distinct groups: sodic
434 plagioclase (An_{23} to An_{26}) and calcic plagioclase (An_{82} to An_{99}). The more sodic
435 plagioclase appears darker in BSE images than the lighter calcic plagioclase.

436 Hydrothermal plagioclase in veins and vein selvages was more calcic, and ranged from
437 An_{86} to An_{98} (Figure 10A). Relict igneous plagioclase, present in the crystalline cores of
438 clasts, consists of intergrown calcic (An_{82}) and sodic plagioclase (An_{23}) and appears
439 albitized in thin section and (Figure 10B). Plagioclase in the core of an altered volcanic

440 clast is calcic (An₉₇ and An₉₈), and forms as a ~ 20 micron rim between a patch of
441 chlorite and the surrounding hornblende amphibole (Figure 10C). Plagioclase present in
442 the hyaloclastite matrix, adjacent to hyaloclastite shards altered to hornblende, consists of
443 intergrown calcic (An₈₅ to An₉₉) and sodic (An₂₆) plagioclase (Figure 10D).

444 **4.4 Strontium Isotope Chemistry**

445 Results for ⁸⁷Sr/⁸⁶Sr epidote vein transects on a sample from 2800.05 m are presented in
446 Figure 11. Plagioclase phenocrysts in the Krafla glass fall in the expected ⁸⁷Sr/⁸⁶Sr ratio
447 range of unaltered basalt (0.7030 – 0.7034; O’Nions and Grönvold, 1973; Sun and Jahn,
448 1975) and overlapped with the value of the sample (0.70328) determined from whole
449 rock dissolution (Elders et al. 2011; Zierenberg et al., 2013). The DWC standard closely
450 matched the expected ⁸⁷Sr/⁸⁶Sr ratio of seawater (0.7092; Dasch et al. 1973; Burke et al.
451 1982; Capo and DePaolo, 1992). The reproducibility of ⁸⁷Sr/⁸⁶Sr on varying substrates
452 supports negligible matrix effects of the laser technique for epidote. ⁸⁷Sr/⁸⁶Sr ratios in
453 epidote veins were higher on the vein edges (~0.70500) compared to the vein centers
454 (~0.70400), and overlap the present-day strontium isotope composition of geothermal
455 fluids from the Reykjanes Geothermal system (0.70412; Millot et al., 2009). The
456 ⁸⁷Sr/⁸⁶Sr ratios in vein epidote from this study (Table 2; Figure 11) are overall shifted
457 towards seawater values, compared to basaltic values, consistent with measurements of
458 epidote in drill cutting samples from RN-17 and RN-17B (Marks et al., 2010b). Lower
459 ⁸⁷Sr/⁸⁶Sr ratios in the vein centers are consistent with the presence of a second generation
460 of epidote observed petrographically in vein centers throughout the RN-17B core (inner
461 vein more yellow in color) and using back scattered electron (BSE) imaging (inner vein
462 brighter and coarser texture). The signal intensity of ⁸⁸Sr increased coincident with the

second generation of epidote, indicating a higher Sr concentration in the vein centers (Figure 11).

Results for epidote crystals hosting fluid inclusions are overall shifted towards seawater $^{87}\text{Sr}/^{86}\text{Sr}$ values (Table 2), consistent with evolved seawater being trapped in both low salinity (<3.2 wt% NaCl) and high salinity (>3.2 wt% NaCl) fluid inclusions. No obvious spatial trend in $^{87}\text{Sr}/^{86}\text{Sr}$ was observed for these samples. This is possibly due to numerous wall rock and acicular green amphibole inclusions present in the samples from 2799.15 m. In contrast, the vein at 2800.05 m in which the epidote vein transects are measured has few wall rock or amphibole inclusions.

Marks et al. (2010b) found no relationship between the Sr concentration measured by EMP and the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio measured by LA-MC-ICP-MS for drill cuttings from different depth intervals from RN-17B, or on the scale of individual zoned epidote grains. We found the same lack of correlation on the scale of individual epidote crystals present in veins. However, the signal intensity for ^{88}Sr measured by LA-MC-ICP-MS does appear to correlate with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in epidote vein transects (Figure 11). The Neptune software Rb correction algorithm was not responsible for the higher ^{88}Sr signal in the later vein generation, as the measured ^{85}Rb intensity did not follow a similar pattern. The conflicting correlation between Sr signal intensity measure by LA-MC-ICP-MS and the Sr concentration measured by EMP may be due to heterogeneous distribution of Sr within individual epidote crystals. The smaller EMP spot size (1 micron) likely reflects Sr heterogeneity within single epidote crystals, while the larger (60 micron) LA-MC-ICP-

486 MS spot size likely reflects the average Sr concentration in the crystals.

487 **5.0 Discussion**

488 The evidence for epidote paragenesis ranging from older on the vein edges to younger in
489 the vein center is supported by three main lines of evidence. Different generations of
490 epidote are apparent petrographically based on the variation in color and crystal
491 morphology. Fluid inclusion homogenization temperatures and salinities in vein centers
492 trend toward the current *in-situ* measured temperature (345°C; Friðleifsson et al., 2011)
493 and the modern salinity of Reykjanes fluids (Figure 5). Additionally, the $^{87}\text{Sr}/^{86}\text{Sr}$ value
494 of epidote in vein centers span the $^{87}\text{Sr}/^{86}\text{Sr}$ of modern Reykjanes fluids (Figure 11).
495 Results from this study show systematic spatial geochemical variability in alteration
496 minerals and indicate evolving temperature and fluid composition with time,
497 accompanied by decreased water/rock ratio recorded by epidote veins in the RN-17B
498 core. These results reflect changes over time in a very localized portion of the Reykjanes
499 geothermal system. Our fluid inclusion and strontium isotope results for epidote in the
500 RN-17B core are similar to fluid inclusions reported from Mid Atlantic Ridge samples
501 (Kelley and Delaney, 1987) and strontium isotope studies of core 504B from the East
502 Pacific Rise (Alt et al., 1996; Teagle et al., 1998; Bach et al., 2003).

503

504 **5.1 Geological Setting**

505 The lithologies present in the RN-17B core are characteristic of hyaloclastite deposits
506 formed during submarine volcanic and subsequent epiclastic processes. Friðleifsson and
507 Richter (2010) interpret the high vesicularity of some clasts, the formation of pillow
508 lavas, and the presence of shell fragments as evidence for deposition of the lithologies in
509 shallow marine conditions.

510

511 Several authors have attempted to link higher degrees of basalt vesicularity with lower
512 hydrostatic pressure related to shallower seafloor depths (i.e. Moore, 1965; Moore, 1979;
513 MacPherson, 1984). However, several samples of highly vesicular basalt have been
514 recovered from seafloor depths of >4000 m in Hawaii (Clague et al., 1990), depths up to
515 3800 m on the Gorda mid-ocean ridge (Clague and Davis, 2003), depths of 1300 to
516 >3400 m offshore central California (Davis and Clague, 2003), and depths between 400
517 and 1700 m on the Mid-Atlantic Ridge (Fouquet et al., 1998; Eissen et al, 2003). Results
518 of these studies suggest that hydrostatic pressure of seawater alone is not sufficient to
519 account for the degree of vesicularity observed in many types of deep submarine basalt.
520 The CO₂ volatile component of basaltic magmas erupting on the seafloor likely
521 determines the explosiveness and vesicularity of submarine eruptions (Davis and Clague,
522 2003; White et al., 2003). Also, observations at mid-ocean ridges indicate the
523 morphology of lava flows is related to the spreading rate (rather than hydrostatic
524 pressure), with pillow lava formation favored at slow spreading ridges (Perfit and
525 Chadwick, 1998). Therefore, the presence or absence of pillow lavas in the RN-17B core
526 may not be a reliable indicator of eruption depth.

527

528 We suggest the characteristics of the RN-17B core are not necessarily indicative of a
529 shallow marine eruptive setting, and coupled with the great vertical depth (2500 m) of the
530 drill core, may be indicative of deeper marine eruptive setting. Within the RN-17B core,
531 Units 2 and 4 contain vesicular volcanic clasts with margins that appear quenched against
532 the enclosing hyaloclastite matrix, indicating the units are preserved *in situ*. Fragmented

533 clasts in Units 1 and 3 with quenched margins on only one side indicate slumping and
534 redeposition, possibly in a debris-flow. Units 2, 3, 4, and 5 are all capped with crudely
535 bedded volcanic sandstone, some with shell fragments and clasts with chilled margins on
536 one side only, indicating epiclastic deposition.

537

538 Similar deep submarine deposits have been described for Pacific Ocean seamounts
539 (Batiza et al., 1984), for the Mid Atlantic Ridge (Schminke et al., 1979), and for fossil
540 seamounts in Hokkaido, Japan (Yamagishi, 1991). Batiza et al. (1984) described crudely
541 bedded hyaloclastite deposits accumulating in depressions present at the crests of
542 seamounts. Schminke et al. (1979) argued that the Mid-Atlantic Ridge hyaloclastite
543 deposits formed by a combination of pillows slumping on the sides of pillow basalt
544 mounds and seawater currents preferentially carrying glass shards, sand-sized crystalline
545 fragments, and clasts with chilled margins from rapidly erupting basalt and depositing
546 them nearby. Yamagishi (1991) described epiclastic volcanic deposits on the flanks of
547 fossil seamounts with crudely graded volcanic sand lithologies, interpreted to have
548 formed during relatively quiescent interludes between eruptions.

549

550 An analogy for a likely depositional setting capable of creating the lithological features
551 observed in the RN-17B core is the modern offshore portions of the Reykjanes Ridge.
552 Parson et al. (1993) mapped submarine portions of the Reykjanes Ridge using high-
553 resolution sidescan sonar, seismic profiles, and magnetic data. Parson et al. (1993)
554 describe various morphologies of the submarine Reykjanes Ridge as a series of en
555 echelon, 5 to 35 km long and 3 to 15 km wide axial volcanic ridges (AVR's) composed

556 of overlapping volcanic edifices, often intersected by fault scarps trending between 0.5
557 km and 10 km long with vertical offsets up to 200 m. Parson et al. (1993) interpret the
558 AVR's as having an early volcanic constructive period followed by a later, tectonically
559 destructive period. Such a depositional setting could produce both the *in situ* eruptive
560 units and re-deposited lithologies observed in RN-17B.

561 **5.2 Fluid Inclusions in Epidote**

562 Fluid inclusion T_h and salinity measurements from RN-17B (0.5 – 7.6 wt.% NaCl) are
563 generally similar to those measured in ophiolite and ocean floor drillcore (i.e. Kelley and
564 Delaney, 1987; Vanko, 1988; Nehlig, 1990; Vanko and Stakes, 1991; Kelley, 1997;
565 Kelley and Früh-Green, 2001). However, several studies report fluid inclusion suites in
566 seafloor hydrothermal systems from drill core with salinities up to and exceeding 50
567 wt.% NaCl (Kelley and Delaney, 1987; Vanko, 1988; Nehlig, 1990; Kelley, 1997; Kelley
568 and Früh-Green, 2001). In comparison, the salinity of fluids expelled from active seafloor
569 hydrothermal vents typically range from 40% less to 30% greater (~1.9 to 4.5 wt.%
570 NaCl) than the 3.2 wt.% NaCl salinity of seawater (Bischoff and Rosenbaur, 1989;
571 Nehlig, 1990 and references therein). Studies of seafloor hydrothermal systems
572 commonly report some fluid inclusion suites with salinities less than the salinity of
573 seawater (Kelley and Delaney, 1987; Vanko, 1988; Nehlig, 1990; Vanko and Stakes,
574 1991; Kelley, 1997; Kelley and Früh-Green, 2001), yet the geological setting precludes
575 the possibility of meteoric recharge.

576

577 In terrestrial hydrothermal systems and shallow seafloor hydrothermal systems, fluid
578 properties are generally governed by the two-phase boiling curve of hydrothermal fluids

579 (Bischoff and Rosenbaur, 1985; Von Damm, 1995). Pressure conditions at oceanic
580 spreading centers are typically in the range of ~200 to 750 bars, which corresponds to
581 temperatures of 300 to 500°C; spanning the critical point (298.5 bars and 407°C) of
582 seawater (Bischoff and Rosenbaur, 1988). Seawater crossing the two-phase boiling
583 curve below the critical point under these conditions would produce a small amount of
584 vapor with much lower concentrations of dissolved salts and metals than the liquid phase
585 (Von Damm, 1995). In contrast, conditions in deep seafloor hydrothermal systems may
586 well exceed the critical point of the hydrothermal fluid due to the large hydrostatic
587 pressure resulting from the ocean mass (Bischoff and Rosenbaur, 1985). Seawater
588 intersecting the two-phase region from above the critical point would remain mostly in
589 the vapor phase and condense a small amount of dense brine (Bischoff and Rosenbaur,
590 1985; Von Damm, 1995). This model fits fluid inclusion evidence from seafloor
591 hydrothermal systems (Bischoff and Rosenbaur, 1985; Kelley and Delaney, 1987;
592 Nehlig, 1990; Vanko and Stakes, 1991; Kelley, 1997; Kelley and Früh-Green, 2001).
593 Some authors have also suggested that the formation of saline brines in seafloor
594 hydrothermal systems can be attributed to direct exsolution of fluids from cooling
595 magma, although direct evidence of this has not been reported (Kelley and Früh-Green,
596 2001 and references therein).

597

598 Fluid inclusion homogenization temperatures and the hydrostatic pressure at the depth of
599 the RN-17B core suggest that the critical point of seawater was not exceeded. Any phase
600 separation of hydrothermal fluids in the Reykjanes system is likely the result of the liquid
601 to vapor phase transition of seawater below the critical point, thus we would not expect to

602 see highly saline fluid inclusions at the depth of the RN-17B core. A key difference
603 between the Reykjanes geothermal system and seafloor hydrothermal systems is that
604 significantly greater depth is required at Reykjanes to approach similar hydrostatic
605 pressure conditions seen in seafloor hydrothermal systems. This is a consequence of the
606 cold hydrostatic pressure of overlying ocean water present in seafloor systems being
607 absent in the Reykjanes system, which instead has a pressure regime dictated by the hot
608 hydrostatic pressure of seawater.

609

610 The spatial relationships in RN-17B epidote veins – specifically that the least saline fluid
611 inclusions are found adjacent to wall rocks as opposed to vein interiors - imply that
612 Category 1 inclusions formed during the earliest stage of epidote precipitation. The low
613 salinities of Category 1 inclusions are less than and approaching the salinity of seawater,
614 and T_h exceeds the temperature of the two-phase boundary for seawater at the estimated
615 hydrostatic pressure for boiling seawater at the sample depth (Figure 12). These results
616 suggest an early stage of boiling due to increased temperatures during the initial
617 formation of the epidote veins.

618

619 Non-homogenizing fluid inclusions (Category 2 inclusions) exhibit salinities approaching
620 seawater, but will not homogenize below the maximum attempted temperature of 480°C.
621 No evidence of decrepitation or the presence of CO₂ was observed, however, the small
622 size of fluid inclusions could possibly obscure the presence of a secondary liquid phase.
623 Anomalous fluid inclusions trapping liquid plus steam effervescing through liquid due to
624 boiling at greater depth would produce $T_h > T_l$ (Roedder, 1984). Evidence for boiling in

625 adjacent, earlier Category 1 fluid inclusions provides support for interpretation of non-
626 homogenizing fluid inclusions forming as a result of boiling seawater below the critical
627 point.

628

629 Category 3 inclusions are present in vein centers and formed during the final stages of
630 vein precipitation. Category 3 inclusions are liquid dominated, have higher salinities than
631 seawater, and lower homogenization temperatures than paragenetically earlier fluid
632 inclusions. Temperatures in Category 3 inclusions trend towards the most recently
633 measured downhole temperature in RN-17B (345°C; Friðleifsson et al., 2011). These
634 inclusions possibly trapped a slightly more saline liquid residue formed from boiling of a
635 liquid with a significant seawater component. This hypothesis is also supported by the
636 fact that homogenization temperatures for the more saline Category 3 fluid inclusions fall
637 below the boiling curve (Figure 12).

638

639 Fluid inclusions trapped during the filling of epidote veins in the RN-17B core appear to
640 record an episode of boiling seawater followed by incursion of a more saline residue.

641 Initial boiling is preserved in Category 1 fluid inclusions based on presence of vapor-rich
642 fluid inclusions with salinities less than that of seawater. Continued boiling is preserved
643 in Category 2 fluid inclusions, which will not homogenize, indicating the presence of two
644 primary fluid phases (steam and liquid) trapped out of equilibrium. A later cooler, more
645 saline residue is preserved in the liquid-dominated Category 3 fluid inclusions. Our fluid
646 inclusion T_h and salinity measurements indicate that epidote veins in the RN-17B core

647 record an episode of boiling and subsequent cooling below the two-phase boundary for
648 RN-17B hydrothermal fluids.

649

650 Fluid inclusion results from epidote veins in the RN-17B core are remarkably similar to
651 those reported by Kelley and Delaney (1987) for altered gabbro samples recovered from
652 the western median valley walls of the Mid Atlantic Ridge near 24°N. Kelley and
653 Delaney (1987) reported an early episode of brittle fracture recorded by fluid inclusions
654 in apatite with homogenization temperatures of 400°C and salinities of 1-2 wt.% NaCl.
655 Kelley and Delaney (1987) concluded that brittle fracture was followed by sealing of the
656 fracture network due to incursion of lower temperature fluids recorded by fluid inclusions
657 in plagioclase, apatite, epidote, and augite with homogenization temperatures of 200-
658 300°C and salinities of 1-7 wt.% NaCl.

659

660 Franzson et al. (2002) described a similar range in fluid inclusion salinities for calcite,
661 quartz and plagioclase at shallower depths (526 to 1338 m) in wells RN-9 and RN-10 in
662 the Reykjanes geothermal system. Fluid inclusion temperatures for calcite at 526 m in
663 RN-10 measured by Franzson (2002) showed a T_h range of ~ 170°C to 250°C, however
664 this study was completed on drill cutting samples, so interpretation of paragenetic
665 relationships of fluid inclusions was not attempted. In the nearby Svartsengi geothermal
666 field, Franzson (1990) concluded from fluid inclusion data that the geothermal system
667 had cooled by 10 to 20°C at 800 m, and by ~50°C at 1,350 m depth. Franzson (1990) also
668 noted that fluid inclusion salinities became more saline with time at Svartsengi. Franzson
669 interpreted the range in fluid inclusion properties from the Reykjanes and Svartsengi

670 geothermal systems as evidence for fluid evolution from fresh to seawater due to
671 meteoric water recharge from a regional icecap present during the last glaciation. Fluid
672 inclusions from the RN-17B core indicate a salinity increase and cooling of $\sim 55^{\circ}\text{C}$ at a
673 vertical depth of 2,560 m, consistent with Franzson's model. An alternative model is that
674 low salinity fluid inclusions reported by Franzson, (2002) are related to phase separation
675 of seawater, an idea supported by the proximity of the fluid inclusion homogenization
676 temperatures to the boiling point-depth curve for seawater (Figure 12). In situations
677 where fluids are near the two-phase boundary, rapid fracture dilation can result in
678 localized isenthalpic phase separation, brine formation, cooling, and supersaturation of
679 mineral phases (Henley and Hughs, 2000). Our very localized fluid inclusion and
680 strontium isotope results do not preclude meteoric recharge in shallower portions of the
681 Reykjanes geothermal system, although are consistent with phase separation of seawater
682 similar to that reported for many submarine hydrothermal systems.

683

684 One explanation for boiling at the depth of the RN-17B core may be a period of elevated
685 temperatures due to injection of a dike or dike swarm. Preliminary examination of the
686 nearby RN-30 core shows evidence of several dikes with chilled margins at a similar
687 vertical depth to the RN-17B core ($\sim 2,200$ m in RN-30 vs 2,560 m in RN-17B). Despite
688 no evidence for meteoric water recharge at the depth of the RN-17B core, our results do
689 not preclude the influence of an icecap on the Reykjanes geothermal system. Boiling at
690 shallow depths may have been influenced by the removal of an icecap, which would
691 possibly result in an initial pressure decrease followed by a temperature decrease once the
692 temperature gradient re-equilibrated with the displaced boiling-depth curve

693 **5.3 Epidote Composition**

694 Epidote zoned with respect to Fe^{3+} - Al^{3+} is commonly formed in hydrothermal systems
695 hosted in a variety of geological settings (Bird et al., 1984; Arnason and Bird, 1992; Bird
696 and Spieler, 2004). Fluid flux, permeability, temperature, oxygen fugacity ($f\text{O}_2$), carbon
697 dioxide fugacity ($f\text{CO}_2$), fluid pH, whole rock composition, and aqueous speciation of
698 Al^{3+} and Fe^{3+} are important variables controlling epidote composition (Arnason and Bird,
699 1993; Bird and Spieler, 2004). Temperature, $f\text{CO}_2$, and $f\text{O}_2$ are the most important
700 intensive variables controlling epidote composition, however elucidation of the key
701 variable is generally not possible (Arnason and Bird, 1993).

702

703 Experimental studies have demonstrated the influence of temperature on the Fe^{3+} content
704 of epidote, and show the relationship is strongly dependent on the buffering mineral
705 assemblage (Liou et al. 1983; Apter and Liou, 1983; Arnason and Bird, 1993). The Fe
706 content of epidote is highly sensitive to small fluctuations in $f\text{CO}_2$ in H_2O -rich fluids with
707 overall low $f\text{CO}_2$, although again, the relationship is strongly dependent on the buffering
708 mineral assemblage (Arnason and Bird 1993). The Fe^{3+} content of epidote appears to be
709 directly related to $f\text{O}_2$; oxidized mineral assemblages contain the most Fe-rich epidote,
710 while reduced assemblages contain the most Fe-poor epidote (Holdaway, 1972; Bird and
711 Helgeson, 1981; Liou et al., 1983; Arnason and Bird, 1993; Bird and Spieler, 2004).

712

713 EMP analyses of epidote from drill cuttings in wells RN-8 and RN-9 in the Reykjanes
714 geothermal system and wells in the nearby Svartsengi geothermal system show no
715 systematic changes in major element concentration with depth or proximity of aquifers,

716 despite documentation of many grains having complex zoning patterns (Lonker et al.,
717 1993; Bird and Spieler, 2004). There are no systematic changes in X_{Fe} of epidote with
718 depth or temperature in RN-17 cuttings (Freedman et al., 2009; Marks et al., 2010a).
719 Epidote in RN-17 drill cuttings from 2,000 to 3,000 m depth show no Al-Fe (III) zoning,
720 and X_{Fe} ranges from 0.17 to 0.36 (Freedman et al., 2009). The results of this study show a
721 comparable range of X_{Fe} in epidote from a single epidote vein at 2799.15 m in the RN-
722 17B core (X_{Fe} of 0.17 to 0.37) compared to drill cuttings from the entire depth of RN-17
723 and other locations in the Reykjanes and Svartsengi geothermal systems. However, the
724 X_{Fe} of epidote grain centers appears to have changed systematically during the infilling of
725 the epidote vein (Figure 6). In contrast to drill cutting results reported by Freedman et al.,
726 (2009) for RN-17 between 2,000 and 3,000 m depth, vein epidote in the RN-17B core
727 commonly has oscillatory zoning with respect to Al-Fe (III), although the zoning within
728 individual crystals shows no systematic pattern.

729

730 The systematic change in X_{Fe} of epidote in the RN-17B core shows little correlation with
731 T_h of fluid inclusions, consistent with results of Marks et al. (2010a) for epidote in drill
732 cuttings compared with depth and temperature. There appears to be a consistent
733 relationship between salinity of fluid inclusions and X_{Fe} of the host epidote (Figure 6).
734 The reported strong dependence of epidote composition on $f\text{O}_2$ suggests that fluid salinity
735 changes were occurring contemporaneously with evolving $f\text{O}_2$ in the geothermal system
736 during vein sealing, rather than salinity being the independent variable, if $f\text{O}_2$ is in fact
737 the controlling variable for the X_{Fe} of epidote in this case. A potential explanation for
738 evolving $f\text{O}_2$ may be partitioning of reduced gasses, primarily H_2S and H_2 , into the vapor

739 phase during boiling that could have left the residual more saline liquid more oxidized.
740 High concentrations of H₂S (26.5% by volume) and H₂ (28.1% by volume) are present in
741 geothermal gas emissions at the Nesjavellir Power Plant, Iceland (Balta et al., 2010),
742 which supports the idea that these gasses may be present in a geothermal system at
743 sufficient concentrations to affect the fO_2 of geothermal fluids at depth. Alternatively, a
744 variation in fluid flux may be the controlling factor for the X_{Fe} of epidote in the RN-17B
745 core, an idea consistent with the mineralization of epidote and subsequent decrease in
746 permeability.

747

748 Sealing of permeability and the related shift to lower fluid-rock ratios may explain the
749 systematic change in X_{Fe} across epidote veins. Lower fluid-rock ratios are consistent with
750 a reduced flux of hydrothermal fluid able to transport dissolved Al from plagioclase in
751 the wall rock (i.e. Alt and Emmerman 1985; Alt, 1995). Mobilization of Al for
752 incorporation into the epidote structure may occur during albitization of calcic
753 plagioclase and/or breakdown of chlorite in the wall rock. Subsequent solutions may have
754 lower Al content possibly due to the fluid being isolated from the wall rock by earlier
755 precipitated Al-rich epidote and/or Al-rich amphibole, thus precipitating more Fe-rich
756 epidote. Alternatively, boiling would increase the pH of fluids and influence the relative
757 proportions of Fe³⁺ and Al³⁺ in solution, although thermodynamic modeling is required to
758 determine if the reduction of Fe³⁺ or Al³⁺ would be larger.

759 **5.4 Amphibole and Plagioclase Major Element Compositions**

760 Despite the pervasive amphibolite grade alteration in the RN-17B core, original textures
761 are preserved, including hyaloclastite glass shards, lithic clasts, and chilled margins of

762 clasts and pillow basalts. Transition from greenschist assemblages to higher-grade
763 amphibolite assemblages is defined in part by replacement of actinolite with hornblende,
764 and the presence of calcic plagioclase. EMP analyses of hydrothermal amphibole reveal
765 that centers of pillow basalt fragments (black rings, Fig. 6) plot closer to the actinolitic
766 end of the pargasite exchange vector. The crystalline cores of pillow basalt fragments
767 retain greenschist grade minerals such as chlorite, and contain sparse, isolated
768 hydrothermal hornblende crystals. The outermost crystalline edges of the pillow
769 fragments (blue rings, Figure 7) are predominantly composed of felted hornblende and
770 plot closer to the aluminous end of the pargasite exchange vector. These results reflect
771 evolution of amphibole composition from actinolitic to pargasitic with increasing
772 water/rock ratios. Protolith type (glass or crystalline) also appears to contribute to the
773 composition of amphiboles. EMP analyses of quenched glassy margins on the pillow
774 basalt (red rings, Figure 8) plot closer to actinolitic compositions than analyses from
775 crystalline parts of the pillow would predict. Amphibole FeO/MgO ratios also suggest
776 protolith composition controls on amphibole composition (Figure 9). Glass protoliths
777 (hyaloclastite shards and quenched clast margins) have higher Mg and lower Fe
778 concentrations than fragments with crystalline protolith, possibly due to enrichment in
779 Mg from earlier palagonitization or chloritization of the glass. The exception to this is
780 hornblende in the very center of crystalline protolith fragments, which spans the entire
781 range of Fe/Mg ratios for crystalline and glass protoliths.

782

783 Hydrothermal amphibole and calcic plagioclase have been described previously in mafic
784 rocks from mid-ocean ridge hydrothermal systems (i.e. Alt et al., 1985; Alt et al., 1986;

785 Vanko and Stakes, 1991; Gillis and Thompson, 1993; Alt, 1995; Alt et al., 1996).
786 Amphibole compositions in seafloor hydrothermal systems commonly change with
787 increasing depth and temperature from actinolite to Mg-hornblende with corresponding
788 increases in Al and Ti contents (Gillis and Thompson, 1993; Alt, 1995). A similar pattern
789 is observed in the results from this study corresponding to the extent of fluid “seen” by
790 the protolith (Figures 8 and 9). Figure 8 indicates an increasing Al content for amphibole
791 in crystalline basalt from the center of a pillow towards the edge of the pillow, while
792 Figure 9 indicates an increase in Mg from the crystalline portion of a basalt pillow
793 towards the glassy margin of the pillow. Overall, EMP results for amphibole indicate that
794 alteration in the RN-17B core has not reached equilibrium in part due to limited
795 permeability. However, timing of alteration events coupled with evolving fluid
796 compositions, and localized chemical controls may also contribute to the systematic
797 variability observed in amphibole compositions.

798 **5.5 Strontium Isotope Ratios in Epidote**

799 Strontium isotope fractionation is not temperature or mineral dependent, and $^{87}\text{Sr}/^{86}\text{Sr}$
800 ratios are significantly different for fresh volcanic rock and seawater (Alt et al., 1996).
801 The low Rb concentration and the young age (<1Ma) of Reykjanes basalts preclude
802 significant radiogenic ^{87}Sr contamination in samples (Marks et al., 2010b). For these
803 reasons $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are particularly useful for evaluating fluid-rock interaction in
804 seawater dominated geothermal systems, and have been used previously to evaluate
805 seawater influence on epidote precipitation in seafloor hydrothermal systems (i.e. Alt et
806 al., 1996) and epidote from drill cuttings in the Reykjanes geothermal system (Marks et
807 al., 2010b).

808

809 $^{87}\text{Sr}/^{86}\text{Sr}$ measurements of epidote from RN-17 and RN-17B drill cuttings, in addition to
810 three powdered samples of RN-17B core whole-rock do not vary with depth or alteration
811 grade, but do vary with increased loss on ignition (LOI) (Marks et al., 2010b). Our
812 results show that lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in vein centers are shifted towards rock values,
813 consistent with the possibility that epidote veins in the RN-17B core record evolution to
814 lower water/rock ratios. Such a relationship between $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and water/rock ratio
815 has been demonstrated previously in whole rock studies of seafloor hydrothermal systems
816 (i.e. Kawahata et al., 1987). Evolving conditions to lower water/rock ratios is consistent
817 with epidote precipitation and a subsequent decrease in water/rock ratio in the fracture
818 network of the RN-17B core due boiling hydrothermal fluids. Conversely, the Sr
819 concentration in epidote is higher in the centers of epidote veins, (based on the increased
820 voltage for ^{88}Sr) as measured by LA-MC-ICP-MS (Figure 11). This observation is in
821 contrast to the relationship observed by Kawahata et al. (1987) for decreasing water/rock
822 ratios in a seafloor hydrothermal system where Sr content remained steady while the
823 $^{87}\text{Sr}/^{86}\text{Sr}$ ratios decreased. Assumedly, lower water/rock ratios would accompany a
824 decrease in Sr content obtained by fluids precipitating epidote due to decreased exposure
825 to the surrounding host rock. However, if the fluids precipitating epidote in vein centers
826 resulted from phase separation and concentration of dissolved solids into a liquid phase
827 due to boiling, a process indicated by our fluid inclusion measurements, fluids in vein
828 centers may have had a higher Sr content than the more dilute fluids forming epidote on
829 the vein edges. Such a partitioning of Sr into a brine or liquid phase has been suggested
830 previously (Bach and Humphries, 1999).

831

832 Salinities lower than seawater in fluid inclusions have previously been interpreted as
833 evidence for evolution from freshwater to seawater fluids in shallow portions of the
834 Reykjanes geothermal system (i.e. Franzson, 1990; Franzson et al., 2002). The
835 implication for the overall seawater $^{87}\text{Sr}/^{86}\text{Sr}$ values in vein epidote is that earlier, less
836 saline hydrothermal fluids had a significant component of seawater. Dilution of 3.2 wt.%
837 NaCl seawater by freshwater to get salinities as low as the ~1 wt.% NaCl we see in the
838 RN-17B core epidote veins would require greater than 3:1 dilution by freshwater and
839 would result in lower overall Sr content, something that is observed (Figure 11).
840 However, this degree of dilution by freshwater at an early stage of epidote vein formation
841 would assumedly shift $^{87}\text{Sr}/^{86}\text{Sr}$ values towards rock values and we see the opposite trend
842 (Figure 11). Therefore, strontium isotope transects across epidote veins suggest that an
843 early meteoric water history for the Reykjanes geothermal system is not recorded by
844 epidote veins in the RN-17B core

845 **6.0 Conclusions**

846 A key difference between the seawater dominated Reykjanes geothermal system and
847 seafloor hydrothermal systems is the lack of cold hydrostatic seawater head.
848 Subsequently, significantly greater depths are required at Reykjanes before temperatures,
849 pressures and mineralization similar to those observed in seafloor hydrothermal systems
850 are observed. The ~2,560-2,570 m vertical depth of the RN-17B core bears many
851 similarities in terms of fluid inclusion characteristics, alteration mineral phases, and
852 alteration mineral compositions as seen in submarine hydrothermal systems. Future
853 recovery of deeper core from the Reykjanes Peninsula may give insights into
854 hydrothermal processes in the deep ocean crust, particularly if depths exceeding the
855 critical point of seawater are achieved.

856
857 This geochemical investigation of samples from drill core RN-17B has revealed evolving
858 geothermal conditions in the Reykjanes geothermal system, evaluated utilizing *in situ*
859 geochemical methods on drill core as opposed to drill cuttings. Fluid inclusion evidence
860 suggests cooling over time of Reykjanes geothermal fluids (below the critical point for
861 seawater) from above the two-phase boundary to below the two-phase boundary. Fluid
862 inclusion salinities indicate that cooling appears to be accompanied by fluids composed
863 of low salinity vapor evolving to more saline residual brine. The Al content of epidote
864 measured by EMP suggests that the cooling of the system is accompanied by sealing of
865 permeability in the geothermal system, supported by strontium isotope ratios measured in
866 epidote vein transects using LA-MC-ICP-MS. Strontium isotope ratios also indicate that
867 hydrothermal fluids had a significant seawater component throughout precipitation of

868 epidote and sealing of permeability in veins. Therefore, incursion of low salinity meteoric
869 fluids is not required to explain fluid inclusion salinities less than seawater at the depth of
870 the RN-17B core. The Al and Mg content of hydrothermal amphibole measured by EMP
871 suggest amphibole compositions are related to protolith type and water-rock ratio, and
872 indicate that alteration mineralogy in the RN-17B core has not reached equilibration in
873 part due to water-rock ratio and in part due to the glassy versus crystalline nature of the
874 protolith.

875

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Figure 1. Location of the Reykjanes geothermal system, Iceland. Wells discussed in the text are indicated (modified after Marks et al. (2010) and Sæmundsson, K. (2011)).

Figure 2. Simplified RN-17B core log. A detailed core log is included as an electronic resource.

Figure 3. BSE images indicating some typical textures described in the text for the RN-17B core. Figure 3A is clast within an altered hyaloclastite matrix (Figure 3B).

Figure 4A through 4C. Transmitted light microphotographs of epidote-hosted fluid inclusions from 2799.15 m. (4A): Primary vapor-rich fluid inclusion adjacent to wall rock. (4B): Primary liquid and vapor fluid inclusions present inwards of the vein wall, shown here present in different crystals. (4C): Liquid dominated fluid inclusions present in the vein center. P(V) = primary vapor inclusion; P(L) primary liquid inclusion; (S)= secondary inclusion.

Figure 5. Plot of T_h and salinity for fluid inclusions hosted by epidote veins from 2799.15 m and 2800.05 m (Left). Error for $T_m = \pm 0.2^\circ\text{C}$, error for $T_h = \pm 5^\circ\text{C}$ (see text for discussion). Spatial distribution of fluid inclusions at 2799.15 m (Right).

Figure 6. Mole fraction of Fe ($X_{\text{Fe}} = \text{Fe}/(\text{Fe}+\text{Al})$) (y-axis) compared to salinity (x-axis) and temperature (legend symbols) in epidote crystals hosting fluid inclusions. Al is more concentrated in epidote that hosts lower salinity, higher temperature fluid inclusions, which are located adjacent to vein walls.

Figures 7A through 7D. BSE images reveal various amphibole morphologies present in the RN-17B core. Red, blue, and black circles indicate locations of EMP analyses presented in Figure 7. Abbreviations: Ab = albite An = anorthite, Ap = apatite, Chl = chlorite, Hbl = hornblende, Mag = magnetite, Ttn = titanite

Figure 8. Spatially resolved amphibole compositions are consistent with those presented by Marks et al. (2011), however variations in amphibole composition in the RN-17B core relate to protolith permeability and composition. Microprobe totals recalculated using the method of Dale et al. (2005).

Figure 9. Fe and Mg concentrations of amphiboles vary according to protolith type with glass protoliths showing higher MgO contents.

Figures 10A through 10D: BSE images and EMP analyses reveal that hydrothermal plagioclase is calcic (Figures 9A and 9C), while altered igneous plagioclase and hyaloclastite matrix consists of intergrown calcic and sodic plagioclase (Figures 9B and 9D). Abbreviations as in Figure 7; FeS₂ = pyrite.

Figure 11. Two LA-ICP-MS transects of an epidote vein at 2800.05 m. The ^{88}Sr signal (Sr abundance) increases in the vein center, and the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio decreases in the vein center.

Figure 12. Reykjanes geothermal system fluid inclusion temperatures in relation to the boiling-depth curve for seawater. Modified after Franzson et al. (2002) and Marks et al. (2010).

Table 1. Epidote fluid inclusion measurements

Table 2. Electron microprobe analyses of epidote reported as oxides (all Fe as Fe₂O₃; based on 25 oxygens per formula unit) [Upper]. ⁸⁷Sr/⁸⁶Sr determined using LA-MC-ICP-MS. Electron microprobe analyses of epidote reported as number of ions [Lower].

Table 3. Amphibole EMP analyses reported as oxides (all Fe as FeO; based on 23 oxygen formula) [Upper]. Amphibole EMP analyses reported as number of ions (FeO recalculated as Fe²⁺ and Fe³⁺ using the method of Dale et al. (2005) [Lower].

Table 4. Plagioclase EMP analyses reported as weight % oxides [Upper]. Plagioclase EMP analyses reported as number of ions [Lower].

Table 1

Epidote Crystal/Inclusion Number	Sample Depth	T_h (°C)	T_m (°C)	Salinity (Wt. % NaCl)
1/1	2799.15	401.5	--	--
2/1	2799.15	400.4	--	--
3/1	2799.15	>480	--	--
4/1	2799.15	378.3	--	--
4/2	2799.15	--	-4.8	7.6
4/3	2799.15	--	-4.8	7.6
4/4	2799.15	--	-4.8	7.6
4/Average		378.3	-4.8	7.6
5/1	2799.15	386.3	-1.1	1.9
6/1	2799.15	366.4	-3.3	5.4
6/2	2799.15	372	-3.2	5.3
6/Average		369.2	-3.3	5.3
7/1	2799.15	390	-1.0	1.7
7/2	2799.15	--	-1.1	1.9
7/Average		390	-1.1	1.8
8/1	2799.15	373.2	-2.8	4.6
8/2	2799.15	370	-3.3	5.4
8/Average		371.6	-3.1	5.0
9/1	2799.15	372.9	-3.8	6.2
10/1	2799.15	385.9	-1.7	2.9
10/2	2799.15	385.8	-1.8	3.1
10/Average		385.9	-1.8	3.0
11/1	2799.15	389.5	-0.6	1.1
12/1	2799.15	400.6	-2.4	4.0
12/2	2799.15	386.6	-2.5	4.2
12/3	2799.15	401.6	-1.7	2.9
12/Average		396.3	-2.2	3.7
13/1	2799.15	--	-0.9	1.6
14/1	2799.15	>480	-0.3	0.5
15/1	2799.15	400.1	--	--
15/2	2799.15	--	-0.6	1.1
15/Average		400.1	-0.6	1.1
16/1	2799.15	370.4	-3.3	5.4
16/3	2799.15	--	-3.4	5.6
16/4	2799.15	370.6	-2.9	4.8
16/5	2799.15	369.7	--	--
16/6	2799.15	360.2	--	--
16/Average		367.7	-3.2	5.3

17/1	2799.15	>480	-1.2	2.1
17/2	2799.15	>480	-1.4	2.4
17/3	2799.15	>480	-0.6	1.1
17/4	2799.15	>480	--	--
17/5	2799.15	>480	-1.5	2.6
17/Average		>480	-1.2	2.0
18/1	2799.15	>480	-1.5	2.6
18/2	2799.15	>480	-1.2	2.1
18/3	2799.15	>480	-0.8	1.4
18/Average		>480	-1.2	2.0
19/1	2799.15	>480	-1.4	2.4
19/2	2799.15	>480	--	--
19/Average		>480	-1.4	2.4
20/1	2799.15	>480	-1.5	2.6
20/2	2799.15	>480	--	--
20/Average		>480	-1.5	2.6
21/1	2800.05	344.9	-2.1	3.5
21/2	2800.05	358.1	-2.2	3.7
21/3	2800.05	354.2	-2.0	3.4
21/Average		352.4	-2.1	3.5
22/1	2800.05	>480	-1.0	1.7
23/1	2800.05	>480	-0.9	1.6
24/1	2800.05	>480	-2.1	3.5
24/2	2800.05	>480	-2.0	3.4
24/Average		>480	-2.1	3.5
25/1	2800.05	>480	-1.0	1.7
26/1	2800.05	>480	-0.8	1.4
26/2	2800.05	>480	--	--
26/3	2800.05	>480	-0.3	0.5
26/4	2800.05	>480	-1.0	1.7
26/5	2800.05	>480	-0.9	1.6
26/Average		>480	-0.8	1.3
27/1	2799.15	354.3	-2.8	4.6
28/1	2799.15	352.2	-2.8	4.6

-- Not measured due to small size, darkening, or decrepitation
T_m Ice melting temperature (°C)
T_h Homogenization temperature (°C)

Table 2

Epidote Crystal	Depth (m)	MgO	CaO	MnO	SrO	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂	TiO ₂	Total	⁸⁷ Sr/ ⁸⁶ Sr
Weight % Oxide											
2	2799.15	0.03	23.12	0.19	0.02	22.96	14.32	37.78	0.32	100.64	0.70437
3	2799.15	0.02	23.27	0.14	0.03	23.89	12.91	37.61	0.06	99.82	0.70449
5	2799.15	0.11	23.53	0.09	0.05	26.27	9.61	37.96	0.02	99.52	0.70457
7	2799.15	0.04	22.94	0.15	0.08	25.99	10.81	37.79	0.09	99.79	0.70475
8	2799.15	0.03	23.05	0.09	0.06	23.04	14.36	37.50	0.08	100.08	0.70511
9	2799.15	0.01	23.29	0.17	0.10	23.10	14.25	37.26	0.19	100.25	0.70568
10	2799.15	0.01	23.76	0.09	0.15	27.90	8.66	38.21	0.04	100.74	0.70456
11	2799.15	0.09	23.67	0.18	0.05	27.41	8.49	38.40	0.14	100.35	0.70449
12	2799.15	0.03	23.33	0.25	0.09	26.17	10.62	38.00	0.01	100.42	0.70464
13	2799.15	0.09	24.10	0.12	0.05	27.16	9.81	38.03	0.00	101.29	0.70441
14	2799.15	--	--	--	--	--	--	--	--		0.70453
15	2799.15	0.04	23.38	0.49	0.08	26.00	10.61	38.27	0.03	100.81	0.70418
16	2799.15	0.02	22.90	0.03	0.04	20.13	18.72	37.30	0.01	101.01	0.70438
17	2799.15	0.05	23.19	0.67	0.14	26.33	10.15	38.27	0.12	100.84	0.70428
18	2799.15	0.03	22.97	0.31	0.14	25.89	11.09	37.63	0.13	100.09	0.70469
19	2799.15	0.04	23.78	0.12	0.06	26.17	10.41	38.22	0.10	100.82	0.70468
20	2799.15	--	--	--	--	--	--	--	--		0.70471
21	2800.05	0.07	23.48	0.05	0.04	25.47	13.02	37.46	0.11	101.61	0.70474
23	2800.05	--	--	--	--	--	--	--	--		0.70457
24	2800.05	--	--	--	--	--	--	--	--		0.70466
25	2800.05	--	--	--	--	--	--	--	--		0.70490
26	2800.05	--	--	--	--	--	--	--	--		0.70478
27	2799.15	0.05	23.57	0.11	0.05	24.66	12.56	37.58	0.18	100.67	0.70465
28	2799.15	0.06	23.31	0.56	0.07	25.64	10.68	38.08	0.17	100.48	0.70421

Epidote Crystal	Depth (m)	Mg	Ca	Mn	Sr	Al	Fe	Si	Ti	Total	X _{Fe}
Number of Ions											
2	2799.15	0.01	3.93	0.03	0.002	4.29	1.71	5.99	0.04	17.98	0.28
3	2799.15	0.01	3.97	0.02	0.003	4.48	1.55	5.98	0.01	18.00	0.26
5	2799.15	0.03	3.97	0.01	0.005	4.88	1.14	5.98	0.00	18.01	0.19
7	2799.15	0.01	3.87	0.02	0.007	4.83	1.28	5.96	0.01	17.98	0.21
8	2799.15	0.01	3.93	0.01	0.005	4.33	1.72	5.98	0.01	17.99	0.28
9	2799.15	0.00	3.98	0.02	0.010	4.34	1.71	5.94	0.02	18.02	0.28
10	2799.15	0.00	3.95	0.01	0.013	5.10	1.01	5.93	0.01	18.02	0.17
11	2799.15	0.02	3.95	0.02	0.005	5.03	0.99	5.97	0.02	18.00	0.17
12	2799.15	0.01	3.92	0.03	0.008	4.83	1.25	5.95	0.00	18.00	0.21
13	2799.15	0.02	4.00	0.02	0.005	4.96	1.15	5.90	0.00	18.05	0.19
15	2799.15	0.01	3.91	0.07	0.007	4.79	1.25	5.98	0.00	18.00	0.21
16	2799.15	0.00	3.93	0.00	0.003	3.80	2.26	5.98	0.00	17.99	0.37
17	2799.15	0.01	3.88	0.09	0.013	4.84	1.19	5.97	0.01	18.00	0.20
18	2799.15	0.01	3.88	0.04	0.013	4.81	1.31	5.93	0.02	18.00	0.21
19	2799.15	0.01	3.97	0.02	0.006	4.81	1.22	5.96	0.01	18.01	0.20
21	2800.05	0.02	3.93	0.01	0.004	4.69	1.53	5.85	0.01	18.03	0.25
27	2799.15	0.01	3.98	0.01	0.004	4.58	1.49	5.92	0.02	18.02	0.25
28	2799.15	0.01	3.92	0.08	0.007	4.74	1.26	5.97	0.02	18.01	0.21

Analyses calculated on the basis of 25 oxygen structural formula
OH fixed at an ion number of 2.00
⁸⁷Sr/⁸⁶Sr Strontium isotope ratio determined by LA-ICP-MS
X_{Fe} = Fe/(Fe+Al)
-- Not analyzed
All Fe calculated as Fe₂O₃

Table 3

All Fe is FeO based on a 23 oxygen formula

Amphibole Sample	Depth (m)	# of Analyses	SiO ₂	Al ₂ O ₃	TiO ₂	MnO	MgO	CaO	Na ₂ O	K ₂ O	FeO	Total
Weight % Oxide												
QE 1	2800.05	2	45.61	8.26	0.76	0.37	12.13	12.18	1.04	0.11	16.09	98.53
QE 2	2800.05	3	46.41	8.55	0.53	0.33	13.04	12.38	1.01	0.09	14.63	98.97
QE 3	2800.05	2	47.05	8.12	0.64	0.38	13.42	12.11	1.07	0.11	14.24	99.16
QE 4	2800.05	2	47.01	7.06	0.46	0.43	12.97	12.49	0.82	0.08	15.42	98.74
QE 5	2800.05	2	47.86	6.69	0.41	0.44	12.74	12.32	0.77	0.09	15.02	98.35
OC 1	2800.05	2	45.96	8.40	1.08	0.37	12.38	12.13	0.98	0.10	15.70	99.11
OC 2	2800.05	3	46.29	8.59	0.64	0.40	12.99	11.94	1.00	0.09	14.86	98.83
OC 3	2800.05	2	42.45	12.34	0.79	0.35	11.44	12.18	1.69	0.08	15.28	98.57
OC 4	2800.05	3	45.08	9.28	0.59	0.30	11.97	12.18	1.19	0.10	16.26	98.95
OC 5	2800.05	3	42.94	11.37	1.12	0.32	11.60	12.25	1.68	0.11	15.92	99.31
OC 6	2800.05	2	42.86	11.77	1.11	0.32	11.93	12.19	1.69	0.05	14.75	98.67
IC 1	2800.05	3	47.40	5.60	0.21	0.44	10.76	12.01	0.67	0.05	19.30	98.41
IC 2	2800.05	3	48.41	6.59	0.58	0.37	13.46	12.51	0.75	0.09	14.66	99.44
IC 3	2800.05	1	47.74	7.88	0.66	0.46	14.61	12.31	0.89	0.06	12.75	99.41
IC 4	2800.05	3	46.90	7.71	0.55	0.36	12.57	11.95	1.01	0.11	15.98	99.14
HS 1	2800.05	2	47.38	7.03	0.44	0.32	13.10	12.35	0.97	0.07	15.20	98.87
HS 2	2800.05	2	47.39	7.69	0.69	0.38	13.18	12.43	0.98	0.07	14.10	98.94
HS 3	2800.05	2	48.93	6.06	0.35	0.29	13.27	12.39	0.71	0.05	14.95	99.02
HS 4	2800.05	3	48.49	6.05	0.36	0.36	13.43	12.36	0.78	0.06	14.81	98.73
HS 5	2798.64	1	45.54	9.18	0.85	0.41	13.08	11.98	1.40	0.05	14.23	98.73
HS 6	2798.64	3	48.30	6.36	0.37	0.26	13.65	12.48	0.88	0.10	14.16	98.57
HS 7	2798.64	3	44.43	10.48	0.93	0.33	12.94	12.25	1.68	0.08	13.92	99.04
HS 8	2798.64	3	44.80	10.40	1.01	0.36	12.73	12.33	1.61	0.06	14.14	99.45
VS 1	2798.64	3	47.57	7.84	0.15	0.47	13.18	12.27	1.04	0.05	14.24	98.82
VS 2	2798.64	3	46.10	8.88	0.14	0.34	13.17	12.19	1.30	0.05	13.98	98.14
VS 3	2798.64	2	45.84	9.59	0.61	0.30	13.47	12.42	1.47	0.08	13.47	99.28
VS 4	2798.64	2	44.29	10.22	0.52	0.36	11.97	12.11	1.47	0.11	15.44	98.47
VS 5	2798.64	3	48.73	6.57	0.36	0.36	14.02	12.33	0.99	0.08	13.49	98.96
VS 6	2798.64	2	48.13	6.52	0.19	0.52	13.40	12.28	1.10	0.07	14.69	98.92
VS 7	2798.64	3	45.71	9.73	0.40	0.32	13.65	12.16	1.62	0.09	13.18	98.88
VS 8	2798.64	3	46.04	8.86	0.49	0.31	12.78	12.30	1.25	0.12	14.80	98.98
VS 9	2798.64	2	45.38	9.46	0.66	0.33	12.33	12.20	1.38	0.09	15.56	99.41
VE 1	2800.35	3	45.31	10.73	0.58	0.33	12.67	12.20	1.22	0.04	14.31	99.42
VE 2	2800.35	2	44.18	10.76	0.64	0.38	12.07	12.21	1.45	0.03	15.16	98.89
VE 3	2800.35	1	46.81	7.12	0.84	0.40	12.54	12.58	0.83	0.06	15.31	98.48
VE 4	2800.35	1	48.71	6.23	0.37	0.42	13.51	12.34	0.65	0.04	14.90	99.20
VE 5	2800.35	2	46.74	7.97	0.50	0.36	12.60	12.32	0.81	0.07	14.87	98.23
VE 6	2800.35	1	47.30	6.51	0.95	0.35	13.12	12.28	0.71	0.07	14.73	98.03
VE 7	2800.35	2	49.71	5.36	0.25	0.44	13.81	11.92	0.61	0.05	14.08	98.25

QE = Quenched Edge (Clast edge)
OC = Outer Crystalline (inwards of quenched edge)
IC = Inner crystalline (clast center)
HS = Hyaloclastite Shard
VS = Acicular green amphibole in vesicles
VE = Acicular green amphibole in vein selvages
FeO recalculated as Fe²⁺ and Fe³⁺ using method of Dale et al., 2005.

Amphibole Sample	Si	Al	Ti	Mn	Mg	Ca	Na	K	Fe2+	Fe3+	Fe Total	Sum	Fe ²⁺ / (Fe ²⁺ + Mg)
Number of Ions													
QE 1	6.78	1.45	0.08	0.05	2.72	1.96	0.30	0.02	1.49	0.51	2.00	15.37	0.35
QE 2	6.82	1.48	0.06	0.04	2.89	1.97	0.29	0.02	1.31	0.48	1.80	15.35	0.31
QE 3	6.88	1.40	0.07	0.05	2.96	1.92	0.31	0.02	1.32	0.43	1.74	15.34	0.31
QE 4	6.94	1.23	0.05	0.05	2.89	2.00	0.24	0.02	1.36	0.55	1.90	15.31	0.32
QE 5	7.09	1.17	0.05	0.06	2.82	1.96	0.22	0.02	0.31	1.55	1.86	15.24	0.10
OC 1	6.78	1.46	0.12	0.05	2.75	1.94	0.29	0.02	1.45	0.48	1.94	15.34	0.35
OC 2	6.80	1.49	0.07	0.05	2.88	1.90	0.29	0.02	1.29	0.54	1.83	15.32	0.31
OC 3	6.32	2.16	0.09	0.04	2.59	1.96	0.49	0.02	1.36	0.54	1.90	15.57	0.34
OC 4	6.67	1.62	0.07	0.04	2.67	1.96	0.35	0.02	1.44	0.57	2.01	15.41	0.35
OC 5	6.36	1.98	0.12	0.04	2.59	1.97	0.49	0.02	1.42	0.56	1.97	15.56	0.35
OC 6	6.36	2.06	0.12	0.04	2.67	1.96	0.49	0.01	1.33	0.50	1.83	15.54	0.33
IC 1	7.14	0.99	0.02	0.06	2.44	1.96	0.20	0.01	1.95	0.48	2.43	15.25	0.44
IC 2	7.07	1.13	0.06	0.05	2.95	1.97	0.21	0.02	1.43	0.36	1.79	15.27	0.33
IC 3	6.91	1.34	0.07	0.06	3.19	1.93	0.25	0.01	1.07	0.48	1.54	15.30	0.25
IC 4	6.91	1.34	0.06	0.04	2.79	1.91	0.29	0.02	1.47	0.49	1.97	15.32	0.35
HS 1	6.98	1.22	0.05	0.04	2.91	1.97	0.28	0.01	1.44	0.43	1.87	15.34	0.33
HS 2	6.97	1.33	0.08	0.05	2.90	1.97	0.28	0.01	1.44	0.29	1.73	15.32	0.33
HS 3	7.18	1.05	0.04	0.04	2.92	1.96	0.20	0.01	1.52	0.32	1.84	15.24	0.34
HS 4	7.14	1.05	0.04	0.05	2.97	1.97	0.22	0.01	1.46	0.36	1.82	15.27	0.33
HS 5	6.71	1.59	0.09	0.05	2.90	1.91	0.40	0.01	1.31	0.44	1.75	15.43	0.31
HS 6	7.11	1.10	0.04	0.03	3.02	1.98	0.25	0.02	1.42	0.32	1.74	15.30	0.32
HS 7	6.54	1.82	0.10	0.04	2.87	1.95	0.48	0.01	1.28	0.43	1.71	15.53	0.31
HS 8	6.58	1.80	0.11	0.04	2.81	1.96	0.46	0.01	1.36	0.37	1.74	15.50	0.33
VS 1	6.99	1.36	0.02	0.06	2.91	1.95	0.30	0.01	1.42	0.33	1.75	15.34	0.33
VS 2	6.82	1.55	0.02	0.04	2.93	1.95	0.38	0.01	1.29	0.43	1.73	15.42	0.31
VS 3	6.70	1.65	0.07	0.04	2.96	1.96	0.42	0.02	1.26	0.39	1.65	15.47	0.30
VS 4	6.59	1.79	0.06	0.05	2.68	1.95	0.43	0.02	1.42	0.50	1.92	15.49	0.35
VS 5	7.13	1.13	0.04	0.04	3.07	1.94	0.28	0.01	1.39	0.26	1.65	15.31	0.31
VS 6	7.09	1.13	0.02	0.07	2.96	1.95	0.32	0.01	1.50	0.31	1.81	15.36	0.34
VS 7	6.70	1.68	0.04	0.04	3.01	1.93	0.46	0.02	1.21	0.41	1.62	15.50	0.29
VS 8	6.78	1.54	0.05	0.04	2.83	1.96	0.36	0.02	1.40	0.43	1.82	15.42	0.33
VS 9	6.68	1.64	0.07	0.04	2.73	1.95	0.40	0.02	1.44	0.48	1.92	15.45	0.34
VE 1	6.61	1.85	0.06	0.04	2.79	1.93	0.35	0.01	1.24	0.51	1.75	15.38	0.31
VE 2	6.53	1.87	0.07	0.05	2.69	1.96	0.42	0.01	1.35	0.52	1.87	15.47	0.33
VE 3	6.94	1.24	0.09	0.05	2.80	2.02	0.24	0.01	1.46	0.44	1.90	15.30	0.34
VE 4	7.12	1.07	0.04	0.05	2.97	1.95	0.19	0.01	1.40	0.42	1.82	15.23	0.32
VE 5	6.93	1.39	0.06	0.05	2.81	1.97	0.23	0.01	1.44	0.40	1.84	15.29	0.34
VE 6	7.02	1.14	0.11	0.04	2.93	1.97	0.21	0.01	1.44	0.39	1.83	15.26	0.33
VE 7	7.31	0.93	0.03	0.06	3.05	1.89	0.17	0.01	1.45	0.28	1.73	15.18	0.32

QE = Quenched Edge (Clast edge)
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HS = Hyaloclastite Shard
VS = Acicular green amphibole in vesicles
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Table 4

Plagioclase Sample	Depth (m)	Na ₂ O	K ₂ O	CaO	Fe ₂ O ₃	Al ₂ O ₃	SiO ₂	Total
Weight % Oxide								
B-plag 1	2799.90	0.21	0.00	20.04	0.34	36.01	42.88	99.48
B-plag 2	2799.90	0.17	0.01	20.36	1.25	34.67	41.91	98.37
C-plag1	2800.35	0.96	0.03	17.82	1.00	34.12	43.30	97.22
C-plag2	2800.35	1.47	0.02	17.03	0.41	33.45	44.73	97.10
C-plag3	2800.35	1.52	0.04	16.93	0.78	33.40	44.92	97.58
F1-plag1	2804.30	2.07	0.05	16.65	0.28	33.47	47.11	99.64
F1-plag2	2804.30	9.12	0.07	4.80	0.23	23.62	62.71	100.54
G2-plag1	2804.50	0.21	0.01	19.39	0.84	36.11	42.54	99.11
G2-plag2	2804.50	0.33	0.02	19.54	0.51	35.69	42.78	98.86
G2-plag3	2804.50	0.23	0.00	19.87	0.45	36.38	43.15	100.08
H1-plag1	2804.80	0.09	0.01	20.21	0.48	36.93	42.48	100.19
H1-plag2	2804.80	1.17	0.04	17.40	0.85	33.80	44.45	97.70
H1-plag3	2804.80	0.33	0.02	19.62	0.60	35.76	43.09	99.42
H1-plag4	2804.80	1.45	0.01	17.49	1.20	32.78	46.06	98.99
H2-plag1	2804.80	1.66	0.00	17.62	0.51	34.07	45.75	99.60
H2-plag2	2804.80	1.26	0.02	18.27	0.46	34.85	45.34	100.20
H2-plag3	2804.80	8.47	0.07	5.52	0.08	24.33	61.27	99.74
H2-plag4	2804.80	1.12	0.02	18.08	0.25	34.90	45.29	99.66

Plagioclase Sample	Depth (m)	Na	K	Ca	Fe	Al	Si	Total	An Content
Number of Ions									
B-plag 1	2799.90	0.02	0.00	1.00	0.01	1.98	2.00	5.01	98
B-plag 2	2799.90	0.02	0.00	1.04	0.05	1.94	1.99	5.03	98
C-plag1	2800.35	0.09	0.00	0.91	0.04	1.91	2.06	5.01	91
C-plag2	2800.35	0.14	0.00	0.87	0.01	1.87	2.12	5.01	87
C-plag3	2800.35	0.14	0.00	0.86	0.03	1.86	2.12	5.01	86
F1-plag1	2804.30	0.19	0.00	0.82	0.01	1.82	2.17	5.01	82
F1-plag2	2804.30	0.78	0.00	0.23	0.01	1.23	2.77	5.01	23
G2-plag1	2804.50	0.02	0.00	0.97	0.03	1.99	1.99	5.01	98
G2-plag2	2804.50	0.03	0.00	0.98	0.02	1.97	2.01	5.01	97
G2-plag3	2804.50	0.02	0.00	0.99	0.02	1.99	2.00	5.01	98
H1-plag1	2804.80	0.01	0.00	1.00	0.02	2.02	1.97	5.02	99
H1-plag2	2804.80	0.11	0.00	0.88	0.03	1.88	2.10	5.00	89
H1-plag3	2804.80	0.03	0.00	0.98	0.02	1.97	2.01	5.01	97
H1-plag4	2804.80	0.13	0.00	0.87	0.04	1.80	2.15	5.00	87
H2-plag1	2804.80	0.15	0.00	0.87	0.02	1.86	2.12	5.02	85
H2-plag2	2804.80	0.11	0.00	0.90	0.02	1.89	2.09	5.01	89
H2-plag3	2804.80	0.73	0.00	0.26	0.00	1.28	2.73	5.00	26
H2-plag4	2804.80	0.10	0.00	0.90	0.01	1.90	2.09	5.00	90

An Content = Ca/(Ca+Na)*100

Figure 1

- Lava from 1226 AD
- 2,000 - 3,000 year-old lava flows
- 3,000-8,000 year-old lava flows
- 8,000-14,500 year-old lava flows
- Pleistocene hyaloclasite ridges
- Drillhole
- Map trace of drillhole
- Extent of surface alteration
- Faults and fractures
- Craters

Figure 2

Hyaloclastite Breccia

Volcanic Sand

Crystalline Basalt

Lithic Breccia

O1

Olivine-phyric Basalt or
Basaltic Clasts

Pl

Plagioclase-phyric Basalt
or Basaltic Clasts

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

- DWC Standard
- ▲ Krafla Plagioclase
- Epidote Vein Transect 1 (2 SE Error Bars)
- - -●- - - Epidote Vein Transect 2 (2 SE Error Bars)
- Seawater (0.7092)
- - - Reykjanes Geothermal Fluids (0.70412) (Millot et al., 2009)
- Unaltered Basalt (0.7030 - 0.7034)
- 88Sr Concentration Transect 1 (Volts)
- - - 88Sr Concentration Transect 2 (Volts)

Figure 12