**Origin of Magnetic Highs at Ultramafic Hosted Hydrothermal Systems: Insights
from the Yokoniwa Site of Central Indian Ridge**

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17 **Abstract**

18 High-resolution vector magnetic measurements were performed on an inactive
19 ultramafic-hosted hydrothermal vent field, called Yokoniwa Hydrothermal Field (YHF),
20 using a deep-sea manned submersible *Shinkai6500* and an autonomous underwater
21 vehicle *r2D4*. The YHF has developed at a non-transform offset massif of the Central
22 Indian Ridge. Dead chimneys were widely observed around the YHF along with a very
23 weak venting of low-temperature fluids so that hydrothermal activity of the YHF was
24 almost finished. The distribution of crustal magnetization from the magnetic anomaly
25 revealed that the YHF is associated with enhanced magnetization, as seen at the
26 ultramafic-hosted Rainbow and Ashadze-1 hydrothermal sites of the Mid-Atlantic Ridge.
27 The results of rock magnetic analysis on seafloor rock samples (including basalt, dolerite,
28 gabbro, serpentinized peridotite, and hydrothermal sulfide) showed that only highly
29 serpentinized peridotite carries high magnetic susceptibility and that the natural remanent
30 magnetization intensity can explain the high magnetization of Yokoniwa. These
31 observations reflect abundant and strongly magnetized magnetite grains within the highly
32 serpentinized peridotite. Comparisons with the Rainbow and Ashadze-1 suggest that in

ultramafic-hosted hydrothermal systems, strongly magnetized magnetite and pyrrhotite form during the progression of hydrothermal alteration of peridotite. After the completion of serpentinization and production of hydrogen, pyrrhotites convert into pyrite or nonmagnetic iron sulfides, which considerably reduces their levels of magnetization. Our results revealed origins of the magnetic high and the development of subsurface chemical processes in ultramafic-hosted hydrothermal systems. Furthermore, the results highlight the use of near-seafloor magnetic field measurements as a powerful tool for detecting and characterizing seafloor hydrothermal systems.

1. Introduction

Hydrothermal activity within oceanic lithosphere is a fundamental process in mid-ocean ridges and arc-backarc systems, and it contributes considerably to the cooling of the oceanic lithosphere and to the geochemical cycles in the oceans (e.g., *Elderfield and Schultz, 1996; Hannington et al., 2011*). The location and spatial extent of hydrothermal activity are difficult to constrain; however, studies of near-seafloor magnetic field can highlight these features because crustal magnetic minerals can be

destroyed or created by hydrothermal processes. Reduced magnetization associated with hydrothermal deposits was widely observed in both the active and inactive lava-hosted hydrothermal fields in varied tectonic settings, including: the hydrothermal site TAG (Trans-Atlantic Geotraverse) and the 4°48'S fields of the Mid-Atlantic Ridge (MAR; *Tivey and Dymant.*, 2010; *Tivey et al.*, 1993, 1996, 2003), the Main Endeavour Field and Raven Field of the Juan de Fuca Ridge (JFR; *Tivey and Johnson*, 2002; *Tivey et al.*, 2014), a site on the Southwest Indian Ridge (*Zhu et al.*, 2010), back-arc-type sites of the Southern Mariana Trough (*Fujii et al.*, 2015; *Nakamura et al.*, 2013), the Hakurei Field within a caldera of the Izu-Ogasawara arc-back-arc volcano (*Honsho et al.*, 2013), and the Brothers Field within a caldera of the Kermadec intraoceanic arc volcano (*Caratori-Tontini et al.*, 2012).

Weakened magnetic anomalies are interpreted as locally altered up-flow zones formed by hydrothermal activity, because hydrothermal alteration processes destroy iron-titanium oxide minerals (titanomagnetite), which carry magnetic remanence in oceanic extrusive (*Ade-Hall et al.*, 1971; *Pariso and Johnson*, 1991; *Rona*, 1978; *Wooldridge et al.*, 1990). Similar magnetic lows were confirmed in the ancient analogs of

seafloor hydrothermal systems found in ophiolites, where zones of demagnetized crust are associated with the mineralized stockwork of ore bodies (*Hall, 1992; Richards et al., 1989; Walls and Hall, 1998*). Furthermore, studies on volcanic geothermal areas in New Zealand showed that hydrothermal alteration of magnetic minerals is the most important mechanism for creating zones of weak magnetization rather than by thermal demagnetization processes (*Hochstein and Soengkono, 1997*). In the seafloor realm, weak magnetization zones were observed in both active and extinct vent areas, also suggesting that differences in the thermal environment do not significantly affect magnetic anomalies (*Fujii et al., 2015; Tivey and Johnson, 2002*). Another explanation for zones of weak magnetic response is the accumulation of thick hydrothermal deposits, which may result in an apparent magnetic low due to the increased distance between the measurement point and the underlying magnetized basalt (*Szitar et al., 2014a*).

In addition to magnetic lows in lava-hosted systems, it was shown that hydrothermal activity has also lead to enhanced crustal magnetization in the Bent Hill massive sulfide (BHMS) deposits, which are located on a sediment-covered axial valley of the JFR (*Gee et al., 2001; Tivey, 1994*), and in ultramafic-hosted active hydrothermal

fields of the MAR (*Tivey and Dymment, 2010; Szitkar et al., 2014b*). Enhanced magnetic anomalies are interpreted as magnetized bodies with large quantities of magnetic minerals (mostly magnetite and pyrrhotite) created by high-temperature hydrothermal precipitation (*Gee et al., 2001; Szitkar et al., 2014b; Tivey, 1994; Tivey and Dymment, 2010*). However, despite high-resolution studies on the magnetization of hydrothermal fields as mentioned above, magnetic mapping studies combined with studies of the magnetic properties of host rocks are limited; therefore, the mineralogical, geochemical and rock magnetic characteristics of magnetic anomalies remain unclear.

Previous studies of near-seafloor magnetics reported high magnetization zones located at active ultramafic-hosted hydrothermal fields where high-temperature venting occurs (*Szitkar et al. 2014b; Tivey and Dymment, 2010*). These magnetic highs are considered to reflect the presence of mineralized stockwork, in which several chemical processes (e.g., serpentinization and sulfide mineral deposition) create and preserve strongly magnetized magnetite (*Szitkar et al., 2014b*). At the Rainbow hydrothermal site of the MAR, the magnetization from magnetic anomaly mapping estimates up to ~30 A/m; however the magnetization of sulfide-impregnated serpentinized peridotite

97 samples collected in the same region are too weak to explain the intensity of the magnetic
98 anomaly (*Szitkar et al.* 2014b).

99 Exposures of ultramafic rocks are extensively distributed within slow spreading
100 environments, where alteration processes significantly influence submarine ecosystems
101 and result in high concentrations of metals (e.g. *Fouquet et al.*, 2010; *Kelley et al.*, 2005;
102 *Nakamura and Takai*, 2014). Therefore, investigating magnetic signatures in these
103 ultramafic-hosted hydrothermal systems is important for detecting active and inactive
104 hydrothermal sites and their mineralization states.

105 In this study, we explored the ultramafic-hosted Yokoniwa Hydrothermal Field
106 (YHF), located on the Central Indian Ridge (CIR). The YHF is primarily an extinct
107 hydrothermal system and represents a unique target for investigating the differences in
108 magnetic signatures between active and inactive hydrothermal systems hosted within
109 ultramafic rocks. We investigated the magnetization distribution from magnetic
110 anomalies obtained by an autonomous underwater vehicle (AUV) and a human occupied
111 vehicle (HOV) observations, and analyzed magnetic properties of rock samples collected
112 around the YHF. On the basis of these multiple analyses, this study provides insight into:

(i) the magnetic signature of ultramafic-hosted inactive hydrothermal systems; (ii) the role of serpentinization in contributing to magnetic anomalies, and; (iii) the formation history of magnetic minerals during hydrothermal activity in ultramafic rocks.

2. Yokoniwa Hydrothermal Field

The YHF (Fig. 1) is located in the southernmost part of the CIR near the Rodriguez triple junction. The tectonic development of the area has been reported by past studies (*Briais, 1995; Honsho et al., 1996; Mendel et al., 2000; Okino et al., 2015; Sato et al., 2009*). The CIR is a slow–intermediate-rate spreading ridge system with a well-developed axial valley, which is segmented by many fracture zones and non-transform discontinuities (NTDs). The full spreading rate in the southernmost segment, CIR-segment1 (CIR-S1), is approximately 47 mm/yr (MORVEL: *DeMets et al., 2010*). The CIR-S1 is a 20-km segment with a deep axial valley, where undeformed volcanic cones and flat-topped volcanic knolls are distributed within a neo-volcanic zone. Segment length shortens with age in off-axis area, implying that the CIR-S1 was newly created during the evolution of the triple junction (*Honsho et al., 1996; Mendel et al.,*

129 2000).

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131

132 Figure 1. Bathymetry and magnetization in the Yokoniwa Rise region
133 Bathymetry (m; A) and magnetization (A/m; B) around the Yokoniwa Rise including the
134 Yokoniwa Hydrothermal Field (YHF; *Okino et al.*, 2015). Rock samples collected by
135 submersible (circles) and dredge (squares) are shown as basalt (red), dolerite (brown),
136 gabbro (yellow), serpentized peridotite (green), and sulfide (purple). CIR = Central
137 Indian Ridge, NTD = non-transform discontinuity, OCC = oceanic core complex, KHF =
138 Kairei hydrothermal field.

139

140 There is a 26-km long lateral offset, i.e. NTD, between CIR-S1 and S2. Along the
141 trace of this NTD, the ridge-parallel abyssal hill pattern is strewn with several domed
142 highs composed of ultramafic rocks. In the northwest off-axis area of this NTD, a
143 corrugated domed high, together with an exposure of highly deformed serpentized
144 peridotite and a high residual mantle Bouguer anomaly, is recognized as an oceanic core
145 complex (OCC), named the 25°S OCC, which is exposed along a detachment fault
146 (*Mitchell et al.*, 1998; *Morishita et al.*, 2009; *Sato et al.*, 2009). Another OCC (the
147 Uraniwa Hills) was identified 15 km east of the ridge axis, where troctolites, olivine

148 gabbros, and dunite were sampled (*Kumagai et al.*, 2008; *Morishita et al.*, 2009;
149 *Nakamura et al.*, 2009). A unique domed high with exposures of serpentinized peridotite,
150 known as the Phoenix knoll, was confirmed within the southern end of the axial valley of
151 CIR-S2 (*Morishita et al.*, 2014; *Okino et al.*, 2015). Between the two nodal basins at the
152 northern end of CIR-S1 and the southern end of CIR-S2, a shallow domed-like massif has
153 developed with height of 1700 m above the rift valley floor. This massif, the Yokoniwa
154 Rise (YR), is recognized as a non-transform offset (NTO) massif (*Okino et al.*, 2015).
155 Previous sea-surface magnetic surveys identified the Brunhes/Matuyama boundary along
156 the eastern off-axis of CIR-S1 and S2 (*Honsho et al.*, 1999; *Sato et al.*, 2009; *Okino et al.*,
157 2015), suggesting that the YR formed in the Brunhes normal chron (~0.78Ma).

158 Two hydrothermal fields are located within CIR-S1. The Kairei Hydrothermal Field
159 (KHF) was identified in 2000, near the top of the ridge-parallel volcanic knoll (the
160 Hakuho Knoll), on the eastern flank of CIR-S1. The Hakuho Knoll is entirely composed
161 of pillow and sheet lavas. The KHF is characterized by the high hydrogen content of
162 venting fluids and associated chemosynthetic ecosystems (*Gamo et al.*, 2001; *Takai et al.*,
163 2004), which are considered to reflect the interaction of seawater with troctolites beneath

the Uraniwa Hills and/or serpentinization processes along an underlying detachment fault
(*Nakamura et al.*, 2009; *Okino et al.*, 2015).

The YHF was identified in 2009, and is located on top of the YR. Hydrothermal
sulfides and small dead chimneys (< 1 m) were observed, along with weak venting of
low-temperature fluids (< 5 °C) and a small biological community (< 1 m²). Submersible
surveys revealed that the serpentinized peridotite is extensively exposed on the eastern
slope of the YR (25°15'–17'S and 70°04'–05'E), while small quantities of gabbro,
dolerite, and basalt were identified in a few outcrops (*Okino et al.*, 2015; *Sato et al.*, 2015;
Fig. 2). Basalts outcrops on the western slopes of the YR (Fig. 2) are partly covered by
sediments.

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176 Figure 2. Magnetic survey tracks and sample collection

177 Magnetic survey tracks and rock sample collection locations in the Yokoniwa

178 hydrothermal fields (YHF), overlain on the regional bathymetry (m). Solid lines denote

179 the AUV r2D4 survey (light blue) and the HOV *Shinkai6500* (blue). Circles denote basalt

(red), dolerite (brown), gabbro (yellow), serpentized peridotite (green), and sulfide (purple). The contour intervals are 100 m.

3. Methods

3.1 Imaging the magnetic anomaly: AUV

A high-resolution magnetic survey of the YHF region was performed using the AUV *r2D4* (dive#68) during a cruise of the R/V *Hakuho-maru* (KH10-6) in November 2010. The AUV collected magnetic field data along four N–S track lines within an area of ~7 by ~2 km and at an altitude of 30–200 m (mean = 80 m; Fig. 2). A three-axis fluxgate type magnetometer with a sampling rate of 8 Hz was attached to the AUV. A calibration spin of the vehicle, called an 8-figure loop, was performed in order to correct for the effect of vehicle magnetization, which was performed using the calibration technique of Isezaki (1986). Attitude data (heading, pitch, and roll) were collected independently using an optical gyroscope (IXEA PHINS IMU) with a sampling rate of 2 Hz. Magnetic data were merged with attitude data after considering a time shift of 3.75 s between them. The ambient geomagnetic field vector was estimated from the International Geomagnetic

Reference Field (IGRF; *IAGA Working Group V-MOD*, 2010). After the vector summation of three components, the mean IGRF value for the area was subtracted to obtain the total magnetic field anomaly.

Magnetization intensity distributions were calculated using the Akaike's Bayesian Information Criterion (ABIC) inversion method developed by *Honsho et al.* (2012). This approach is a space domain inversion method; therefore, the solution does not suffer from smoothing due to high-cut filtering or error caused by reducing data onto a flat surface, as can occur in other inversion methods (e.g., *Parker and Huestis*, 1974). During this procedure, the magnetic data were merged with positioning and altitude data from a Doppler velocity log (WHN-300, Teledyne RD Instruments). Seafloor bathymetric data utilized in the analysis were obtained by reinterpolation of a 20-m grid based on surface multibeam echo-sounding. The magnetization direction was assumed to be parallel to the IGRF at 25°16'S, 70°04'E (inclination: -60° and declination: -23°). For the magnetized source geometry, we considered four cases: constant 100 m and 500 m thick layers, a half-infinite body, and a layer with a flat lower boundary 200 m below a maximum water depth of seafloor.

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213 ***3.2 Imaging the magnetic anomaly: HOV***

214 Near-seafloor magnetic surveys around the YHF were performed using the HOV
215 *Shinkai6500* during two scientific cruises of the R/V *Yokosuka*. Two dives (6K#1170 and
216 6K#1176) were conducted during the YK09-13-leg2 cruise in August 2009 and another
217 dive (6K#1331) was completed during the YK13-03 cruise of March 2013 (Fig. 2). The
218 HOV *Shinkai6500* covered 7.3 km in distance and provided magnetic data for both the
219 western and eastern slopes of YR between 3500 and 2400 m depth. A three-axis fluxgate
220 type magnetometer was rigidly fixed to the front side of the *Shinkai6500*. Both vector
221 magnetic data and the attitude of the submersible (heading, pitch, and roll) were collected
222 at a sampling rate of 10 Hz. Attitude data were collected using a gyrocompass and an
223 OCTANS motion sensor (IXBLUE Inc., United States). The calibration of vehicle
224 magnetization was performed following the method of *Isezaki* (1986) and using data
225 obtained during the helical loops performed spontaneously at depth of 500 m beneath the
226 sea surface and 500 m above the seafloor. To obtain an optimal solution, a time shift of
227 0.4 s was added to attitude data. All dives were navigated by acoustic ranging using the

super-short baseline positioning system. The distance between the altimeter and the magnetic sensor were taken into account in order to obtain the geometry of the experiment. To estimate magnetization intensity distribution, two analytical methods were applied to all data: a forward modeling technique combined with frequency analysis (*Honsho et al., 2009; Talwani and Heirtzler, 1964*) for short-wavelength variations in the magnetic anomaly, and an inversion technique in the frequency domain (*Honsho et al., 2009; Parker and Huestis, 1974*) for long-wavelength anomalies. To conduct these two-dimensional analyses, the dive tracks of 6K#1170 and 1331 were projected onto a straight line, with the track of 6K#1176 divided into two straight lines because its course changed significantly.

Based on short-wavelength variations in the magnetic anomaly, the absolute magnetization was estimated from the ratio of the observed magnetic anomaly to the synthetic anomaly computed for a 1 A/m magnetization intensity representing the unit magnetization intensity of the seafloor (*Honsho et al., 2009*). The horizontal (x) and vertical (z) components of the synthetic anomaly were calculated using the forward technique of a two-dimensional polygonal body (*Talwani and Heirtzler 1964*), and were

244 obtained every 1 m along each dive paths by summation of anomalies from the upper
245 boundary of the magnetized layer, which was constrained by the topographic surface.
246 Seafloor topography was defined by the depth and altitude of the vehicle at the same
247 interval. The magnetized layer was assumed to be uniformly magnetized with a
248 half-infinite thickness and a magnetization direction parallel to IGRF (inclination: -60° ,
249 declination: -23°). For short-wavelength anomaly variations, the contribution of
250 magnetic sources shallower than 50 m was significant and synthetic anomalies were not
251 affected by the differences in assumed layer thickness when the thickness was more than
252 100 m (Fig. 3). Before comparison between the observed and synthetic anomaly, a
253 band-pass filter of between 20 and 300 m was applied in order to remove both
254 long-wavelength variations and spiky noise. The ratio of the observed magnetic anomaly
255 to the synthetic anomaly ratio, representing absolute magnetization, was calculated by the
256 technique improved by *Honsho et al.*, (2009), based on a linear transfer function in the
257 Fourier domain used to acquire quantitative solutions. Absolute magnetization was
258 estimated at 16 m intervals along the track using a 128 m wide sliding window (segment).
259 Spectrums were then averaged over wavelengths between 32 and 128 m in order to obtain

260 a ratio of observed to synthetic anomalies, together with error, coherency, and phase.
261 These calculations were applied to all sets of segments for the x and z components,
262 providing individual results. The averaged values between the x and z ratio of observed to
263 synthetic anomalies were adopted as our best estimate of absolute magnetization intensity.
264 Estimations with a coherency greater than 0.5 in both components were retained for
265 further interpretation as with *Honsho et al.* (2009). The magnetization polarity was
266 regarded as normal for phases between -90° and 90° and reversed for phases between
267 90° and 270° .

268

269

270 Figure 3. Synthetic downward magnetic anomalies

271 Synthetic downward magnetic anomaly (nT; A) calculated by 2D forward modeling

272 along the track of dive 6K#1331 for dive altitudes of 0–20 m, assuming different

273 magnetic source layer thicknesses: 10 m (blue line), 30 m (light blue line), 50 m (green

274 line), 100 m (red line), and infinite (black broken line). Vehicle altitude (m; B) and

275 vehicle depth (m; C) are shown along the same track, with seafloor topography defined

276 by summation of vehicle altitude and depth.

277

278 While the short-wavelength variation of a magnetic anomaly provides the magnetic
279 structure of the uppermost sub-seafloor, long-wavelength variation gives information on
280 the magnetic layer over a greater depth range. We applied the magnetic inversion
281 technique improved by *Honsho et al.* (2009) in order to take full advantage of the varying
282 altitude of the vehicle above the seafloor. First, magnetic anomaly was inverted to
283 magnetization distribution in a substitute geometry, using the method of *Parker and*
284 *Huestis* (1974), as defined by a level observation plane and pseudo topography (vehicle
285 altitude), the same as equivalent geometry proposed by *Hussenoeder et al.* (1995). In the
286 next step, a synthetic magnetic anomaly was calculated in the real geometry using this
287 inversion solution and the forward technique of *Talwani and Heirtzler* (1964). The
288 difference between the observed and calculated synthetic magnetic anomaly was
289 obtained, and this differential anomaly was inverted for magnetization using the
290 substitute geometry. Finally, the differential magnetization solution was added to the last
291 magnetization obtained in first step. In order to correct the error due to switching

geometry, these procedures were repeated until the inversion solution of magnetization converged (*Honsho et al.*, 2009). A high-cut filter, cosine tapered curves of 1.0 (full pass) at a wavelength of 300 m and 0.0 at a wavelength of 250 m, was used for the inversion calculation. In forward modeling with real geometry, an upper boundary of the magnetized layer was defined by depth and altitude. A uniform magnetized layer with 500 m thickness, and a direction of magnetization parallel to the IGRF was assumed. The annihilator was also solved by the same procedure.

3.3 Rock magnetic properties measurements

Rock magnetic property measurements were performed on collected rock samples, including; 5 sulfides, 11 basalts, 4 dolerites, 2 gabbros, and 30 serpentinized peridotites. Samples were mostly collected on the seafloor around the YHF during the submersible *Shinkai6500* dives of 6K#1170, 6K#1176, and 6K#1331 with the exception of 5 serpentinized peridotite samples, which were obtained by a dredge haul of KH10-6-DR11 conducted during the KH10-6 cruise (Fig. 2). Detailed analyses of the serpentinized peridotites were described in another paper.

308 Rock samples were divided into four cubic subsamples (2-cm cubic) and several
309 chips (~500 mg). For two, small serpentized peridotites, just two cubed materials were
310 prepared. Cubic sub-samples were used for measurements of natural remanent
311 magnetization (NRM), magnetic susceptibility, and bulk grain density. Chips were used
312 for measurements of saturation magnetization, and low- and high-temperature magnetic
313 experiments.

314 NRM intensity and magnetic susceptibility measurements were conducted on all
315 cubed materials using a spinner magnetometer (Natsuhara Giken DSPIN) at the
316 Atmosphere and Ocean Research Institute (AORI), the University of Tokyo, and an
317 AGICO KLY-3 Kappabridge at the Center for Advanced Marine Core Research (CMCR),
318 Kochi University, respectively. Thermomagnetic analyses were performed on chips of
319 representative rock samples. For low-temperature measurements, an isothermal remanent
320 magnetization (IRM) was imparted to the samples using a field of 3 T at 6 K after cooling
321 in a zero field, which was followed by measurement of the IRM during warming up to
322 300 K using a magnetic property measurement system (MPMS, Quantum Design Inc.) at
323 the CMCR. The temperature dependence of saturation magnetization above room

temperature was measured using a magnetic balance (Natsuhara Giken MNB-89) at the CMCR. Samples were heated from 50 to 700°C and cooled back to 30°C in a vacuum (~1 Pa) at a rate of 12 °C/min. The inducing magnetic field was 0.5 T.

4. Results

4.1 Magnetization from AUV data

The distribution of equivalent magnetization (EM) was estimated from the total magnetic anomaly obtained by AUV *r2D4* using the ABIC inversion analysis technique of *Honsho et al.* (2012). The results are shown in Figure 4 and are summarized in Table 1. High amplitude magnetization (−5 to 28 A/m) was estimated when we assumed a 100 m-thick magnetic layer (Fig. 4B). Magnetization intensities showed smaller variation: −2 to 11 A/m in a 500 m-thick magnetic layer model, −2 to 10 A/m in a half-infinite body model, and −2 to 8 A/m in a flat lower boundary model (Fig. 4C–E). Regardless of the magnetic layer model employed, significant positive magnetization (approximately ~30 A/m using 100 m-thick layer model and ~10 A/m using the other thickness layer models) was located on the YHF.

340

Observation		Analysis		Assumption		Result
Platform	Dive#	Applied method	Data	Layer thickness	Magnetization direction	Magnetization amplitude
AUV <i>r2D4</i> (altitude: 80 m)	68	ABIC inversion ^{*1}	All	100 m	IGRF	33 A/m
				500 m	IGRF	13 A/m
				half-inifinite	IGRF	12 A/m
				flat boundary	IGRF	10 A/m
HOV <i>Shinkai6500</i> (altitude: 10 m)	1170, 1176, 1331	Forward modeling ^{*2, *3}	Line 1-4 ($\lambda = 20\text{-}300$ m)	half-inifinite	IGRF	17 A/m
	1331	FFT inversion ^{*2, *4}	Line 1 ($\lambda > 300$ m)	half-inifinite	IGRF	7 A/m
	1170	FFT inversion ^{*2, *4}	Line 2 ($\lambda > 300$ m)	half-inifinite	IGRF	7 A/m
	1176	FFT inversion ^{*2, *4}	Line 3 ($\lambda > 300$ m)	half-inifinite	IGRF	10 A/m
	1176	FFT inversion ^{*2, *4}	Line 4 ($\lambda > 300$ m)	half-inifinite	IGRF	3 A/m

References are *1, *Honsho et al.* (2011); *2, *Honsho et al.* (2009); *3, *Talwani and Heirtzler* (1964); *4, *Parker and Huestis* (1974). Abbreviations are AUV, autonomous underwater vehicle; HOV, human occupied vehicle; λ , wavelength; IGRF, international geomagnetic reference field; ABIC, Akaike's Bayesian information criterion.

341

342 Table 1. Summary table of the magnetic inversion and forward modeling results

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344

Figure 4. Total magnetic anomaly and equivalent magnetization distribution

Total magnetic anomaly (nT; A) for track AUV *r2D4*, and the equivalent magnetization (A/m) distribution from the magnetic anomaly assuming a 100 m-thick layer (B), a 500 m-thick layer (C), half-infinite body (D), and a layer with flat lower boundary at maximum water depth plus 200 m (E). The inversion solution was estimated using the Akaike's Bayesian Information Criterion (ABIC) inversion technique developed by *Honsho et al.* (2012).

4.2 Magnetization from HOV data

Absolute magnetization (AM) distribution was estimated from the short-wavelength magnetic anomaly obtained by HOV *Shinkai6500* using forward modeling and frequency analysis techniques (*Honsho et al.* 2009; *Talwani and Heirtzler*, 1964). The results for all 4 lines were plotted together with the bathymetric signature in Figure 5. The estimated values of AM with high coherency (> 0.5) ranged from 0.4–16.7 A/m (Fig. 5B). The AM values of lines 1, 2, 3, and 4 were 2.4–16.7, 1.0–6.9, 0.4–4.6, and 1.4–2.4 A/m, respectively. Strong magnetization higher than 7 A/m was

observed only for line 1 on the western slope of YR, where basalt and hydrothermal sulfide samples were collected. The AM values were up to 8 A/m around the YHF. Approximately 50% of the reliable estimates within every segment were acquired along line 1, and almost all estimates with high coherency along line 1 showed a positive phase (-90 to 90° ; Fig. 5B), representing the acquisition of remanence during the normal (positive) magnetic chron. In contrast, on the eastern slope, where serpentinized peridotite dominates, low AM values less than 7 A/m were observed (Figs. 5C, 5D, and 5E), although gabbro and dolerite were also collected at this location. For lines 2, 3, and 4 on the eastern slope, reliable estimates with high coherency represented less than 20 % for which the phase distribution was complicated (Figs. 5C, 5D, and 5E), possibly due to the low signal-to-noise ratio caused by the weakly magnetized crust, or reflecting a disturbance in the magnetization direction resulting from faulting, and/or block rotation or due to geometry error caused by the sedimentation.

Figure 5. Forward results of the short-wavelength anomaly from the submersible

(A) Projected 2D tracks of the vehicle (solid lines) overlain on regional bathymetry (km), with the original vehicle tracks (broken line). The results of absolute magnetization (AM; A/m), and phase calculated for each set of the segments from short-wavelength magnetic anomalies in lines 1, 2, 3, and 4 are shown in (B), (C), (D), and (E). For B–E, the top plots show horizontal (x) and vertical (z) components of both observed (red line) and synthetic (blue line) magnetic anomalies; the middle plots show the distribution of AM averaged by the calculated ratio of the x (broken lines) and z (solid lines) components; the bottom plots show the distribution of phase of the x (broken lines) and z (solid lines) components. Reliable results, as confirmed by high coherency (> 0.5 in both x and z), are indicated by solid circles. Gray and open circles show results with middle (0.3–0.5) and low (< 0.3) coherency. Collected rock sample locations are shown together, including basalt (BS), dolerite (DL), gabbro (GB), serpentized peridotite (SP), and hydrothermal sulfide (HS). Resultant phase values with high coherency (> 0.5) were shown as squares for x component and as triangles for z component.

EM distribution was estimated from the long-wavelength magnetic anomaly obtained by HOV *Shinkai6500* using an inversion method in equivalent geometry (Honsho *et al.*, 2009; Parker and Huestis, 1974). The results along all 4 lines were plotted together with the bathymetric signature and collected rock species in Figure 6. Along line 1, relatively high EM intensity was recognized in the western region ($x = 100\text{--}700$ m) where basalts were observed (Fig. 6B). The EM value increased from -5 A/m at the minimum point ($x = 1000$ m) to 2 A/m at the eastern end, where the YHF is located (Fig. 6B). For line 2, the maximum EM value of 4 A/m was observed in the easternmost part at 3500 m deep. The EM intensity decreased to -3 A/m near the area of $x = 1600\text{--}1900$ m at a depth of $2900\text{--}3000$ m, and increased to $0\text{--}1$ A/m at 2600 m depth (Fig. 6C). The largest variation in EM, up to 10 A/m, was observed in line 3 (Fig. 6D), with the lowest value at the eastern end of the slope at a depth of 3050 m, and increasing to the west. High EM values were observed in the area of $x = 1100\text{--}1700$ m. For line 4, the EM showed a high value (~ 3 A/m) near the YHF ($x = 0$), which was larger than that of the southern area (Fig. 6E).

Figure 6. Inversion results of the long-wavelength anomaly from the submersible (A) Equivalent magnetization (A/m) distribution along projected 2D tracks in map view (overlain on regional bathymetry; km). Dotted line with colored circles show the original HOV track and collected rock species. (B–E) Results of line 1–4, respectively, with the top plots showing obtained equivalent magnetization (A/m; black) derived from downward component, and the annihilator (grey). Middle plots show a comparison between the observed anomaly (red line) and the synthetic anomaly (blue line), reproduced from the obtained magnetization distribution and the real geometry between the vehicle and topography. The bottom two plots show altitude (m) and the depth of the vehicle (km).

4.3 Magnetic properties of rock samples

The NRM intensity (A/m) and magnetic susceptibility (SI) were 0.02–0.09 (mean: 0.05) and 0.0008–0.0016 (mean: 0.0010), respectively, in sulfide, 0.01–6.33 (mean: 1.48) and 0.0007–0.0110 (mean: 0.0025) in basalt, 0.05–0.44 (mean: 0.18) and 0.0017–0.0143 (mean: 0.0063) in dolerite, 0.85–1.78 (mean: 1.31) and 0.0061–0.0268 (mean: 0.0165) in

gabbro, and 0.20–8.35 (mean: 2.17) and 0.0016–0.0867 (mean: 0.0347) in serpentinized peridotite (Fig. 7A). Koenigsberger ratio values (Q : NRM/induced magnetization) were 0.7–2.7 (mean: 1.3) in sulfide, 0.3–74.1 (mean: 37.1) in basalt, 0.2–3.6 (mean: 1.5) in dolerite, 0.9–8.1 (mean: 4.5) in gabbro, and 0.4–5.1 (mean: 1.9) in serpentinized peridotite (Fig. 7A). Induced magnetization was calculated for a magnetic field of 45000 nT.

Figure 7. Results of rock magnetic properties for various host rock samples

(A) Log-log plot of magnetic susceptibility (K ; SI) and natural remanent magnetization (NRM; A/m) of host rocks collected from the Yokoniwa Rise. Broken lines show constant Q (Koenigsberger ratio), calculated for a magnetic field of 45000 nT. (B) Magnetization (A/m) estimated by the sum of NRM, with induced magnetization

calculated for a magnetic field of 45000 nT. Rock samples include: sulfide (purple),
basalt (red), dolerite (brown), gabbro (yellow), and serpentinized peridotite (green).

Basalt samples were divided into two groups of low and high Q values, possibly
reflecting the degree of low-temperature oxidation and grain size distribution of
ferromagnetic minerals (mostly titanomagnetite). A similar signature was also observed
in dolerites. Sulfide samples had weak magnetic signatures with low magnetic
susceptibility and NRM intensity. Serpentinized peridotites carried strong NRM and had
high magnetic susceptibility. In more than 80% of serpentinized peridotites,
magnetization primarily originated from remanence rather than from induced
magnetization ($Q > 1$). Magnetization for each rock sample was estimated using the sum
of NRM with induced magnetization (Fig. 7B). It was assumed that the NRM vectors
were parallel to the current geomagnetic field vector; therefore magnetization values
represent the maximum estimation for in-situ magnetization contributing to the magnetic
anomaly. This assumption was reasonable because the rocks appeared to have acquired
the NRM (thermoremanent magnetization and chemical remanent magnetization) during

the Brunhes normal polarity epoch when the YR evolved (*Honsho et al.*, 1999; *Sato et al.*, 2009; *Okino et al.*, 2015). However, there is no available data showing the direction of their NRM. The serpentinized peridotites exhibited the strongest magnetization among the tested samples, although there is a large variation in the magnetization values. High- and low-temperature experiments showed a Curie temperature of $\sim 580^{\circ}\text{C}$ and a rapid transition between 90 and 120 K, which were interpreted as the Verwey transition (e.g., *Ozdemir et al.*, 1993), suggesting that the magnetic carrier in serpentinized peridotite was magnetite.

The variation of magnetization (under the 45000 nT field) in serpentinized peridotites was considered as a function of serpentinization degree (Fig. 8). We estimated the degree of serpentinization from bulk grain density and saturation magnetization (equivalent to magnetite content) using the empirical equation proposed by *Oufi et al.* (2002). The magnetization appeared to vary nonlinearly, as a function of serpentinization degree. The maximum value of magnetization was up to 11 A/m for a sample with a serpentinization degree of 96%. The details of this analysis will be described in another paper.

469

470

471 Figure 8. Magnetization and degree of serpentinization of serpentinized peridotites

472 Magnetization (A/m) and serpentinization degree (%) of 30 serpentinized peridotites.

473 Magnetization was estimated by the sum of natural remanent magnetization (NRM) with

474 induced magnetization calculated for a magnetic field of 45000 nT.

475

476 **5. Discussion**

477 **5.1 Basaltic volcanism on the western slope of the Yokoniwa Rise**

478 Magnetization polarities from the HOV magnetic anomaly were entirely positive on
479 the lava-dominant area of the western slope. These results are consistent with previous
480 sea-surface magnetic studies, which identified the Brunhes/Matuyama boundary along
481 the eastern off-axis of CIR-S1 and suggested that the YR formed in the Brunhes normal
482 chron (*Honsho et al.*, 1999; *Sato et al.*, 2009; *Okino et al.*, 2015). On the basis of these
483 results, we consider that these basaltic lavas erupted and acquired thermal remanent
484 magnetization during the Brunhes normal epoch.

485 The absolute magnetization from the HOV magnetic data, analyzed for a short
486 wavelength signal, showed the highest magnetization (~16 A/m) for areas in which
487 basaltic pillow lavas are present (Fig. 5B). However, this signature was not observed in
488 the inversion results using a long wavelength signal (Figs. 6A and 6B), or in the results
489 from the AUV data (Fig. 4). Short wavelength anomalies strongly reflect shallow
490 sub-seafloor magnetization, so that these results imply that the locally high magnetization
491 of the lava-hosted area reflects high magnetization of the basaltic lavas. Previous studies
492 of oceanic basalt NRM intensities at spreading ridges proposed that zero-age basalts carry

strong NRM (several tens A/m) and their intensity is drastically decreased to several A/m with age following low-temperature oxidation (e.g., *Gee and Kent, 1994; Johnson and Atwater, 1977; Johnson and Tivey, 1995*). Moreover, the geomagnetic field intensity has doubled over past several tens of thousands of years (e.g., *Guyodo and Valet, 1996*), which constitutes a considerable signal that cannot be ignored. Forward models of near-seafloor magnetic data (*Schouten et al., 1999*) demonstrated that observed character of the central anomaly magnetization high (known as CAMH) was successfully reproduced by the convolution of a lava deposition distribution with age-dependent lava magnetization function, which is the product of geomagnetic paleo-geomagnetic field intensity and low-temperature alteration. Therefore, the high magnetization observed on the western slope of the YR (line 1) appears to originate from relatively young basalt.

However, basaltic rock samples, collected and analyzed in this study, showed low values of NRM (~2 A/m). Because these samples were collected from the lava surface, we consider that the low NRM values are attributed to low-temperature oxidation affecting on the lava surface region (e.g. *Gee and Kent, 1994; Irving, 1970; Johnson and Atwater, 1977*). Indeed, the high-temperature saturation magnetization measurements for

these basaltic rock samples showed large differences between heating and cooling processes, suggesting that low-temperature oxidation of the rock samples has progressed to some extent. The difference between magnetization from the magnetic anomaly and from rock samples possibly reflects spatial differences in the oxidation degree within the lava-scale (Fujii *et al.*, 2015). The low-temperature oxidation process must be faster on the surface than in the interior of the lavas, resulting in weaker NRM of rock samples than magnetization from magnetic anomaly representing bulk magnetization of accumulated lava flows.

5.2. Formation of the Yokoniwa Hydrothermal Field strong magnetization zone

One of the most significant features observed in our magnetic data was the clear association between the YHF and enhanced crustal magnetization, which was higher than the surrounding seafloor (Figs. 4, 5, and 6). Equivalent magnetization estimated by the AUV for the assumed layers of 100 m and > 200 m increased by approximately 30 A/m and 10 A/m at the YHF, respectively (Fig. 4). Absolute magnetization estimated from the HOV anomaly, which reflects the uniform magnetization of the uppermost sub-seafloor

525 < 50 m, was ~8 A/m at the YHF (Fig. 5). Equivalent magnetization of the
526 long-wavelength anomaly in the HOV data increased by 8 A/m at the YHF (Fig. 6). The
527 magnetization within 100 m layer thickness (30 A/m) for explaining the AUV data is not
528 consistent with the absolute magnetization (~8 A/m) from the HOV *Shinkai6500*.
529 Therefore, shallow concentration of magnetization of 30 A/m overestimated the magnetic
530 anomaly from the HOV *Shinkai6500*. Consequently, the AUV and HOV datasets imply
531 that approximately 10 A/m of the magnetization for the > 200 m thickness layer
532 represents reliable explanation for the magnetic anomaly highs in the YHF.

533 The rock magnetic properties showed that the highly serpentinized peridotite was
534 sufficient to represent a reasonable source for the magnetic high at the YHF (Fig. 7 and
535 10). The value of magnetization in serpentinized peridotites with a serpentinization
536 degree of 96% was up to 11 A/m, consistent with the high magnetization value of the
537 YHF deduced from the magnetic anomaly observations. The measured Curie temperature
538 (~580°C) and confirmed Verwey transition (90–120 K) in serpentinized peridotites
539 showed that magnetite is the main magnetic carrier, while the wide range in
540 Koenigsberger ratios (0.4–5.1) suggests the variation in grain size of magnetite.

541 Saturation magnetization measurements showed that magnetite abundances are up to
542 5.5 wt% with grains occurring within veins in olivine mesh structures. These
543 observations strongly suggest that serpentinized peridotites hosted magnetite grains
544 formed due to a reaction with hydrothermal fluids, and appear to have acquired their
545 chemical remanent magnetization during their growth.

546 As discussed in section 5.1, the highest absolute magnetization based on the
547 short-wavelength anomaly was distributed in the lavas-hosted area, but basaltic lavas
548 cannot be a source of the magnetization high around the YHF, because no lava flows were
549 observed around the YHF. Even if there were an intruded body beneath the hydrothermal
550 sulfide deposit, it is unlikely that the intruded body could carry a strong remanence (up to
551 10 A/m) with a spatial extent of a few hundred meters. Dike sections generally show low
552 NRM intensity, and provide only minor contributions to marine magnetic anomalies (e.g.
553 *Pariso and Johnson, 1991; Tivey, 1996*). Hydrothermal processes decrease the
554 magnetization of lavas due to the alteration of the strongly magnetic titanomagnetite to
555 less magnetic titanomaghemite and to nonmagnetic minerals (e.g., *Irving, 1970; Johnson*
556 *and Atwater, 1977*).

Tivey (1994) and *Gee et al.* (2001) proposed that a combination of magnetic pyrrhotite and magnetite mineralization was responsible source for the magnetic anomaly at the Bent Hill massive sulfide in the Middle Valley, the JFR. There, samples of the Ocean Drilling Program (ODP) drill cores contained predominantly pyrite (90%), with 2–10% magnetite and smaller amounts of hematite, pyrrhotite and other sulfides (*Korner*, 1994). In the YHF, sulfide samples show low magnetic susceptibility (<0.002 SI) and NRM intensity (<0.1 A/m). Thermomagnetic curves show no evidence for the presence of pyrrhotite. Therefore, we concluded that sulfide mineralization does not contribute to crustal magnetization. Although systematic variation in the degree of oxidation with depth was not evident, we speculate that enhanced seawater or hydrothermal circulation may have led to oxidative conditions under which pyrrhotite cannot exist.

In summary, based on the results of this study, we propose that enhanced magnetization at the YHF results from an abundance of magnetite within the highly serpentinized peridotites. The accumulation of dead chimneys and low-temperature venting in the YHF implies a long-lived hydrothermal system that is significant enough to react with sub-seafloor ultramafic rocks.

573

574 ***5.3 Magnetic mineral formation in ultramafic-hosted hydrothermal systems***

575 Enhanced magnetization in ultramafic-hosted hydrothermal fields has also been
576 reported at the Rainbow and Ashadze-1 sites of the MAR (*Szitkar et al.*, 2014; *Tivey and*
577 *Dyment et al.*, 2010). Both are high-temperature active vents with fluid temperatures of >
578 300°C (*Fouquet et al.*, 2010). *Szitkar et al.* (2014) interpreted the high magnetization zone
579 to reflect strongly magnetized magnetite in the stockwork created during serpentinization
580 of ultramafic rocks. *Tivey and Dyment et al.* (2010) proposed that the magnetization high
581 at the Rainbow reflects the presence of a magnetic iron sulfide (pyrrhotite).
582 Magnetization intensity from magnetic anomalies increases ~30 A/m at the Rainbow
583 (*Szitkar et al.*, 2014b), which is much larger than that observed at the YHF (~10 A/m). In
584 contrast, the magnetization intensity of the Ashadze 1 (~4 A/m) is smaller than that of the
585 YHF (*Szitkar et al.*, 2014b). Moreover, strong magnetization associated with
586 hydrothermal sites was also observed at the Bent Hill massive sulfide on the JFR, where
587 the magnetic source was considered to be the deposition of significant magnetite and
588 pyrrhotite (*Gee et al.*, 2001; *Tivey*, 1994).

589 At the Rainbow site, the maximum values for magnetization of serpentized
590 peridotites in the stockwork zone are less than 13 A/m, as calculated using magnetic
591 susceptibility (0.007 SI) and NRM intensity (12 A/m) (*Szitkar et al.*, 2014b). The main
592 magnetic bearer is magnetite, based on the Curie temperature of ~580 °C (*Szitkar et al.*,
593 2014b). However, this magnetization value does not sufficiently explain the
594 magnetization intensity from the magnetic anomaly (~30 A/m) (*Szitkar et al.*, 2014b). As
595 discussed by *Tivey and Dymant et al.* (2010), they considered the presence of strongly
596 magnetized magnetic iron sulfides (pyrrhotite) to be as the part of sources of the magnetic
597 anomaly in the Rainbow site. The abundant hydrogen content (~16 mM) resulting from
598 the serpentinization process created a more reducing environment, which may protect
599 pyrrhotite from oxidation (*Charlou et al.*, 2010). One noteworthy feature is that the
600 highest magnetization intensity (~13 A/m) of the serpentized peridotite from the
601 Rainbow is consistent with the highest value (~11 A/m) for the Yokoniwa sample,
602 implying that magnetization of ~10–15 A/m is the contributing magnetization value of
603 magnetite-rich serpentized peridotite. The contribution of pyrrhotite needs to be
604 investigated in future studies.

605 Based on the differences of magnetization intensity, we propose a 3-stage model for
606 magnetic mineral formation in ultramafic-hosted hydrothermal fields (Fig. 9). During the
607 initial stage of an ultramafic-hosted hydrothermal system, magnetite forms with
608 serpentine and hydrogen through hydrothermal alteration of peridotites. Produced
609 magnetite grains during serpentinization are increased as reaction progresses (*Klein et al.*,
610 2013a; *Oufi et al.*, 2002; *Malvoisin et al.*, 2012). Strongly magnetized pyrrhotite creation
611 also occurs through sulfide mineralization under reductive conditions (with the presence
612 of pyrrhotite confirmed at the several hydrothermal fields; *Fouquet et al.*, 2010). We
613 consider the Ashadze-1 site to be within this stage of development (Stage 1 in Fig. 9). The
614 hydrothermal alteration in the Ashadze-1 site has not progressed so that the magnetite and
615 pyrrhotite concentrations remain low; thus, the magnetization intensity is also low (~4
616 A/m). Once serpentinization reaction has progressed, the amount of created magnetite
617 increases drastically, strengthening magnetization (*Klein et al.*, 2013a; *Oufi et al.*, 2002;
618 *Malvoisin et al.*, 2012). Pyrrhotite creation continues as long as the hydrogen content of
619 the hydrothermal fluids continues to create a reducing environment (e.g., *Bach et al.*,
620 2004). Moreover, hot and reduced hydrothermal fluids keep magnetite and pyrrhotite

621 stable without low-temperature oxidation, i.e., reacting with cold seawater; thus, it retains
622 magnetization. We consider the Rainbow (~30 A/m) to be in this stage of development
623 (Stage 2 in Fig. 9). Once the reaction of ultramafic rocks ceases, hydrogen is no longer
624 formed in the system and conditions become oxidative, which allows pyrrhotites to
625 convert into nonmagnetic iron sulfide or oxide, reducing their magnetization considerably.
626 We consider the YHF (~10 A/m) to be in this stage of development (Stage 3 in Fig. 9).

627 It should be noted that the amount of magnetite creation is known to be controlled
628 by various factors such as temperature and primary rock compositions. Natural samples
629 (*Klein et al.*, 2009; 2013a) and thermodynamic equilibrium models (*Klein et al.*, 2013b)
630 demonstrated that serpentinized peridotite with abundant magnetite forms at
631 temperatures higher than 200°C with the involvement of Fe-poor brucite. While
632 serpentinization of olivine-dominated lithology results in the formation of serpentine,
633 brucite, and magnetite, that of orthopyroxene-dominated lithology results in the
634 formation of serpentine and talc with little magnetite (*Klein et al.*, 2013b). We have not
635 taken into account these factors affecting the progress of serpentinization in our model,
636 because alteration conditions of the Yokoniwa peridotites have still been uncertain. In

addition, the formation and weathering condition of magnetic sulfide minerals, as well as their relation to the magnetite creation during serpentinization, have not been fully understood. However, our simplified conceptual evolutionary model can explain well the observations. In order to confirm our preferred 3-stage model for magnetic mineral formation in ultramafic-hosted hydrothermal system, as well as to make a more comprehensive model, the contribution of pyrrhotite to crustal magnetization needs to be understood by future investigation based on drilled sub-seafloor samples.

Figure 9. Model of magnetic mineral formation

Model showing the 3-stage formation history of magnetic minerals in ultramafic-hosted hydrothermal systems. Stage 1: some magnetized magnetite and pyrrhotite form thorough serpentinization and sulfide mineralization under reductive conditions, Stage 2: large quantity of magnetized magnetite and pyrrhotite are accumulated thorough serpentinization

and sulfide mineralization under reductive conditions, Stage 3: only magnetized magnetite remains as main magnetic source under oxidative condition after serpentinization of host rock.

6. Conclusions

We completed near-seafloor magnetic field measurements using the AUV *r2D4*, and the HOV *Shinkai6500* and rock sampling by dredge haul and the *Shinkai6500* from around the inactive hydrothermal field Yokoniwa of the CIR. The integrated analysis of the magnetic anomaly and rock magnetic properties leads to the following conclusions:

1. The inactive ultramafic-hosted hydrothermal site Yokoniwa is associated with a strong magnetization intensity of ~ 10 A/m. Conversely, the surrounding seafloor does not exhibit such a magnetic signature.

2. The highly serpentinized peridotites have higher magnetic susceptibility and NRM intensities than basalts, dolerites, gabbros, and sulfides. The value of magnetization for the highly serpentinized peridotites, which is carried by constituent magnetite, is sufficient to explain the crustal magnetization intensity observed near the YHF. Therefore,

667 we conclude that the strong magnetization intensity at the YHF reflects the formation of
668 abundant magnetite within highly serpentinized peridotites, resulting from hydrothermal
669 circulation.

670 3. The YHF is considered to be in the final stages of an ultramafic-hosted
671 hydrothermal system, which has developed in a non-transform offset, similar to the
672 Rainbow and Ashadze-1 sites of the MAR. Its magnetization intensity implies that the
673 ultramafic bodies were completely serpentinized and that sulfide mineralization does not
674 contribute to the magnetic anomaly (i.e., strongly magnetized iron sulfides such as
675 pyrrhotite do not exist here).

676 Near-seafloor magnetic field measurements reveal the site-to-site variations
677 between ultramafic-hosted hydrothermal sites, which result from differing degrees of
678 serpentinization and sulfide mineralization through the reaction of host rock with
679 hydrothermal fluids. These observations highlight the importance of subsurface chemical
680 processes in hydrothermal systems and in the geochemical budget of oceanic lithosphere.
681 Moreover, the detection of a magnetic signature provides a powerful tool for
682 investigating active and inactive hydrothermal fields and the mineral deposits associated

683 with hydrothermal activity.

684

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702

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