The Magnetic Signature of Hydrothermal Systems in Slow Spreading Environments

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1. Abstract

Slow-spreading midocean ridges like the Mid-Atlantic Ridge host a remarkable diversity of hydrothermal systems including vent systems located on the neovolcanic axis, large axial volcanoes, in transform faults and non-transform offsets, and associated with low-angle detachment faults, now recognized as a major tectonic feature of slow spreading environments. Hydrothermal systems are hosted in various lithologies from basalt to serpentinized peridotite and exposed lower oceanic crust. The substantial variations of hydrothermal processes active in these environments have important implications for the magnetic structure of oceanic crust and upper mantle.

Hydrothermal processes can both destroy the magnetic minerals in basalt, diabase and gabbro and create magnetic minerals by serpentinization of ultramafic rocks and deposition of magnetic minerals. We report on the diversity of magnetic anomaly signatures over the vent systems at slow spreading ridges and show that the lateral scale of hydrothermal alteration is fundamentally a local phenomenon. This highly focused process leads to magnetic anomalies on the scale of individual vent fields, typically a few 100 meters or less in size. To detect such features, high-resolution, near-bottom magnetic surveys are required rather than sea surface surveys. High-resolution surveys are now more tractable with deep-towed systems, dynamically-positioned ships and with the recent development of autonomous underwater vehicles, which allow detailed mapping of the seafloor on a scale relevant to hydrothermal activity. By understanding these present-day active hydrothermal systems, we can explore for yet to be discovered buried deposits preserved off-axis, both to determine past history of hydrothermal activity and for resource assessment.
2. Introduction

It has long been recognized that geothermal activity can produce distinctive magnetic anomalies in terrestrial volcanic rocks (e.g. Hochstein and Soengkono, 1997). There are numerous ground and aeromagnetic surveys documenting the occurrence of reduced or weakly magnetic anomalies over geothermal areas in New Zealand and Iceland (e.g. Watson-Munro, 1938; Studt, 1959; Palmason, 1975). The source of the reduced magnetization has variously been attributed either to the high temperatures found in these systems that exceed the Curie temperature of the magnetic minerals in the host rock [Watson-Munro, 1938] or to hydrothermal alteration and destruction of the magnetic minerals [Studt, 1959; Browne, 1978]. A review of the early literature by Hochstein and Soengkono [1997] revealed that alteration of the magnetic mineral magnetite to less magnetic minerals was by far the most important mechanism in generating the anomalies. Interestingly, they also found that in liquid-dominated geothermal systems, magnetite was the first mineral to be replaced by non-magnetic minerals such as pyrite, but in vapor-dominated systems that were oxidizing, magnetite is stable and remains magnetic. It was further noted, however, that geothermal systems often went through phases of liquid- and then vapor-dominated periods, so that even a temporary stage of liquid-dominated activity could permanently erase the magnetic properties of the host rock [Hochstein and Soengkono, 1997].

In the marine realm, the recognition of magnetic anomalies associated with hydrothermal systems has been slower to gain acceptance. While early studies of the rock magnetic properties of basalt ranging from terrestrial lava outcrops [Ade-Hall et al., 1971], to ocean crustal drilling [Ade-Hall et al., 1973], dredged samples [Irving, 1970] and lab experiments [Johnson and Merrill, 1972; 1973] all demonstrated that magnetization was significantly affected by low-temperature alteration,
this process was considered relatively regional and broad-scale in effect [Johnson and Atwater, 1977]. Other than an overall decrease in magnetic anomaly amplitude, the effect of concentrated high-temperature hydrothermal alteration was not considered widespread or common. Although it was recognized early on that hydrothermal mineral deposits could form at mid-ocean ridge (MOR) spreading centers [Bostrom and Peterson, 1966; Spooner and Fyfe, 1973; Hutchinson, 1973; Sillitoe, 1973; and Bonatti, 1975], it was not until the discovery of the mineral-rich sediments in deep brine pools of the Red Sea [Degens and Ross, 1969] and active hot springs on the seafloor in the Galapagos Rift [Corliss et al., 1979] that the possibility of crustal alteration associated with hydrothermal vent systems became evident. Clues to the possibility of magnetically disturbed areas in upflow zones related to hydrothermal activity came from studies in ophiolites. A detailed ground level magnetic survey over a massive sulfide body in the Troodos ophiolite complex in Cyprus revealed a strong magnetic low over the stockwork zone, which is hosted within a basaltic lava sequence at Agrokipia [Johnson et al., 1982]. Rock magnetic studies revealed that the titanomagnetite in the basalt had been replaced with non-magnetic sulfide minerals and the basalt itself had been altered to a clay, silica and chlorite assemblage within the stockwork upflow zone [Johnson et al., 1982; Hall, 1992].

Rona [1978] published on the possible magnetic signature of hydrothermal alteration and volcanogenic mineral deposits in oceanic crust and proposed that magnetic anomalies might be an effective method of detecting and characterizing such areas and deposits. Since that time, we have seen the discovery of over 200 active vents on the seafloor in many different environments [Hannington et al., 2005]. As we have come to learn, the detection of water column plumes from active vents and seafloor photography and visual imagery of distinctive biological communities
have become a successful method of vent discovery. The geophysical signature of hydrothermal systems based on crustal properties has been a lot more difficult to employ primarily due to the small scale of the hydrothermal features (e.g. typically only a few 100s of meters across) compared with the ocean depth (kilometers), which imposes a resolution limitation on sea surface measurements. Progress has been made, however, with the advent of deep-towed geophysical surveys [Tivey et al., 2003], near-bottom submersible and Remotely-Operated Vehicle (ROV) surveys [Tivey et al., 1993; 1996; Dyment et al., 2005] and most recently autonomous underwater vehicle (AUV) surveys [Tivey et al., 1997; 1998; 2006]. These high-resolution surveys now reveal that hydrothermal systems do indeed have a detectable geophysical response and we can now embark on the next challenge of documenting the detailed subsurface structure of these systems and to apply these techniques to finding older and non-venting deposits both on and off the MOR. In the following discussion, we have chosen to focus primarily on the magnetization response, which removes the latitude skewing effect of the magnetic field and magnetization vector directions on the magnetic anomaly signal. Magnetization is typically computed from magnetic field data by assuming the direction of the magnetic field and magnetization vector and a source body geometry, either constant layer thickness or some geometrical shape (e.g. pipe) [e.g. Parker and Huestis, 1974].

Not all types of hydrothermal activity will result in reduced magnetization and magnetic mineral destruction. In some environments hydrothermal activity can result in magnetite formation. The process of serpentinization can lead to significant magnetite deposition and there is now clear evidence that magnetite formation can occur not only through normal seafloor alteration of peridotite [Dunlop and Prevot, 1982; Smith and Banerjee, 1985; Bina and Henry, 1990;...
Nazarova, 1994; Oufi et al., 2002], but also can occur where there are vigorous hydrothermal systems hosted in serpentinized peridotite bodies [Murton et al., 1994; German et al., 1996; Gebruk et al., 1997; Charlou et al., 1998; Barriga et al., 1998; Kelley et al., 2001] that have ongoing magnetite formation. These ultramafic-hosted vent systems appear to produce positive magnetic anomalies either related to the formation of magnetite through serpentinization [Dyment et al., 2005] or by the mineralization process.

Magnetite can also form as part of the mineralization process. An example is the sediment-hosted Bent Hill hydrothermal system in Middle Valley on the northern Juan de Fuca Ridge, which produces a positive magnetic anomaly related to a magnetite zone at depth [Tivey, 1994], that was subsequently confirmed by drilling [Zierenberg et al., 1998]. Other magnetic minerals can form as a consequence of hydrothermal activity. The most common is pyrrhotite (Fe$_7$S$_8$), an iron sulfide that has magnetic behavior that varies according to its crystal structure and composition. Monoclinic pyrrhotite is weakly magnetic (ferromagnetic) with a Curie temperature of 325°C, but the magnetic properties weaken with greater sulfur content as the iron sulfide becomes more pyritic in composition (FeS$_2$). Hexagonal pyrrhotite (Fe$_9$S$_{10}$) is nonmagnetic (antiferromagnetic) at room temperature although it is weakly magnetic (ferromagnetic) above its lambda transition temperature of ~200°C when thermally activated vacancy ordering occurs until its Curie temperature at 265°C is reached. Some of the vent structures on the Juan de Fuca ridge contain pyrrhotite [Tivey and Delaney, 1985; 1986].

The slow-spreading Mid-Atlantic Ridge (MAR) shows a remarkable diversity in the types of hydrothermal systems with significant differences in the tectonic and volcanic setting combined
with highly variable host rock lithologies (see Table-1 and Fig. 1). Basalt-hosted hydrothermal vent systems range from what are now recognized as low-angle detachment fault controlled systems such as TAG (26°N) and the recently discovered systems at 13-14°N [Murton et al., 2007; Searle et al., 2007], to vents directly located at the neovolcanic axis such as Snake Pit at 23°N and Turtle Pits at 4°48’S, to those associated with axial volcanoes such as Lucky Strike and Menez Gwen [Langmuir et al., 1997; Ondreas et al., 1997]. Ultramafic hydrothermal vent systems range from the peridotite-hosted Logatchev and Rainbow vent systems found at segment discontinuities [German et al., 1996; Fouquet et al., 1997; Gebruk et al., 1997] to the carbonate dominated Lost City located on the Atlantis transform at 30°N [Kelley et al., 2001]. Below we discuss the magnetic properties and resultant magnetic anomalies found over this remarkable range of hydrothermal systems present on the slow-spreading MAR. Magnetic anomalies are one of the few methods at present that provide insight into the subsurface structure of active hydrothermal systems. Understanding the geometry and distribution of this subsurface crustal structure of active hydrothermal vent systems should allow us to more confidently search and characterize those ancient deposits now preserved off-axis on the flanks of the MOR system.

3. Magnetism of basalt-hosted systems on the MAR

3.1 TAG Hydrothermal Field (26° 08’N 44°45’W)

The Trans-Atlantic Geotraverse (TAG) program in 1973 provided the first clues to the existence of hydrothermal activity on the slow spreading Mid-Atlantic Ridge near 26°N [Rona, 1973]. A water temperature and camera survey of the eastern rift valley wall showed evidence of elevated temperature near the seafloor and low temperature alteration and mineral deposition [Rona et al., 1975; Scott et al., 1974]. A detailed sea surface magnetic survey of the region (Plate 1a,b)
revealed a distinctive zone of reduced magnetic intensity directly over the spreading axis [McGregor and Rona, 1975; McGregor et al., 1977; Rona, 1978]. Detailed magnetic modeling of this anomalous zone suggested that block rotation could not cause the reduction in magnetic intensity and it was suggested instead that hydrothermal alteration was responsible for this reduction in magnetic intensity [McGregor et al., 1977]. It wasn’t until many years later in 1985 that active black-smoker high temperature hydrothermal activity was discovered in the TAG area [Rona 1985; Rona et al., 1986]. A large circular mound ~100 m in diameter and 50 m high was documented at 26°08’N and 44°45’W at the foot of the eastern rift valley scarp with a central black smoker complex discharging 350°C fluids [Campbell et al., 1988].

To investigate the relationship between reduced magnetic anomaly intensity and crustal magnetization, Wooldridge et al., [1990] studied the rock magnetic properties of dredged and submersible rock samples from TAG and other known or suspected hydrothermal sites on the MAR and Gorda Ridge in the Pacific. Contrary to what was expected, the basalt samples showed no evidence of pervasive hydrothermal alteration and showed instead high natural remanent magnetization (NRM), high Koenigsberger ratio values (the ratio of remanent to induced magnetization) and strong stability, in other words, the typical magnetic parameters of young, relatively fresh, MOR basalts. These results suggest that if hydrothermal activity were indeed affecting oceanic crustal magnetization, then the zones of alteration are discrete and localized perhaps along faults or the alteration is confined to the crust at depth beneath more recent lava flows.
The original magnetic field data from the 1973 survey were reprocessed by Wooldridge et al. [1992] and analysed using the inversion techniques of Parker and Huestis [1974] modified for three-dimensions by Macdonald et al. [1980]. In addition to TAG, Wooldridge et al., [1992] also investigated sea surface magnetic anomalies over several other known and suspected sites of hydrothermal activity on the MAR at 15°N and 17°N and also the Sea Cliff vent field on the Gorda Ridge at 42°45’N. They suggested that magnetic anomaly lows over these regions were primarily caused by regional-scale high-temperature hydrothermal alteration of the crust with subsidiary factors at each site varying from captured reversely magnetized crust, to thinner extrusive crust arising from magmatic cyclicity or due to the raised temperatures in the crust above an axial magma chamber. In the latter case, an increase in crustal temperature in a region above a magma chamber would cause the magnetic minerals to exceed their Curie temperature i.e. the temperature above which magnetism ceases to occur, which for the dominant magnetic mineral in basalt, titanomagnetite, is approx. 160-300°C [Wooldridge et al., 1992].

Sea surface magnetic surveys are necessarily limited in their resolution by the water depth (3-4 km), the distance from the source region. However, most of the discovered hydrothermal vent fields to date have spatial extents on the scale of 100s of meters [Hannington et al. 2005], which means that sea surface data could detect regional kilometer scale alteration but lacks the spatial resolution needed to resolve individual vent systems. Near-bottom surveys provide magnetic measurements on the same spatial scale as individual vent deposits and thus can tie the anomalous field more directly to the deposits. The first unambiguous documentation of the magnetic effects of a seafloor hydrothermal field in detail was obtained by the submersible ALVIN in a 1990 survey over the active TAG mound [Tivey et al., 1993]. A distinct zone of reduced magnetization was
computed from magnetic field data collected along submersible tracks at an altitude of 20 meters above the mound (Plate 2). Magnetic modeling of this zone of reduced magnetization suggested a pipe-like source body with a radius of 100 m, a depth extent of 500 m and a magnetization contrast of 12 A/m. Inversion results suggested a narrower zone some 80 m across (Plate 2). This narrow pipe-like body is directly comparable in size to the upflow zones and stockwork pipes typically found in terrestrial volcanogenic massive sulfide ore deposits such as the Cyprus ophiolite [Johnson et al. 1982; Richards et al., 1989]. To answer the question of whether the zone of reduced magnetization beneath the active TAG mound could be responsible for the more regional anomaly measured at the sea surface, a computation of the total magnetic moment suggested that this discrete zone of demagnetization could not produce the anomaly [Tivey et al., 1993]. The estimated total magnetic moment for the TAG active mound is $10^8$ Am$^2$; which is calculated from the computed magnetization contrast, the area of the low magnetization zone and the thickness of the source body. A similar calculation for the magnetic moment required to generate the sea surface anomaly is $5 \times 10^{10}$ Am$^2$, (see Tivey et al., 1993), which means that several hundreds of sources (500) similar to the TAG active mound would be needed to create the surface anomaly. Such sources clearly do not exist. Tivey et al., [1993] concluded that the regional zone of reduced magnetization measured at the sea surface was a reflection of crustal scale processes rather than an ensemble of individual vent deposit anomalies.

In 1994, new insight into the source of the regional magnetic anomaly at TAG came from the analysis of a deeptow sidescan and magnetic survey of the area [Kleinrock and Humphris, 1996] as part of a site survey effort prior to drilling by the Ocean Drilling Program (ODP) [Humphris et al. 1995; Herzig et al., 1998]. The deeptow sidescan and magnetic data were
collected at an altitude of 100 m, with line spacing of 700 m, and encompassed an area of 200 sq. km extending across the TAG segment spreading axis and 10 km from north to south. The magnetic data analysis revealed that the zone of reduced magnetization is clearly located over the eastern wall of the rift valley and is not directly associated with the active and inactive hydrothermal mounds (Plate 1c,d). Combined with the earlier discovery of gabbros on the eastern valley wall [Zonenshain et al., 1989], further analysis suggested that the regional magnetic low is the result of crustal thinning and the loss of the extrusive crustal layer caused by 4 km of horizontal extension on a low-angle normal fault [Tivey et al., 2003]. It was suggested that the location of the active TAG mounds and adjacent relict mounds on the hanging wall of this detachment fault was one reason why these systems had been active for such a long period of time [Lalou et al., 1993; 1998; You and Bickle, 1998]. Further confirmation of the geometry of the detachment faulting was provided by a seismic refraction survey and microearthquake study, which imaged the crustal structure and revealed the steep dip of the detachment fault at its toe [Canales et al., 2007; deMartin et al., 2007]. The recognition that low-angle detachment fault systems are much more prevalent than first thought in slow spreading environments [Smith et al., 2006; 2008] coupled with recent discoveries of additional active and inactive hydrothermal activity [Searle et al., 2007; Murton et al., 2007; Cannat et al., 2007; Fouquet et al. 2007] associated with these types of fault systems suggests that these environments are a major new addition to the inventory of hydrothermal activity at slow spreading ridges [Escartin et al., 2008].

ODP drilling at TAG (Leg 158) was intended to sample the subsurface structure of an active hydrothermal system and to investigate the mineral and host rock zonation and fluid-rock interactions [Herzig et al. 1998; Humphris et al., 1996]. Drilling conditions were difficult due to
the harsh conditions and core recovery was generally poor (av. 12%). However, a penetration of 125 m was accomplished and the samples helped reveal the processes active within the mound. The mound was found to consist of 4 different types of breccias: a massive pyrite breccia, an anhydrite-rich breccia, a quartz-sulfide breccia stockwork at depths > 40 mbsf and a quartz-chlorite breccia stockwork at depths >100 mbsf. The drill-hole results confirmed that the zone of upflow is narrowly constrained beneath the active mound and is likely to be on the order of 80 m in diameter, consistent with the submersible magnetic inversion results (Plate 2).

Rock magnetic measurements of the breccias reveal very weak magnetization with NRM values generally >> 1 A/m and low magnetic susceptibility [Zhao et al., 1998]. Magnetic tests of the sulfide samples suggest that pyrrhotite and trace amounts of magnetite are present and contribute to the magnetic properties of these samples. Holes TAG-2 and TAG-4 drilled the margins of the active mound and did intercept basalt and silicified basalt breccia at depth. Magnetic measurements of these samples show NRM values more typical for basalt (~10 A/m) and higher magnetic susceptibility. Magnetic tests suggest a pseudo-single domain grain size and high Curie temperatures of 500-540°C indicative of low-Ti titanomagnetite. Unoxidized titanomagnetite has a Curie temperature of 160°C, while completely oxidized titanomaghemite has a Curie temperature of 350°C [Ade-Hall et al., 1971]. Zhao et al. [1998] suggests the possibility of high-T deuteritic oxidation of these basalt samples. Perhaps the most surprising result, however, was the fact that the basalts also showed much shallower inclinations (20.8°) than predicted (55°) based on the site latitude [Zhao et al., 1998]. The explanation for this shallow inclination remains to be resolved. The inclination could be the result of tilting/rotation due to faulting, although no obvious signs of this are seen in other data. The magnetization could simply be remagnetized through chemical
alteration, although such remagnetization is unlikely to be in a direction different than the ambient field. Another speculation is that the basalts may have captured a geomagnetic excursion for which there are several candidates in the Brunhes recent past (e.g. Laschamps, Lake Mungo, etc) [Zhao et al., 1998]. Age dating and magnetic paleointensity measurements could help to test this possibility.

In summary, hydrothermal activity at the TAG active mound and nearby inactive mounds does result in localized demagnetized zones but these are not responsible for the observed sea surface anomaly. This regional sea surface anomaly is the result of crustal-scale processes related to a loss in magnetic source layer (extrusive lavas) by low-angle detachment faulting which appears to act as a mechanism to promote hydrothermal activity on the hanging wall of the fault.

3.2 Snake Pit Hydrothermal zone (23°22.08’N 44°57.00’W)

The Snake Pit hydrothermal area was discovered by deeptow camera in May 1985 during a site survey for bare-rock drilling sites on the MAR just south of the Kane fracture zone at 23°22N [Kong et al., 1985]. The site is located on the neovolcanic ridge within the axial rift valley, 25 km south of Kane fracture zone. The deeptow camera imaged pockets of greenish-white mottled sediment populated by crabs and spindly worms subsequently found to be shrimp and eels (after which the vent site was named). The vent area was explored and drilled as part of ODP Leg 106, site 649 [Detrick et al., 1986]. The pre-drill camera survey found an 11 m tall black smoker chimney and an extensive area of mineralization and hydrothermal chimney debris. A transect of holes were drilled with very limited core recovery; only one hole, site 649B recovered 6 m of ore minerals before intercepting hard substrate. Only a small amount of basaltic material (chips) was recovered from the drilling. This basalt was mostly glassy and aphyric and apparently unaffected
by hydrothermal alteration. The basalt also had a relatively high MgO content, which indicates slightly more primitive lava than the basalt recovered to the north at site 648 (Serocki volcano)

[Detrick et al., 1986]. The area was subsequently visited by submersibles Alvin [Karson and Brown, 1988], Nautile [Mevel et al., 1989; Dubois et al., 1994] and Mir [Lisitsyn et al., 1989]. The pre-ODP drilling surveys included sea surface magnetic surveys [Schulz et al., 1988], which showed no pronounced magnetic anomaly low as was noted over the TAG area [Wooldridge et al., 1992; Tivey et al., 1993]. Inspection of the magnetic inversion for crustal magnetization [Schulz et al., 1988, Fig.9] shows that ODP site 649 at the Snake Pit vent site is located between two along-axis magnetization highs, but as discussed earlier, it is unlikely that the magnetic effects of hydrothermal alteration related to individual vent fields can be resolved by the sea surface magnetic data. To date, no near-bottom magnetic data have been collected over the Snake Pit area.

3.3 Lucky Strike (37°17’N 32°17’W)

Following a series of research cruises to survey the MAR axis in the early 1990’s under the FARA program (French-American cooperative agreement, Needham et al., 1992), a detailed sampling program (FAZAR) collecting rock samples every 5 km along the neovolcanic zone between 34° and 41°N recovered one dredge at 37°17’N with sulfides and hydrothermal vent organisms [Langmuir et al., 1997]. This serendipitous event thus coined the name of this area as “Lucky Strike”. The Lucky Strike ridge segment located between 37° and 37°35’N is the third segment south of the Azores platform and represents a hot-spot influenced segment that provides an opportunity to document the potential impact of hot-spot magmatism on a hydrothermal system. The ridge segment is 65 km long and marked by a rift valley that has remarkably uniform width of 11 km wide and a prominent axial volcano 430 m high reaching 1550 m at its summit located at the
center of the segment (Plate 3). The summit of the axial volcano can be divided into two halves, a
linear narrow ridge to the west and a more circular edifice to the east that is marked by 3 small
cones which enclose a lava lake at the summit [Fouquet et al., 1995]. Hydrothermal activity is
located on the periphery of this lava lake. Several submersible dive programs documented the
presence of high temperature black smoker chimneys, extensive areas of diffuse flow and sulfide
deposits distributed around the lava lake margins [Fouquet et al., 1994a, b; Langmuir et al., 1997;
Ondreas et al., 2009]. The presence of a lava lake at the summit also suggests recent magmatic
activity and the potential for an active magma chamber directly beneath the edifice. Confirmation
of the presence of an active axial crustal magma chamber at a depth of 3 km was obtained by a
seismic reflection survey over the Lucky Strike volcano [Singh et al., 2006]. Ongoing studies
including a recent on-bottom gravity study have been completed across the area to detect the
presence of this magma chamber as part of the MOMAR observatory effort.

Publicly available sea surface magnetic data (National Geophysical Data Center (NGDC):
www.ngdc.noaa.gov) were compiled over the Lucky Strike area. Inversion for crustal
magnetization was performed using the Parker and Huestis [1974] approach assuming a source
layer thickness of 1 km, a geocentric dipole direction for the magnetization and a pass band with
wavelength cutoffs of 237 and 2 km respectively. The magnetization inversion shows a slight
weakening of the axial magnetization signal over the center of the segment compared with the
segment ends (Plate 3b), which is typical of MAR segments in general [Grindlay et al., 1992;
Ravilly et al., 1998; Tivey and Tucholke 1998;] and has been explained in various ways including
changes in along-axis crustal thickness and composition, in the depth of the Curie isotherm, in the
basalt petrology, and in the degree of alteration. In the case of the Lucky Strike segment, the
presence of a central volcano argues against thinned crust but might support thermal
demagnetization due to an elevated Curie isotherm. The inversion and analysis of a dense sea
surface magnetic anomaly compilation has revealed a strong, ~7 nT magnetization low over the
Lucky Strike lava lake and hydrothermal area [Miranda et al., 2005], suggesting that the
hydrothermal systems located on the summit of the volcano may have sufficient volume to be
imaged magnetically from the sea surface. A detailed near-bottom sidescan sonar, phase
bathymetry and magnetic survey was completed over the axial volcano in 1996 at an altitude of 100
m [Fornari et al., 1996; Scheirer et al., 2000; Humphris et al., 2002]. The near-bottom magnetic
data were upward continued to a level plane, bordered with the regional data and inverted for
crustal magnetization assuming a source layer thickness of 1 km. A geocentric dipole direction for
the magnetization was assumed and the inversion solution was filtered with a wavelength passband
of 2 and 237 km to ensure convergence. The magnetization inversion results, shown in Plate 3d,
shows two distinct zones of reduced magnetization. A 1.5 x 3 km zone of low magnetization is
located over the easternmost cone of the summit and a longer, linear magnetization low is found to
the west of the summit (Plate 3d). These zones appear to align with the crustal fault systems
imaged by the reflection seismic data (Plate 3) that bound the summit area [Singh et al., 2006].
These bounding faults are interpreted by Singh et al. [2006] to be pathways of hydrothermal
recharge for the axial hydrothermal system beneath the summit and thus may have suffered
enhanced alteration resulting in the linear zones of reduced magnetization on a crustal scale (Plate
3d). Confirmation of this crustal alteration within these fault zones remains to be confirmed by in
situ observations and rock sample recovery and analysis. More detailed surveys are ongoing at
Lucky Strike as part of the MOMAR observatory effort and high resolution camera-tow magnetic
data were collected as part of this effort but remain to be analyzed [Escartin et al., 2006].
3.4 Menez Gwen (37°50’N 31°30’W)

The Menez Gwen hydrothermal field was discovered during submersible dives on the ridge segment between 37°35’N and 38°N just north of the Lucky Strike segment [Fouquet et al., 1994a,b]. This segment also has a well-developed circular volcano similar to Lucky Strike, which is centrally located within the rift valley (Plate 4). The 800 m high volcano is 16 km in diameter and has a 2 km wide by 6 km long axial graben at its summit, which is ~300 m deep, open to the north and south and floored by very fresh volcanics [Ondreas et al., 1997]. A younger 120 m high, 700 m diameter volcanic cone is located towards the northern end of the graben. The vent field was discovered on the east and southeast slopes of this volcanic cone at depths of 840 to 865 m and hosted in young fresh pillow lava [Desbruyeres et al., 2001]. In terms of magnetism, the area has not been mapped in detail by near-bottom sensors but sea surface magnetic data do exist. An inversion of sea surface magnetic field data compiled from NGDC (using the Parker and Huestis inversion approach [1974] and assuming a 1 km thick source layer and geocentric dipole direction for the magnetization) shows that the central portion of the Menez Gwen ridge segment dominated by the large axial volcano does have a reduced magnetization zone (Plate 4). This reduced zone of magnetization is similar to other ridge segments of the MAR, which show similar reduced magnetization at segment centers and enhanced magnetization at segment ends [Grindlay et al., 1992; Ravilly et al., 1998; Tivey and Tucholke, 1998]. Nevertheless, the juxtaposition of reduced magnetization at the magmatically robust crustal center certainly suggests that processes other than crustal thickness are acting to reduce the magnetic effect of the crust (Plate 4).

3.5 South Atlantic Vent Sites
Over the past few years, MOR exploration has also begun to include the MAR south of the large-offset Romanche fracture zone in the south Atlantic to test ideas of biogeographic barriers to vent faunal populations. Plume mapping [German et al., 2002, 2005; Devey et al., 2005] and in situ mapping and sampling [Koschinsky et al., 2006; Haase et al., 2007; German et al., 2008] has now revealed the presence of several active hydrothermal sites in the South Atlantic between 2° and 14°S. Many of these sites appear to be hosted on basaltic substrate. At 4°48’S, there are several active black smoker fields (Comfortless Cove, Turtle Pits and Red Lion) at depths ranging from 3050 m at Red Lion field to 2990 m at Turtle Pits. The Turtle Pits sites are boiling and the measured vent exit temperatures reach 407°C [Haase et al., 2007; Koschinsky et al., 2007]. These are located within the axial rift on very fresh volcanic lava, which may have recently erupted. The vent fluid chemistry has high hydrogen content relative to methane along with significantly reduced chloride concentrations indicative of phase separation presumably related to a very recent intrusion and lava eruption [Koschinsky et al., 2005; 2008]. Interestingly, the inactive chimneys at Turtle Pits have abundant hematite in association with pyrite and magnetite [Haase et al., 2007] and the presence of such oxidizing conditions may suggest that positive magnetic anomalies might be associated with these deposits. Only sparse sea surface magnetic coverage is available on this portion of the MAR [Bruguier et al., 2003] and the low latitude is not favorable for collecting high quality magnetic data. Near-bottom ABE magnetic data were collected over the 4°48’S area in 2005 (Plate 5), flown at an altitude of approx. 50 m with a nominal 40-50 m line spacing [German et al., 2008; Haase et al., 2007]. The primary magnetic sensor of ABE was not functioning for this survey and so we have processed the back-up ABE sensor (Crossbow) which results in a noisier but satisfactory result. Inversion analysis of these ABE magnetic data reveal zones of discrete reduced magnetization associated with the Red Lion and Comfortless Cove vents areas (Plate 5d). The
Turtle Pits site also lies on the edge of a large ~200 m diameter zone of reduced magnetization (Plate 5d) perhaps suggesting a wider area of alteration buried under the recently erupted lava flows [Haase et al., 2007].

Further south, the Nibelungen vent field (8°17’S) was discovered at 2905 m depth [Koschinsky et al., 2005; 2006; German et al., 2008; Yoerger et al., 2007]. This vent field is located on the face of a steep fault facing outward from the rift axis. Although the rocks adjacent to the vent site appear to be dominantly volcanic pillow lava, the vent fluid chemistry suggests a serpentinite influence. Finally, a small low temperature vent field (max. temperature 17°C), the Lilliput Field, was discovered at 9°33’S at 1500 m on the summit of an axial volcanic ridge [Haase et al., 2005; Koschinsky et al., 2006; Yoerger et al., 2007]. Near-bottom magnetic data over these areas remain to be analyzed.

3.6 Ultramafic hosted vent deposits

Rocks from the mantle and the deep crust are observed on transform fault and fracture zone scarps in different spreading environments, from fast (e.g. Garrett Fracture Zone or FZ) to slow spreading centers (e.g. Marie-Celeste FZ; Kane FZ; Saint Paul FZ). However, only slow and ultraslow spreading centers expose such deep seated rocks at the spreading axis, both as a result of low magma supply [e.g. Cannat, 1993; Dick et al., 2003; Cannat et al., 2006] and tectonic denudation [e.g. Tucholke et al., 1998; Smith et al., 2008]. Although outcropping mantle rocks likely reflect a colder thermal regime, several active hydrothermal sites have nevertheless been discovered in ultramafic environments on the Mid-Atlantic Ridge. These sites exhibit a rather large diversity, including low temperature (7-9°C) fluids and diffuse flow at Saldanha (see Table-1);
focused low temperature (40-75°C) fluids and carbonate chimneys at Lost City and focused high
temperature fluids and sulfide precipitates at Rainbow (365°C) and Logatchev (350°C). In addition
to the northern MAR, evidence for ultramafic-hosted hydrothermal vents have also been found in at
least three locations on the Southwest Indian Ridge [Tao et al., 2007].

To date only the Rainbow area has been surveyed for magnetism in any detail, i.e. both sea-
surface and deep-sea magnetic surveys. Rainbow is located at 36°14’N, 33°54’W, on the western
side of a N-S hill, 13 km long and 8 km wide, which separates two adjacent segments (Plate 6a).
This hill is mostly comprised of serpentinized peridotite and may represent an oceanic core
complex at the footwall of a detachment fault, although no clear striations or mullions have been
identified so far. The hydrothermal site extends over 300 m in a E-W direction and 200 m in a N-S
direction, at a depth of about 2300 m (Plate 7a). It was first explored in 1997 [Fouquet et al., 1997]
shortly after the associated hydrothermal plume was detected [German et al., 1996; Charlou e
1996]. This plume is one of the strongest in the Atlantic Ocean as far as CH₄, H₂ and Fe contents
are concerned [Charlou et al., 2002], suggesting fluid-mantle rock interaction and serpentinization.
Rainbow vent fluids appear to be very reducing, highly acidic (pH 2.8) and very high in Fe
[Douville et al., 2002].

Plate 6b shows the sea-surface scalar magnetic anomaly compiled from the available
shipborne data, and Plate 6c shows the equivalent magnetization derived from this anomaly
assuming a 500 m thick magnetic layer whose top is bounded by the topography. The magnetic
lineations show a local N20°E direction and a general N40°E direction, in agreement with,
respectively, the local structural direction of each ridge segment and the overall direction of the
Mid-Atlantic Ridge in the area, as mirrored in the topography (Plate 6a). The axial magnetic anomaly (the normal Brunhes period, 0-0.78 Ma) is marked by the highest amplitudes, up to 40 A/m, crossing the map on the NE-SW diagonal. The reversed Matuyama period corresponds to the negative equivalent magnetization of -20 A/m that dominates the SE and NW corners and only interrupted by Chron 2 (the normal polarity Olduvai interval) marked by a positive magnetization up to 20 A/m. The axial magnetic anomaly displays significant along-axis variations. The AMAR segment (north of Rainbow) shows a strong equivalent magnetization (35 A/m) at the segment southern end near 35°17’N and a much weaker one (5 to 10 A/m) at the segment centre near 35°22’N, in agreement with the general observation of Ravilly et al. [1998]. Conversely, the segment south of Rainbow shows a strong equivalent magnetization all along the segment, maybe because this segment is too short to exhibit the characteristic signature of a typical MAR segment. The Rainbow vent site is located between the two segments within the inside corner of a non-transform offset in an area which exhibits a lower equivalent magnetization of 15 A/m (Plate 6c). Such a value may however not be meaningful, as the computation of an equivalent magnetization assumes a constant thickness of the magnetic layer – an assumption which is probably violated at segment ends.

A deep-sea magnetic survey was carried out at Rainbow in 2001 by ROV Victor of IFREMER using the deep-sea vector magnetometer developed by the Ocean Research Institute of the University of Tokyo. The vent site was systematically surveyed at an altitude of about 10 m, with profiles spaced by about 60 m. No high-resolution bathymetric data was collected, as this capability was yet not available on ROV Victor. The magnetic data were corrected for the magnetic effect of the ROV and inverted to magnetization using the topography and the altitude of
the submersible [Honsho, 1999, Nakase, 2003; Kitazawa, 2006; Honsho et al., 2009]. The resulting map (Plate 7b) shows a very strong positive magnetization (up to 35 A/m) associated to the western side of site Rainbow, whereas it does not exceed 10 A/m in other areas. Seafloor submersible observations suggest that the western part of Rainbow consists of well-developed and relatively mature vents edifices, whereas the three easternmost vents (Plate 7a) may have appeared more recently and may still be immature [Y. Fouquet, pers. comm., 2004].

The host rocks of the Rainbow region are comprised of a non-mineralized, lizardite-dominated serpentinite resulting from a static, relatively low-temperature, retrograde serpentinization [Marques et al., 2006]. The sulfide-bearing serpentinites, on the other hand, are characterized by a stockwork zone of semi-massive sulfides (Cu-Zn-(Co) type) that are found restricted to the vent area and is not thought to be genetically linked to the earlier serpentinization history [Marques et al., 2007]. This overprinting of sulfide mineralization results in zonation and replacement textures more typically found in mafic systems [Doyle and Allen, 2003], but with some ultramafic tendencies including Ni-rich stockwork sulfides and methane anomalies [Marques et al., 2007]. In terms of the source of the strong magnetization found over the main Rainbow vent area, it would appear that the main serpentinite host rock is only weakly magnetic so that the formation of magnetite by the serpentinization process may not be the main source of magnetization. At the vent site, the serpentinized peridotite would therefore be overprinted by the stockwork sulphide mineralization inferring that the source of the magnetization anomaly could be due to the presence of magnetic iron sulfide such as pyrrhotite or even nickel, if present in sufficient quantity. Rock magnetic measurements of the sulfide phases are needed to test this assertion.
In summary, the Rainbow hydrothermal site and the serpentinized peridotite hill upon which it stands is associated with relatively low equivalent magnetization inverted from the sea-surface magnetic anomalies, between two magnetic highs associated with the nearby ridge segment ends (Plate 6d). Conversely, the vent field itself corresponds to a very strong magnetic signal at its western edge as inverted from deep sea vector magnetic anomalies. These observations can be reconciled by recognizing that both data sets do not reflect the same scale of features of the seafloor. Deep-sea magnetic anomalies recorded at 1 – 10 m above seafloor are much more sensitive to shallow sources (i.e. in the uppermost tens to hundreds of meters) which dominate the signal, whereas sea surface magnetic anomalies, recorded 2500 m above the seafloor, integrate the effect of the magnetization carried by all layers of the oceanic crust. The “crust”, in this case serpentinized peridotite, around the Rainbow vent site has a relatively low magnetization, upon which is superimposed a smaller but more strongly magnetized shallow body. A likely magnetic source for this superficial body is magnetic sulfide mineralization possibly a stockwork zone of pyrrhotite or perhaps more exotic minerals of Ni composition.

The high-temperature ultramafic-hosted Logatchev and Achadze vent sites have also recently been the target of deep-sea high resolution bathymetric and magnetic measurements [Fouquet et al., 2007; Ondreas et al., 2007]. These results are currently being analyzed but are required to generalize the observations gathered at Rainbow to the focused, high-temperature hydrothermal vents in an ultramafic environment. Furthermore, a very high resolution photographic, bathymetric and magnetic study carried out at Rainbow in mid-2008 will help to better constrain the source of the strong Rainbow magnetic anomaly. The magnetic signature of other low temperature ultramafic sites such as Saldanha (diffuse, very low temperature flow) and
Lost City (focused, low temperature vents) remains relatively unknown at present. Rock magnetic measurements have been made on basaltic samples from the Saldanha area showing typical low-temperature alteration but no measurement of the magnetic properties was carried out on ultramafic rocks from the hydrothermal area [Miranda et al., 2002]. A deepwater magnetic anomaly profile across the Lost City vent field showed a pronounced low suggesting that serpentinization is not contributing to the magnetic signal [Gee and Blackman, 2004]. Finally, at least three hydrothermal plumes have recently been detected by Chinese investigators along the ultraslow Southwest Indian Ridge [Tao et al., 2007], an area which displays many serpentinite outcrops [Cannat et al., 2006]. Given these recent results, it is likely that the number of hydrothermal sites located on serpentinite outcrops are more numerous than first thought. Indeed, it is also likely that slow spreading ridges in general show a more diverse range of hydrothermal systems than might have been predicted only 5 years ago (e.g. Escartín et al., 2008).

Conclusions

We can make the following generalizations and conclusions concerning the magnetic response of hydrothermal systems in the slow spreading midocean ridge environment.

1) In basaltic-hosted hydrothermal vent systems, crustal magnetization is strongly impacted by hydrothermal fluid circulation, particularly in the upflow zones where buoyant high-temperature fluid rises to the seafloor through the crust. The lateral scale of these upflow zones appears to be on the order of 100 m or less compatible with observations of ancient systems found in ophiolites [e.g. Richards et al., 1989]. The depth of these upflow zones remains to be quantified but they are likely to extend through the extrusive crust into the intrusive dike section. Further, also by analogy
to ancient deposits, the lateral alteration gradients in the upflow zone are likely to be quite sharp resulting in well-defined magnetic contrasts [e.g. Tivey and Johnson 2002].

2) From the few high-temperature, focused-flow ultramafic-hosted systems surveyed to date (Rainbow), we find strong positive magnetic anomalies possibly associated with magnetic sulfide deposition or perhaps other metallic mineral compounds that are magnetic such as Ni (or Cr-Fe). If confirmed, the role of magnetite formation as a result of serpentinization of the host peridotite may be only of secondary effect in producing magnetic anomalies. More work is needed to fully characterize this conclusion.

3) Little to no magnetic work has been done over low-temperature ultramafic hosted vent systems like Saldanha or Lost City. Preliminary indications from a near-bottom magnetic profile over Lost City are that while serpentinization is occurring at depth, it is not producing sufficient magnetite to create a positive anomaly at the vent site [Gee and Blackman, 2004]. More work is also needed here to fully quantify the magnetic effects in these low temperature, low flow ultramafic hosted systems.

4) The slow spreading environment as exemplified by the Mid-Atlantic Ridge shows a remarkable diversity of hydrothermal systems that was perhaps unanticipated a few years ago. The diversity of host rocks found in slow spreading environments also allows for a greater range in the types of mineralization found at slow spreading ridges. Also, the role of low-angle faulting and detachment tectonics provides a newly recognized pathway for fluid flow that will allow for hydrothermal systems to maintain activity away from the conventional spreading axis environment. It was
already known that the slow spreading environment hosts longer lived hydrothermal systems than fast to intermediate spreading centers where volcanic activity may bury vent systems on a regular basis. This proclivity to longer lived activity would potentially lead to larger hydrothermal deposits and perhaps a greater chance of preservation in the off-axis realm.

5) Magnetic imaging provides a viable method to search for these hydrothermal deposits and with advances in gradiometer and vector measurement technology, which would allow for source depth estimation and improved lateral interpolation of spatial anomaly character, we foresee better quantification of these anomalous magnetic zones in the future.

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**Figure Caption 1.** Location map of known hydrothermal sites on the Mid-Atlantic Ridge between 20°S and 60°N.

**Plate 1.** Magnetic data over the TAG hydrothermal field on the Mid-Atlantic Ridge (From Tivey et al., 2003). a) Multibeam bathymetry showing the location of the active TAG mound (star). Bold box shows the outline of the deeptow sidescan and magnetic survey [Kleinrock and Humphris, 1996]. b) Crustal magnetization computed from sea surface magnetic field data showing a magnetization low located within the Brunhes normal chron. Star is the TAG active mound. c) High-resolution phase bathymetry map from the deeptow survey of Kleinrock and Humphris, [1996]. White stars represent from left to right, TAG, Alvin zone and Mir mound. d) Crustal magnetization computed from near-bottom magnetic data [Tivey et al., 2003] showing the zone of reduced magnetization over the eastern wall of the spreading segment indicating exposed less magnetic lower crustal rocks. White stars as in c).

**Plate 2.** Detailed bathymetry and magnetic data over the TAG active hydrothermal mound. Top panel shows high resolution multibeam bathymetry of the active TAG mound obtained by ROV Jason, contour interval is 5 m [Roman et al., 2007]. Bottom panel shows high-resolution near bottom magnetic data collected by submersible Alvin over the TAG active mound showing reduced magnetization directly over the mound (cont. int. 1 A/m) [Tivey et al., 1993]. The north-south extension of the reduced zone suggests a possible fault trend influence in the upflow zone. Bold line is the 3645 m isobath delineating the mound extent. Figure modified from Tivey et al., [1993].
**Plate 3.** Bathymetry and magnetic data over the Lucky Strike hydrothermal area. a) Ship multibeam bathymetry of the Lucky Strike segment showing the axial volcano within the rift valley that hosts the hydrothermal activity. b) Crustal magnetization computed from the sea surface magnetic data (see text for details) showing the Brunhes anomaly and the weaker crustal magnetization over the summit of Lucky Strike volcano. Bold line is the 1900 m isobath delineating the volcano. c) High-resolution phase bathymetry of the Lucky Strike volcano collected by a deeptow DSL120 survey [Scheirer et al., 2000]. Red stars show the location of hydrothermal activity on the summit of the volcano. d) Crustal magnetization computed from the DSL120 near bottom magnetic data (see text for details) showing magnetization lows associated with the major fault systems (shown in heavy bold dash) imaged by multichannel seismic data [Singh et al., 2006]. Red star indicates the summit hydrothermal vent area. These major fault systems may be hosting recharge for the hydrothermal system [Singh et al., 2006] resulting in enhanced low temperature alteration on a regional crustal scale.

**Plate 4.** Magnetic and bathymetry data of the Menez Gwen ridge segment: a) regional multibeam bathymetry data from [Lourenço et al. 1998] showing the presence of the axial split volcano within the rift valley reaching 800 m depth. Bold black line marks the spreading axis. b) A compilation of sea surface magnetic field data over the Menez Gwen segment (data from National Geophysical Data Center). c) Multibeam bathymetry [Cannat et al., 1999; Escartin et al., 2001]. Red star shows the location of the hydrothermal vent site within the central rift of the split axial volcano. d) Crustal magnetization computed from the sea surface magnetic data and showing the overall...
reduced magnetization over the axial volcano. Bold line is the 1000 meter isobath outlining the axial volcano. Red star is the location of the active vent site.

Plate 5. Magnetic and bathymetry data from the 4°48’S region on the southern Mid-Atlantic Ridge:

- a) Regional ship-based multibeam bathymetry courtesy of German et al., [2008].
- b) High resolution near bottom multibeam bathymetry data from 2005 autonomous vehicle ABE missions 151/153.
- c) ABE multibeam data showing the location and names of active vent sites with red stars.
- d) ABE near-bottom magnetic data inverted for relative crustal magnetization showing zones of reduced magnetization associated with the active vent sites show by red stars.

Plate 6. Sea surface magnetic and bathymetry data over the Rainbow hydrothermal area of the Mid-Atlantic Ridge at 36°14’N.

- a) Multibeam bathymetry from [Thibaud et al., 1998] showing the location of Rainbow ridge at the non-transform offset between the AMAR segment to the north and the South AMAR segment to the south.
- b) Sea surface magnetic data compilation of the Rainbow area (data from National Geophysical Data Center and cruise Iris, chief scientist Y. Fouquet) with ship tracks shown by fine black lines.
- c) Regional crustal magnetization computed from the sea surface magnetic field data showing the central Brunhes anomaly and the weak magnetic field over the Rainbow area.


Individual vents are shown by the triangles and circles. Near bottom magnetic survey tracks shown
dotted. b) Crustal inversion of the near bottom magnetic data showing that a strong positive anomaly is located over the main active vent field to the west end of the field.
Table 1 Hydrothermal vent sites along the Mid-Atlantic Ridge

<table>
<thead>
<tr>
<th>Vent Field</th>
<th>Location</th>
<th>Active (A)</th>
<th>Inact. (I)</th>
<th>Type $^1$</th>
<th>Terrain $^2$</th>
<th>Depth (m)</th>
<th>Areal Size $^3$ (sq. km (m))</th>
<th>Reference</th>
</tr>
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<td>Loki’s Castle</td>
<td>~73.5°N, 4°W</td>
<td>A</td>
<td>A</td>
<td>High-T 300°C</td>
<td>Volcanic</td>
<td>2352</td>
<td>0.01 (100 x 100)</td>
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<td>Gallionella</td>
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<td>A</td>
<td>High-T 270°C</td>
<td>Volcanic</td>
<td>550</td>
<td>(500 x ?)</td>
<td>Pederson et al., 2005</td>
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<td>A</td>
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<td>Volcanic</td>
<td>700</td>
<td>(100 x 100)</td>
<td>Pederson et al., 2005</td>
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<td>A</td>
<td>A</td>
<td>Low-T 131°C</td>
<td>Volcanic</td>
<td>100-110</td>
<td>unreported</td>
<td>Scholten et al., 1999, Hannington et al., 2001</td>
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<td>Grimsey</td>
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<td>A</td>
<td>High-T 248-</td>
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<td>400</td>
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<td>Scholten et al., 1999, Hannington et al., 2001</td>
</tr>
<tr>
<td>Location</td>
<td>Latitude/Longitude</td>
<td>Activity</td>
<td>Temperature</td>
<td>Type</td>
<td>Mass Flow Rate</td>
<td>Reference</td>
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<td>Volcanic</td>
<td>220-350</td>
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<td>Steinaholl</td>
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<td>unknown</td>
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<td>265-285°C</td>
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<td>Lucky Strike</td>
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<td>Volcanic</td>
<td>1700</td>
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<td>170-324°C</td>
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<td>Ultramafic</td>
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<td>356-</td>
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<td>(7850 sq. m)</td>
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<td></td>
<td>Murton et al., 1994</td>
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<td></td>
<td>German et al, 1999</td>
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<tr>
<td>Location</td>
<td>Lat/Long</td>
<td>Type</td>
<td>Temperature</td>
<td>Depth</td>
<td>Width</td>
<td>References</td>
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<tr>
<td>TAG</td>
<td>26.14°N, 44.83°W</td>
<td>A</td>
<td>High-T</td>
<td>3620-3700</td>
<td>25 (5000 x 5000)</td>
<td><em>Rona et al., 1993</em></td>
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<td>MAR 24°30’N</td>
<td>24.5°N, 46.15°W</td>
<td>I</td>
<td>n/a</td>
<td>3900</td>
<td>Unknown</td>
<td><em>Sudarikov et al., 1990</em></td>
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<td>Snake Pit</td>
<td>23.37°N, 44.95°W</td>
<td>A</td>
<td>High-T</td>
<td>3440</td>
<td>0.045 (150 x 300)</td>
<td><em>Thompson et al., 1988</em>; <em>Fouquet et al., 1993</em></td>
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<td>Puy des Folles</td>
<td>20.50°N, 45.64°W</td>
<td>I</td>
<td>n/a</td>
<td>1940-2000</td>
<td>0.858</td>
<td><em>Gente et al., 1996</em></td>
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<td>Zenith Victory</td>
<td>20.13°N, 45.65°W</td>
<td>I</td>
<td>n/a</td>
<td>2370-2390</td>
<td>0.495</td>
<td><em>G. Cherkashov, pers. comm. 2009</em></td>
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<td>Krasnov</td>
<td>16.63°N, 46.47°N</td>
<td>I</td>
<td>n/a</td>
<td>3700-3750</td>
<td>0.15 (500 x 300)</td>
<td><em>Beltenev et al., 2004</em></td>
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<td>MAR 15°20’N</td>
<td>15.08°N, 44.8°W</td>
<td>A</td>
<td>Low-T</td>
<td>2452-4668</td>
<td>Unknown</td>
<td><em>Charlou et al., 1998</em></td>
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<tr>
<td>Logatchev-1</td>
<td>14.75°N, 44.98°W</td>
<td>A</td>
<td>High-T</td>
<td>2900-3050</td>
<td>0.032 (400 x 80)</td>
<td><em>Gebruk et al., 1997</em></td>
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<td>Latitude, Longitude</td>
<td>Type</td>
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<td>Depth</td>
<td>Temperature</td>
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<td>Logatchev-2</td>
<td>14.72°N, 44.94°W</td>
<td>A</td>
<td>High-T</td>
<td>320°C</td>
<td>2650</td>
<td>Cherkashev et al., 2000</td>
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<td>Logatchev-5</td>
<td>14.75°N, 44.97°W</td>
<td>I</td>
<td>n/a</td>
<td>Ultramafic</td>
<td>2900</td>
<td>unreported</td>
<td>Fouquet et al., 2008</td>
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<td>Semyonov</td>
<td>13.52°N, 44.95°W</td>
<td>I</td>
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<td>Volcanic</td>
<td>2400-2600</td>
<td>G. Cherkashov, pers. comm. 2009</td>
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<td>MAR 13°30’N</td>
<td>13.50°N, 44.83°W</td>
<td>I</td>
<td>n/a</td>
<td>unreported</td>
<td>unreported</td>
<td>unreported</td>
<td>Searle et al., 2007</td>
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<td>MAR 13°20’N</td>
<td>13.33°N, 44.83°W</td>
<td>I</td>
<td>n/a</td>
<td>unreported</td>
<td>unreported</td>
<td>unreported</td>
<td>Murton et al., 2007</td>
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<td>Ashadze 1</td>
<td>12.97°N, 44.86°W</td>
<td>A</td>
<td>High-T</td>
<td>370°C</td>
<td>4080</td>
<td>Beltenev et al., 2003</td>
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<td>Ashadze-2,3</td>
<td>12.97°N, 44.86°W</td>
<td>A</td>
<td>High-T</td>
<td>Ultramafic</td>
<td>3300</td>
<td>0.106</td>
<td>Fouquet et al., 2007</td>
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<td>Ashadze-4</td>
<td>~12.97°N, ~44.86°W</td>
<td>I</td>
<td>n/a</td>
<td>Volcanic</td>
<td>4530</td>
<td>unreported</td>
<td>Fouquet et al., 2007</td>
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<td>Comfortless Cove</td>
<td>4.80°S, 12.37°W</td>
<td>A</td>
<td>High-T</td>
<td>400-464°C</td>
<td>2996</td>
<td>0.0001-0.0002</td>
<td>German et al., 2005, Haase et al., 2005</td>
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<td>Turtle Pits</td>
<td>4.81°S, 12.37°W</td>
<td>A</td>
<td>High-T</td>
<td>Volcanic</td>
<td>2990</td>
<td>0.0004</td>
<td>German et al., 2005</td>
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<td>Latitude, Longitude</td>
<td>Terrain</td>
<td>Exit Temperature</td>
<td>Deposit Type</td>
<td>Areal Extent</td>
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<td>Red Lion</td>
<td>4.80°S, 12.38°W</td>
<td>A</td>
<td>High-T 193-349°C</td>
<td>Volcanic</td>
<td>3050</td>
<td>German et al., 2005 Haase et al., 2005 Koschinsky et al., 2008</td>
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<tr>
<td>Wideawake</td>
<td>4.81°S, 12.37°W</td>
<td>A</td>
<td>Diffuse</td>
<td>Volcanic</td>
<td>&lt;2990</td>
<td>German et al., 2005 Haase et al., 2005 Koschinsky et al., 2008</td>
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<td>Nibelungen</td>
<td>8.30°S, 13.51°W</td>
<td>A</td>
<td>High-T 153°C</td>
<td>Volcanic?</td>
<td>2915</td>
<td>Koschinsky et al., 2006 Melchert et al., 2008</td>
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<td>(Drachenschlund)</td>
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<tr>
<td>Lilliput</td>
<td>9.55°S, 13.21°W</td>
<td>A</td>
<td>Low-T Diffuse</td>
<td>Volcanic</td>
<td>1500</td>
<td>Koschinsky et al., 2006</td>
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</table>

1 High (T > 150°C) vs. low (T < 50°C) exit-fluid temperature.

2 terrain refers to the primary substrate hosting the vent site Volcanic

3 Reported areal extent of hydrothermal deposits in sq. kilometers and equivalent area in meters in parentheses.
Figure 1.
Plate 5.