Diversity of Hydrothermal Systems on Slow Spreading Ocean Ridges

Peter A. Rona, Colin W. Devey,
Jérôme Dyment, and Bramley J. Murton
Editors

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American Geophysical Union
Washington, DC

Library of Congress Cataloging-in-Publication Data

Diversity of hydrothermal systems on slow spreading ocean ridges / Peter A. Rona ... [et al.], editors.
p. cm. -- (Geophysical monograph, ISSN 0065-8448 ; 188)
Includes bibliographical references and index.
ISBN 978-0-87590-478-8 (alk. paper)
1. Sea-floor spreading. 2. Hydrothermal deposits. 3. Chemical oceanography. 4. Hydrothermal vents. I. Rona, Peter A.
QE511.7.D58 2010
551.2'309162--dc22
2010023962

ISBN: 978-0-87590-478-8
ISSN: 0065-8448

Cover Photo: Vent shrimp *Rimicaris exoculata* Williams and Rona, 1986, swarming on an active black smoker chimney in the TAG hydrothermal field on the Mid-Atlantic Ridge near latitude 26°N. Photo credit: IMAX film *Volcanoes of the Deep Sea*, produced by The Stephen Low Company. Reproduced with permission.

Printed in the United States of America.

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PREFACE

Presentations at three sessions entitled "Diversity of Hydrothermal Systems at Slow-Spreading Ocean Ridges" convened by P. A. Rona, R. Reves-Sohn, B. J. Murton, J. Dyment, and C. W. Devey at the American Geophysical Union (AGU) 2005 Fall Meeting initiated preparation of this volume. We thank R. Reves-Sohn for his strong constructive role in this process. C. R. German provided highly valued guidance for preparation of the volume. The authors are the heroes of this volume for advancing and communicating knowledge of seafloor hydrothermal systems. We are grateful to the reviewers who did yeoman's duty in advising the authors on how to improve their manuscripts.

This volume is part of the progression of knowledge of seafloor hydrothermal systems and their geologic settings represented by a series of AGU volumes: *Seafloor Hydrothermal Systems: Physical, Chemical, Biological, and Geological Interactions* (edited by S. E. Humphris, R. A. Zierenberg, L. S. Mullineaux, and R. E. Thomson, 1995); *Faulting and Magmatism at Mid-Ocean Ridges* (edited by W. R. Buck, P. T. Delaney, J. A. Karson, and Y. Lagabrielle, 1998); *The Subseafloor Biosphere at Mid-Ocean Ridges* (edited by W. S. D. Wilcock, E. F. DeLong, D. S. Kelley, J. A. Baross, and S. C. Cary, 2004); *Mid-Ocean Ridges: Hydrothermal Interactions Between the Lithosphere and Oceans* (edited by C. R. German, J. Lin, and L. M. Parson, 2004); and *Magma to Microbe: Modeling Hydrothermal Processes at Oceanic Spreading Centers* (edited by R. P. Lowell, J. S. Seewald, A. Metaxas, and M. R. Perfit, 2008).

The full diversity of hydrothermal systems on slow spreading ocean ridges, reflected in the contributions to this volume, is only now beginning to emerge and opens an exciting new frontier for ocean ridge exploration.

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Geophysical Monograph Series 188

10.1029/2010GM000970

Diversity of Hydrothermal Systems on Slow Spreading Ocean Ridges: Introduction

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Emerging findings reveal that hydrothermal systems at slow to ultraslow spreading ocean ridges (full spreading rate $\leq$3 cm a^{-1}) exhibit greater diversity than that at intermediate to fast spreading ridges, related to geological diversity of the lithosphere that hosts the systems. This volume presents studies of hydrothermal systems on slow spreading ocean ridges that convey our present knowledge of their diversity. *Rona* [this volume] places them in the perspective of the development of seafloor hydrothermal research. *Lowell* [this volume] reviews the various heat sources and heat transfer processes at slow spreading ridges, including mantle heat flux, mining of crustal heat, the role of exothermal chemical reactions as the principal heat sources in low-temperature hydrothermal fields, and replenishing magmatic heat sources as the principal heat sources in high-temperature hydrothermal fields. *Edmonds* [this volume] synthesizes the compositional variation of low- and high-temperature fluids discharged from hydrothermal systems hosted in the various lithologies on slow spreading ridges. *Tivey and Dyment* [this volume] explain the use of different magnetic signatures to infer the nature of hydrothermal alteration associated with the diverse geologic settings on slow spreading ocean ridges and the potential application of these signatures for future off-axis exploration.

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10.1029/2010GM000969

The Mid-Atlantic Ridge is the archetype of slow spreading ocean ridges and, under a variety of geographic names, extends from the Arctic Mid-Ocean Ridge southward to (but not including) the Southwest Indian Ridge. *Pedersen et al*. [this volume] show that, contrary to prediction, hydrothermal activity exceeds what would be expected by extrapolation from observations on faster spreading ridges. Taking advantage of Iceland as the part of the Mid-Atlantic Ridge most accessible to surface and subsurface studies, *Elders and Friðleifsson* [this volume] report encountering an active rhyolitic magma intrusion 2.1 km below the surface in their drilling of a geothermal recovery well in the Krafla field.

Many hydrothermal fields have been found on the northern Mid-Atlantic Ridge since the discovery of the TAG hydrothermal field in the axial valley at 26°N in 1985. Of these, investigation of the Lucky Strike hydrothermal field described by *Crawford et al*. [this volume] is the first to clearly reveal an axial magma chamber reflector similar to those seismically imaged at faster spreading axes in the Pacific. *Devey et al*. [this volume] describe the first five hydrothermal fields discovered on the southern Mid-Atlantic Ridge and present criteria to guide ongoing exploration.

Sauter and Cannat [this volume] describe the ultraslow spreading (full rate 1.4 cm a^{-1}) Southwest Indian Ridge as a melt-poor end-member of the ridge system with thin or no crust similar to the Gakkel Ridge in the Arctic Ocean.

Detachment faulting and associated core complexes represent a fundamental mode of crustal extension initially recognized in continental crust and emerging as a major mode of deformation of oceanic lithosphere, as presented by *John and Cheadle* [this volume]. *McCaig et al*. [this volume] review geological and geochemical data to show the key role of detachment faulting in spreading of ocean lithosphere and hydrothermal circulation over the whole range of slow spreading settings and apply findings to a detachment fault zone at the TAG hydrothermal field.

Geochemical and thermal outputs of serpentinization reactions are widespread on magma-poor sections of slow spreading ridges where hydrothermal circulation accesses and hydrates mantle peridotites. *Cannat et al*. [this volume] estimate significant hydrogen and abiogenic methane fluxes generated by serpentinization reactions related to hydration of mantle-derived ultramafic rocks that outcrop in detachment faults and magma-starved sections predominantly of the axial valley. *Charlou et al.* [this volume] demonstrate that ongoing serpentinization is mainly responsible for hydrogen and abiogenic methane production at axial ultramafic exposures that are common along sections of the northern Mid-Atlantic Ridge. They extrapolate a global hydrogen flux from the flux measured at the Rainbow field and infer significance for generation of abiogenic hydrocarbons with implications for energy and life. *Seyfried et al*. [this volume] combine theoretical reaction path models with experimental data to determine fluid-mineral equilibria controls on the chemistry of vent fluids discharging from ultramafic-hosted hydrothermal systems (e.g., Rainbow and Logatchev) on the Mid-Atlantic Ridge.

Mineralization is a byproduct of hydrothermal systems. *Fouquet et al*. [this volume] contribute a comprehensive review of similarities and differences of high- and low-temperature mineralization produced by ultramafic-hosted (peridotites and serpentinites) and mafic-hosted (basalt) hydrothermal systems on slow spreading ocean ridges.

Hydrothermal ecosystems can be grouped into several biogeographic provinces, sometimes showing variations within sections of slow spreading ocean ridges. *Kelley and Shank* [this volume] present an overview of geologic settings and associated macrofauna on slow spreading ocean ridges. *Le Bris and Duperron* [this volume] relate distribution, abundances, and nutritional role of two chemosynthetic mussel species to available electron donors and energy sources in hydrothermal fluids to explain their adaptation to diverse hydrothermal systems along the Mid-Atlantic Ridge.

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Emerging Diversity of Hydrothermal Systems on Slow Spreading Ocean Ridges

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The development of seafloor hydrothermal research has followed a classic scientific progression in which discoveries were initially interpreted as special cases until further exploration revealed their more general significance. The first high-temperature seafloor hydrothermal system was found at the Atlantis II Deep of the slow spreading Red Sea in 1963. At that time, the hydrothermal activity was largely discounted as an anomaly associated with continental rifting rather than as part of an early stage of opening of an ocean basin that could continue with the development of ocean ridges as in the Atlantic. When high-temperature black smoker hydrothermal venting was found on the East Pacific Rise in 1979, the scientific consensus then held that the relatively high rate of magma supply at intermediate to fast spreading rates was required for such activity. Accordingly, high-temperature hydrothermal activity could not occur on the slow spreading half of the global ocean ridge system. High-temperature black smokers like those on the East Pacific Rise were first discovered on a slow spreading ocean ridge at the TAG hydrothermal field on the Mid-Atlantic Ridge in 1985. The scientific consensus then ruled out the possibility for such activity on the ultraslow portion of the ocean ridge system. Plumes indicative of active high-temperature black smokers were found on the ultraslow spreading Gakkel Ridge in the Arctic in 2001, and active black smokers were found on the Southwest Indian Ridge in 2006. A diversity of high-temperature hydrothermal systems remains to be found on ocean ridges, particularly at slow spreading rates.

1. EARLY RIFTING

The first hydrothermal system discovered at a slow spreading divergent plate boundary was the Atlantis II Deep system at 21°N in the Red Sea, which contains the most efficient metallic ore-forming system and the largest seafloor hydrothermal mineral deposit found to date. A temperature and salinity (T-S) anomaly was recorded on a hydrocast made at this site by the Swedish oceanographic research vessel *Albatross* while transiting the Red Sea on return from the Indian Ocean in 1948 [*Bruneau et al*., 1953; *Pettersson et al*., 1951]. The anomaly was overlooked at the time because the cruise geochemist, G. Arrhenius, had left the ship to secure his engagement after the long separation of the expedition. Arrhenius (personal communication, 12 January 2010) now comments, “So our sixty two year happy marriage was saved at the expense of not tying down the discovery of the Red Sea hot brine.”

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10.1029/2010GM000944

Fifteen years later in November 1963, scientists traversing the Red Sea on board the British oceanographic vessel HMS *Discovery* as part of the International Indian Ocean Expedition (1963–1965) noted on their echo sounder profile a reflecting interface anomalously near the seafloor, coincidentally at the same site as the *Albatross* T-S anomaly [*Swallow*, 1969]. Water samples of hot brine were recovered from this site by the R/V *Atlantis II* in July 1963, which is now known as the Atlantis II Deep [*Miller*, 1964; *Swallow*, 1969]. Additional deeps containing hot brines and metalliferous sediments were subsequently found in the Red Sea [*Charnock*, 1964; *Swallow and Crease*, 1965; *Dietrich and Krause*, 1969]. At the time of discovery, the Red Sea hot brines and metalliferous sediments were considered an anomalous phenomena related to continental rifting rather than as part of the opening of an ocean basin from early rifting to development of an ocean ridge, as in the Atlantic Ocean.

2. HYDROTHERMAL CIRCULATION AT OCEAN RIDGES

Evidence that hydrothermal circulation is a major process at ocean ridges and the theory of plate tectonics developed together in the late 1960s and 1970s. Hydrothermal circulation and plate tectonics changed the understanding of ocean basins from passive sinks for material derived from land to active sources of new lithosphere and fluids at divergent plate boundaries. Thermal and mineralogical studies provided early evidence for hydrothermal circulation in the ocean lithosphere at ocean ridges. Comparison of the theoretical amount of heat produced by the generation of lithosphere with measurements of conductive heat flow at ridge crests revealed a discrepancy that was attributed to cooling by hydrothermal circulation of seawater through ocean crust and upper mantle [*Elder*, 1965; *Deffayes*, 1970; *Lister*, 1972; *Williams and Von Herzen*, 1974; *Wolery and Sleep*, 1976]. Alteration in ophiolites [*Muehlenbachs and Clayton*, 1972; *Spooner and Fyfe*, 1973], the association of metalliferous sediments with ocean ridges [*Skornyakova*, 1965; *Bostrom and Peterson*, 1966; *Corliss*, 1971; *Bonatti et al*., 1976], and measurement of manganese accumulation rates and fractionation in metalliferous sediments [*Bender et al*., 1971] and crusts [*M. R. Scott et al*., 1974] indicated subseafloor hydrothermal circulation and discharge into the near-bottom water column.

3. GALAPAGOS RIFT

The first discovery of an active hydrothermal system at a submerged ocean ridge was of low-temperature diffuse flow at the intermediate spreading rate Galapagos rift (full rate 6 cm a^{-1}) at the equator near 86°W longitude in 1977. The discovery was made on the basis of several lines of evidence comprising measurement of anomalously low conductive heat flow indicative of hydrothermal cooling [*Williams et al*., 1974], delineation of hydrothermal plumes in the water column by detection of the conservative primordial isotope ^{3}He derived from mantle outgassing associated with small positive temperature anomalies [*Weiss et al*., 1977], and total dissolvable manganese anomalies [*Klinkhammer et al*., 1977] in the near-bottom water column. A chemosynthetic vent ecosystem with tubeworms and clams was first imaged [*Lonsdale*, 1977; *Corliss and Ballard*, 1977] and sampled [*Grassle*, 1983] at this site.

Jenkins et al. [1978] measured the ratio of dissolved ^{3}He to transported heat ($7.6 \pm 0.5 \times 10^{-8}$ cal atom^{-1} ^{3}He) over the temperature range of the Galapagos hydrothermal solutions (3°C to 13°C). They extrapolated the observed ^{3}He to heat ratio to the global oceanic flux of ^{3}He (4 ± 1 atoms cm^{-2}) estimated by the global integration of the ^{3}He anomaly measured at mid depth in the water column [*Craig et al*., 1975] to determine a global seafloor hydrothermal heat flux assumed to be focused at ocean ridges ($4.9 \pm 1.2 \times 10^{19}$ cal a^{-1}). *Edmond et al*. [1979] used dissolved silicon concentration as a proxy for temperature and used magnesium as an indicator of mixing with seawater over the narrow temperature range of the Galapagos diffuse vent fluids. They extrapolated measured concentrations of dissolved major and minor elements versus silicon to predict the composition and temperature of a high-temperature end-member solution with zero magnesium at about 350°C. They extrapolated the measured temperature dependence of the concentration anomalies (moles per calorie) to the estimated global seafloor hydrothermal heat flux [*Jenkins et al*., 1978] to compute global fluxes of the elements. They determined that the hydrothermal fluxes for Mg and SO_4 balance river input, that Li and Rb exceed river input by factors between 5 and 10, and that K, Ba, and Si are between one third and two thirds of river load.

4. EAST PACIFIC RISE

Extinct massive sulfide chimneys were first found on the intermediate spreading rate (full rate 6 cm a^{-1}) East Pacific Rise at 21°N,103°W in 1978 using the French human-occupied vehicle (HOV) *Cyana* [*Francheteau et al*., 1979]. A dive series in the same area the following year with the American HOV *Alvin* discovered black smoker chimneys discharging hydrothermal solutions [*Spiess et al*., 1980] with temperature (350°C) and composition of the end-member solutions predicted by *Edmond et al*. [1979]. This stunning corroboration supported the estimates by *Edmond et al*. [1979]

of the large global magnitudes of hydrothermal fluxes from vents on ocean ridges. Estimates of the global seafloor hydrothermal heat flux based on the discrepancy between calculated theoretical heat production by emplacement of lithosphere at divergent plate boundaries and measured conductive heat flow on ocean ridges [*Williams and Von Herzen*, 1974; *Wolery and Sleep*, 1976] indicated that a substantial fraction (~40%) of global heat loss derives from the cooling of relatively young oceanic lithosphere by hydrothermal circulation, consistent with the estimate by *Jenkins et al*. [1978]. The large estimated global magnitudes of chemical and thermal fluxes of hydrothermal circulation assumed to be focused at ocean ridges and the associated chemosynthetic ecosystems effectively launched hydrothermal research at ocean ridges.

A NATO Advanced Research Institute on Hydrothermal Processes at Seafloor Spreading Centers was convened at the University of Cambridge, England, in 1982 and brought together some 63 scientists, virtually the entire seafloor hydrothermal community at that time [*Rona et al*., 1983]. The participants are now recognized as pioneers and founders of the field (Figure 1). The early hydrothermal discoveries at Pacific Ocean ridges initiated two paradigms at that time: (1) Relatively high magma supply rates at intermediate to fast spreading rates were required to drive high-temperature hydrothermal activity, thus eliminating the slow spreading half of the global ocean ridge system as prospective for such activity. (2) Seafloor hydrothermal systems involve the reaction of seawater convectively driven by magmatic heat with ocean crust; since the compositions of seawater and of ocean crust (basalt and gabbro) are relatively uniform, the solution chemistry at vents on ocean ridges was expected to be uniform. Therefore, J. M. Edmond initially declared that the study of solution chemistry of seafloor hydrothermal systems would be "stamp collecting" [*Rona et al*., 1982].

5. MID-ATLANTIC RIDGE

The majority consensus that favored intermediate to fast spreading for high-temperature hydrothermal activity advocated that seafloor hydrothermal research be focused on ocean ridges in the Pacific and criticized such work elsewhere as a waste of resources. A minority view contended that the slow spreading portion of the global ocean ridge system was prospective for the occurrence of high-temperature hydrothermal systems. Evidence favoring the occurrence of high-temperature hydrothermal activity on slow spreading

Figure 1. Photograph of participants in the NATO Advanced Research Institute, Hydrothermal Processes at Seafloor Spreading Centers, convened 5–8 April 1982 at Cambridge University, England. Front row (left to right): H. Craig, D. S. Cronan, J. Francheteau, C. R. B. Lister, G. Thompson, K. C. Macdonald, F. Machada, P. A. Rona, J. Honnorez, R. F. Dill, R. D. Ballard, N. A. Ostenso, R. Hessler, H. Thiel, and F. Grassle. Second row (left to right): J. Verhoef, R. Whitmarch, V. Stefansson, B. E. Parsons, T. Juteau, G. A. Gross, H. P. Taylor Jr., F. Albarede, H. Jannasch, E. Bonatti, K. Crane, J. Lydon, I. D. MacGregor, and E. R. Oxburgh. Third row (left to right): R. Hekinian, B. J. Skinner, C. Mevel, L. Widenfalk, R. Bowen, H. Bougault, T. H. van Andel, J. R. Cann, R. J. Rosenbauer, D. T. Rickard, A. Malahoff, S. P. Varavas, and M. J. Mottl. Fourth row (left to right): K. Brooks, J. W. Elder, B. Stuart, K. Gunnesch, A. Fleet, H. T. Papunen, A. H. F. Robertson, S. A. Moorby, J. Boyle, C. Lalou, and V. Ittekott. Top row (left to right): K. K. Turekian, J. Hertogen, J. A. Pearce, J. M. Edmond, S. D. Scott, D. B. Duane, A. S. Laughton, H.-W. Hubberton, R. Chesselet, and R. L. Chase. From *Rona* [1982].

ocean ridges included the Atlantis II Deep hydrothermal system in the Red Sea and the spectacular diversity of hydrothermal systems on Iceland as an emergent section of the slow spreading Mid-Atlantic Ridge. The trans-Atlantic geotraverse (TAG) project was initiated in 1970 to develop a standard crustal section across the central North Atlantic [*Rona and Orlin*, 1971], as a contribution to the International Decade of Ocean Exploration [*Intergovernmental Oceanographic Commission*, 1974]. The crustal section comprised a 330-km-wide corridor that followed mean flow lines of seafloor spreading between points that were conjugate in the *Bullard et al*. [1965] fit prior to opening of the Atlantic (Cape Hatteras, North America and Cap Blanc, northwest Africa). In addition to conducting underway geotraverses (bathymetry, magnetics, and gravity [*Rona*, 1980]), the TAG project studied representative areas of the continental margins, abyssal plain, and Mid-Atlantic Ridge within the corridor.

Dredging of the east wall of the section of the axial valley of the Mid-Atlantic Ridge within the TAG corridor unexpectedly recovered patchy manganese crusts that were thicker (centimeters), more fractionated (~40% Mn), and more rapidly accumulated (radiometrically measured rates to 200 mm 10^6 a^{-1}) than hydrogenous crusts previously recovered from ocean ridges. These properties indicated a hydrothermal origin for the manganese crusts [*M. R. Scott et al*., 1974]. In the same area of the east wall, thermistor tows recorded near-bottom temperature anomalies (0.01°C–0.1°C) with gradients that warmed downward, indicative of hydrothermal discharge from the seafloor [*Rona et al*., 1975; *Lowell and Rona*, 1976; *Rona*, 1978]. Water sampling in this area of the east wall revealed near-bottom anomalies of ^{3}He [*Jenkins et al*., 1980] and water column anomalies of dissolved and particulate manganese and iron oxides [*Klinkhammer et al*., 1984]. A metalliferous component is present in cores recovered from thin sediments (typically several centimeters thick with up to 1 m thickness in discrete ponds) in this area. The sediments are characterized by relatively rapid metal accumulation rates [*Scott et al*., 1978] and metal contents varying from disseminated [*Shearme et al*., 1983] to distinct layers including metals indicative of high-temperature discharge (Cu, Fe, and Zn [*Metz et al*., 1988]). These lines of evidence indicated proximal ongoing low- and high-temperature hydrothermal activity in this area named the trans-Atlantic geotraverse or TAG hydrothermal field [*R. B. Scott et al*., 1974]. The remaining challenge was to track the elusive thermal and chemical hydrothermal anomalies as wisps in the water column and the metals in the seafloor sediments to their source.

In 1984, N. A. Ostenso, the distinguished geophysicist who was then serving at a high level of the National Oceanic and Atmospheric Administration (NOAA), obtained congressional funding to initiate the NOAA Vents Program, dedicated to studying seafloor hydrothermal systems [*Hammond et al*., 1991]. The funding provided support to lease a long-baseline acoustic navigation system for use on an August 1985 cruise of the NOAA ship *Researcher* to the TAG field. Working within the fixed transponder navigation framework, deep-sea camera-temperature tows and water sampling tracked hydrothermal signals to their source near the base of the east wall of the axial valley [*Rona et al*., 1986]. The source is a massive sulfide mound some 200 m in diameter and 35 m high between water depths of 3635 and 3670 m surmounted by vigorously venting black smoker chimneys. The active high-temperature sulfide mound is populated by a vent ecosystem dominated by the shrimp *Rimicaris exoculata* [*Williams and Rona*, 1986], different from the ecosystem at Pacific vent sites. TAG is the first high-temperature hydrothermal system, massive sulfide deposit, and vent ecosystem found in the Atlantic and the first found on any slow spreading ocean ridge.

The summary in the article that reports this Atlantic discovery [*Rona et al*., 1986, p. 33] states: "The discovery of black smokers, massive sulfides and vent biota in the rift valley of the Mid-Atlantic Ridge demonstrates that this assemblage of hydrothermal phenomena is not limited to intermediate- to fast-spreading oceanic ridges. Hydrothermal exchange processes may thus be important at the ridges which extend though the Atlantic Ocean and western Indian Ocean, comprising more than half the 55,000-km global length of seafloor spreading centres." This statement sets the scene for the present volume with subsequent discoveries of active high-temperature hydrothermal systems on the slow spreading southern Mid-Atlantic Ridge in 2006 [*Haase et al*., 2007; *Devey et al*., this volume], the Central Indian Ridge in 2000 [*Hashimoto et al*., 2001; *Van Dover et al*., 2001], the ultraslow Southwest Indian Ridge in 2006 [*Tao et al*., 2007; *Sauter and Cannat*, this volume], the Arctic Mid-Ocean Ridge beginning in 1997 [*Hannington et al*., 2001; *Kuhn et al*., 2003; *Pedersen et al*., this volume], and plumes indicative of high-temperature black smoker venting on the ultraslow spreading Gakkel Ridge in 2002 [*Edmonds et al*., 2003]. A diversity of high-temperature hydrothermal systems remains to be found on ocean ridges particularly at slow spreading rates.

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Hydrothermal Circulation at Slow Spreading Ridges: Analysis of Heat Sources and Heat Transfer Processes

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Hydrothermal processes on slow spreading ridges exhibit several features that distinguish them from their counterparts at fast and intermediate rate spreading centers. These differences may reflect differences in magma supply rates, type of host rock, and the interplay between magmatism and tectonic extension. As a result, the heat sources and driving mechanisms for hydrothermal circulation at slow spreading ridges may differ from those on fast and intermediate spreading ridges. This paper reviews various heat sources and heat transfer processes at slow spreading ridges, including mantle heat flux, mining of crustal heat, the role of exothermic chemical reactions, and magmatic heat sources. The analyses suggest that for high-temperature, high-output systems such as TAG, Rainbow, and Lucky Strike on the Mid-Atlantic Ridge and Kairei on the Central Indian Ridge, heat transfer from convecting, an actively replenished subaxial magma chamber is required to maintain these systems on decadal time scales. Low-temperature off-axis systems such as Lost City are likely driven by heat extraction from the crust, perhaps in conjunction with downward fluid migration in reactivated faults. Serpentinization reactions appear to play a smaller role. Broken Spur is a relatively low heat output system that is likely driven by magma, but it may be in a waning phase.

1. INTRODUCTION

Hydrothermal processes at slow spreading ridges exhibit some significant differences from their counterparts at fast and intermediate spreading ridges. For example, some systems, such as the Rainbow and Logatchev vent fields on the northern Mid-Atlantic Ridge [*Douville et al.*, 2002] and the off-axis Lost City vent field [*Kelley et al.*, 2001], are hosted in ultramafic rocks rather than basalt. This difference in host rock is reflected, in part, in the chemistry of the vent fluids [e.g., *Wetzel and Shock*, 2000; *Douville et al.*, 2002; *Allen and Seyfried*, 2004] and possibly in the microbial ecosystems that extract their energy from the vent fluids [e.g., *Früh-Green et al.*, 2003, 2004]. Some of the vent fields at slow spreading ridges tend to be spatially more extensive and appear to have longer histories than their fast spreading counterparts. For example, hydrothermal circulation at the TAG mound has occurred episodically for ~10^5 years [*Lalou et al.*, 1993], and *Cave et al.* [2002] argue that the Rainbow field has been active continuously for ~10^4 years. Finally, in comparison to processes at fast and intermediate spreading centers, hydrothermal processes at slow spreading ridges may reflect differences in the nature of heat supply and the depth of hydrothermal circulation.

At fast and intermediate spreading ridges, there is growing evidence that hydrothermal venting is closely associated with the presence of thin, mostly liquid subaxial magma bodies [*Singh et al.*, 1999; *Canales et al.*, 2006]. The presence

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of such subsurface magma is inferred from seismic imaging that constrains the spatial extent of magma, the thickness of the sill [e.g., *Detrick et al.*, 1987; *Kent et al.*, 1990; *Van Ark et al.*, 2007; *Jacobs et al.*, 2007], and the extent of the underlying zone of partial melt [*Dunn et al.*, 2000]. At slow spreading ridges, the lower rate of magma supply suggests that liquid magma may be present intermittently and that partially molten zones may be less extensive [*Sinton and Detrick*, 1992]. Nevertheless, hydrothermal activity is likely to be closely associated with magmatic and volcanic processes in these settings. Seismic reflectors indicative of subsurface magma have been imaged at depths of 3 and 1.2 km beneath the Lucky Strike [*Singh et al*., 2006] and Snake Pit [*Calvert*, 1995] vent fields on the northern Mid-Atlantic Ridge, respectively. Even at the ultraslow spreading Gakkel Ridge in the Arctic Ocean, hydrothermal activity appears to be linked to recent volcanism [*Michael et al.*, 2003; *Baker et al*., 2004]. On the other hand, *Canales et al.* [2007] did not detect magma at shallow to midcrustal levels beneath the TAG hydrothermal field on the Mid-Atlantic Ridge. There may be magma yet to be detected at the base of the crust or in the upper mantle; however, the detection of earthquakes at depths of several kilometers beneath TAG suggests that hydrothermal circulation may penetrate deeply into the crust along detachment faults [*deMartin et al.*, 2007].

Although magmatic heat is still likely to be important at many hydrothermal sites along slow spreading ridges, heat from exothermic reactions such as serpentinization of peridotite [*Macdonald and Fyfe*, 1985] might also be important [e.g., *Rona et al.*, 1987; *Kelley et al.*, 2001; *Lowell and Rona*, 2002; *Emmanuel and Berkowitz*, 2006]. Moreover, heat to drive hydrothermal circulation may also come directly from the mantle, or from cooling crust and lithosphere. This paper reviews the contributions of various heat sources and heat transport processes as drivers for hydrothermal systems at slow spreading ridges. By understanding the contributions of these heat sources and heat transfer processes in various hydrothermal systems at slow spreading ridges, one may gain important insight into the physics and temporal evolution of hydrothermal processes in these complex settings. In sections 2 and 3, I first review some of the observational constraints on hydrothermal activity at slow spreading ridges. Then I discuss possible heat sources and heat transfer processes and evaluate their likely importance in the context of the observational constraints.

2. OBSERVATIONAL CONSTRAINTS AND MODELING PARAMETERS

In order to construct mathematical models of hydrothermal circulation, several pieces of data are required to constrain certain model parameters [e.g., see *Lowell and Germanovich*, 2004; *Lowell et al.*, 2008]. The primary observational constraints include the temperature and heat output of the hydrothermal system and the areal extent of the discharge zone at the seafloor. These parameters are important because together they control the mass flux and fluid velocity through the system, respectively. Because hydrothermal heat output is ultimately limited by heat conduction from rock to the hydrothermal fluid, heat output and temperature data also constrain the area over which heat conduction occurs, regardless of the source of heat supply. As a result, these data also shed light on the likely temporal evolution of the system. Finally, seafloor hydrothermal systems have a finite duration. In most cases this is somewhat uncertain. Dating of samples near the TAG mound indicate that the hydrothermal circulation has occurred episodically for ~10^5 years [*Lalou et al.*, 1993, 1995] and the most recent venting episode has lasted ~90 years [*Lalou et al.*, 1998]. *Cave et al* [2002] argue that the Rainbow vent field has lasted for 10^4 years. For the discussion of heat sources in this paper, I assume that vent fields last from decades to ~10^2 years.

Chemistry of the vent fluid, whether or not magma is present, and estimates of its volume also help to constrain mathematical models. Unfortunately, trace and major element chemical data, including chloride, have not yet been incorporated directly into the transport models; but these data provide indirect model constraints by providing information on subsurface reaction temperature and pressure [e.g., *Von Damm*, 1988, 2004; *Seyfried et al.*, 1991, 2004], type of host rock, and the importance of serpentinization [e.g., *Wetzel and Shock*, 2000; *Allen and Seyfried*, 2004]. Models incorporating precipitation of anhydrite from seawater have been used to investigate the spatial extent of hydrothermal recharge zones [*Lowell and Yao*, 2002] and mixing during hydrothermal discharge [*Lowell et al.*, 2003]. *Sleep* [1991] and *Fontaine et al.* [2001] also discuss the role of anhydrite precipitation on hydrothermal circulation. To date, reactive transport models have not been used extensively to model high-temperature hydrothermal systems at oceanic spreading centers [e.g., *Alt-Epping and Smith*, 2001]. *Alt-Epping and Diamond* [2008] provide an up-to-date review of reactive transport modeling in the oceanic crust. The presence of magma and its volume allow a calculation of the magmatic heat supply. Simple models of magma convection, crystallization, and replenishment have been incorporated into recent hydrothermal models [*Lowell et al.*, 2008; *Liu and Lowell*, 2009]. Table 1 lists known hydrothermal systems at slow spreading ridges for which the main observational constraints are available.

Table 1. Observational Data for Hydrothermal Systems on Slow Spreading Ridges

Vent Field	T_f (°C)	Heat Output H_f(MW)	Vent Field Area A_d (m^2)	Heat Transfer Area A_m[a] (m^2)	Lifetime[b] (years)
TAG[c]	360–364	1700	3×10^4	10^7	~90
Broken Spur[d]	350–365	28–275[e]	4×10^8	10^6	?
Rainbow[f]	364	1810 ± 693	3×10^4	10^7	10^4
Lucky Strike[g]	333	3800 ± 1200	2×10^5	2.1×10^7	?
Kairei[h]	360	70–120	3×10^3	10^6	?

[a]Heat transfer area is an estimate except for Lucky Strike, which comes from the imaged magma chamber [*Singh et al.*, 2006].

[b]Vent field lifetimes are poorly known. Lifetime of the current venting episode at TAG is from *Lalou et al.* [1998] based on radiometric dating of mound samples, estimate for Rainbow is from *Cave et al.* [2002] from interpretation of sediment trap data.

[c]*Rona et al.* [1993], *Rudnicki and Elderfield* [1992], *Wichers et al.* [2005], *James and Elderfield* [1996], and *Lalou et al.* [1998].

[d]*James et al.* [1995] and *Murton et al.* [1999].

[e]High temperature venting is ~28 MW over an area of ~1.5×10^4 m^2.

[f]*German and Lin* [2004], *Jean-Baptiste et al.* [2004], *Thurnherr and Richards* [2001], and *Cave et al.* [2002].

[g]*Langmuir et al.* [1997] and *Jean-Baptiste et al.* [1998].

[h]*Rudnicki and German* [2002].

In addition to the observational constraints, the construction of hydrothermal models requires the input of several thermodynamic and fluid flow parameters. Many of these are known to within ±20% or so. Because the heat output constraint is known with much less certainty, model results are considered to be relatively insensitive to the choice of these parameter values. Table 2 list the main parameters and typical values used. The most uncertain parameter is crustal permeability. Estimates of this parameter can be deduced from modeling of magma-hydrothermal systems [e.g., *Lowell and Germanovich*, 1994, 2004; *Wilcock and McNabb*, 1996; *Lowell et al.*, 2008]. Also see section 3.4.

At slow spreading ridges, magma supply is lower than at fast spreading ridges; consequently, heat sources other than magma may play a role in driving hydrothermal processes at slow spreading ridges. For example, *Lowell and Rona* [2002] have suggested that heat resulting from exothermic serpentinization reactions coupled with heat transfer from the underlying lithosphere may drive the Lost City vent field. In addition, heat derived from downward crack propagation and cooling of crustal rocks [*Lister*, 1974; *Wilcock and Delaney*, 1996] may also play a role. *Maclennan et al.* [2004] suggest that extensive hydrothermal cooling of the crust occurs near the ridge axis. Models of cellular convection have also been developed [e.g., *Rabinowicz et al.*, 1999; *Fontaine et al.*, 2008], where the planform of the convection cells is thought to be controlled, in part, by the sloping boundary of the lithosphere.

Many of these models only seek to reproduce observed hydrothermal vent temperatures but not the heat output from the hydrothermal systems. For the most part this may be because these data were not available. Advective heat flow data have recently become available from a number of sites on slow spreading ridges (see *Baker* [2007] and Table 1), however, and these data place additional constraints on hydrothermal models [see *Lowell and Germanovich*, 2004; *Lowell et al.*, 2008]. These data will be used to constrain the contributions of various heat sources and heat transfer processes in driving hydrothermal processes at slow spreading ridges.

3. HYDROTHERMAL HEAT SOURCES AND HEAT TRANSFER PROCESSES

There are four potential heat sources that may play a role in driving hydrothermal circulation in the oceanic crust. These are (1) mantle upwelling, (2) heat extraction from rocks of the crust and lithosphere, (3) heat associated with exothermic reactions such as serpentinization of peridotite, and (4) magmatic heat sources. In this paper, I briefly discuss the potential contributions of each of these.

3.1. Mantle Heat Source

The heat flux H_M transported by mantle of density ρ_r, specific heat c_r, and temperature T_M, ascending at a velocity u_M, is given by

$$H_M \sim [\rho_r c_r (T_M - T_0) + \rho_r L] u_M \times 10^6 \text{ W km}^{-2}, \qquad (1)$$

where T_0 is a background temperature to which mantle is cooled and L is the latent heat of crystallization. Symbols and parameter values are given in Table 2. Using these parameters and u_M = 0.02 m a^{-1}, typical of slow spreading ridges, one obtains

Table 2. Symbols, Definitions, and Values

Symbol	Definition	Value/Units
a	Thermal diffusivity	10^{-6} m^2 s^{-1}
A_d	Cross-sectional area discharge zone	m^2
A_m	Cross-sectional area of magma chamber	m^2
c_f	High temperature specific heat of fluid	6×10^3 J (kg °C)$^{-1}$
c_r	Specific heat of rock	10^3 J (kg °C)$^{-1}$
d	Width of crack or fault zone	m
D	Depth to lithospheric isotherm	km
g	Acceleration due to gravity	10 m s^{-2}
h	Height of fluid-filled fracture or fault zone	km
H_f	Hydrothermal heat output	MW
H_m	Heat content per unit volume of magma	10^9 J m^{-3}
H_M	Heat flux from ascending mantle	MW
H_r	Heat flux to propagating fracture/reactivated fault	W m^{-2}
$\langle H_r \rangle$	Mean of H_r for a propagating fracture/fault	W m^{-2}
k	Rock permeability	m^2
l	Length of fracture perpendicular to flow direction	m
L	Latent heat of crystallization	5×10^5 J kg^{-1}
Nu	Nusselt number	
q	Mass flow per unit length of fracture	kg (m s)$^{-1}$
Q	Total mass flux	kg s^{-1}
Ra	Rayleigh number	
t	Time	
T_f	Temperature of hydrothermal fluid	°C
T_m	Bulk temperature of convecting magma	°C
T_M	Temperature of mantle	1300°C
T_r	Temperature of rock	400–700°C
T_s	Solidus temperature of magma	860°C
T_L	Liquidus temperature of magma	1200°C
u_f	Velocity of fluid	m s^{-1}
u_m	Magma replenishment velocity	m^3 (m^2 s)$^{-1}$
u_M	Velocity of ascending mantle	0.01 m a^{-1}
$<v>$	Velocity of downward propagating fault	0.1–10 m a^{-1}
x	Horizontal coordinate	
X	Horizontal distance between fractures	
z	Vertical Cartesian coordinate	
	Greek Symbols	
α	Thermal expansion coefficient of fluid	~10^{-3} °C^{-1}
χ	Crystallinity of magma	
δ	Conductive thermal boundary layer between magma and base of hydrothermal system	m
ζ	Latent heat of reaction in serpentinization	2.5×10^5 J kg^{-1}
λ_r	Thermal conductivity of rock	2.0 W (m °C)$^{-1}$
ν	Kinematic viscosity of fluid	~10^{-7} m^2 s^{-1}
ρ_f	Fluid density	10^3 kg m^{-3}
ρ_m	Density of magma	2.8×10^3 kg m^{-3}
ρ_r	Density of rock	3.3×10^3 kg m^{-3}
τ	Time interval	
	Subscripts	
0	Reference, background or boundary value	
d	Discharge zone	
f	Fluid	
m	Magma	
M	Mantle	
r	Rock	

$$H_M \sim 1.8 \text{ MW km}^{-2}. \quad (2)$$

Heat output from hydrothermal systems at slow spreading ridge typically ranges between 10^2 and 10^3 MW (Table 1), suggesting that for steady state mantle upwelling to drive hydrothermal flow, each hydrothermal system would need to extract the mantle heat from an area between 10^2 and 10^3 km^2. Hydrothermal recharge would occur outside of this region. Finally, to tap such an extensive region, the convective circulation would have to be extremely heterogeneous, and even if the heat supply could be tapped, the temperature of the resulting hydrothermal system is essentially unconstrained. For this heat to be extracted by a single high-temperature hydrothermal system would require a very fortuitous permeability distribution. It seems unlikely that heat transfer associated with broadscale mantle upwelling will contribute significantly to high-temperature hydrothermal venting.

3.2. Heat Transfer From Cooling Crust and Lithosphere

The earliest models of hydrothermal circulation were used to explain the low values of conductive heat flow compared with models of lithospheric cooling [e.g., *Sclater et al.*, 1980]. Because conductive heat flow data were obtained in sedimented regions far from the ridge axis, the models were generally concerned with mining of heat from crustal rocks by hydrothermal circulation [e.g., *Bodvarsson and Lowell*, 1972; *Williams et al.*, 1974; *Lowell*, 1975]. Adapting a suggestion of *Bodvarsson and Lowell* [1972] that thermal cooling and contraction could generate cracks and enhance permeability, *Lister* [1974, 1982] developed a model of hydrothermal circulation based on the downward propagation of thermal contraction cracks. The Lister model not only predicted the possibility of high-temperature hydrothermal venting, but his idea of a downward propagating "cracking front" has often been invoked over the years as a mechanism of hydrothermal heat transfer [e.g., *Mével and Cannat*, 1991; *Lowell and Germanovich*, 1994; *Seyfried and Ding*, 1995; *Wilcock and Delaney*, 1996; *Sohn et al.*, 1998, 1999; *Johnson et al.*, 2000; *Wilcock and Fisher*, 2004; *Tolstoy*, 2008].

When high-temperature vents were discovered [*Spiess et al.*, 1980], attempts were made to explain the observations as a result of extracting heat from crustal rocks. It became quickly apparent that high-temperature venting could not be maintained for very long by this process [*Strens and Cann*, 1982; *Lowell and Rona*, 1985], and magmatic heat sources were invoked [*Cann and Strens*, 1982; *Lowell and Rona*, 1985; *Lowell and Burnell*, 1991].

In this section, I first briefly review the difficulties in driving hydrothermal circulation by mining heat from a fixed region of crustal rock. I then discuss heat transfer resulting from downward migration of fluid into the crust. Such fluid migration may occur, in principle, as a result of downward propagating fluid-filled cracks as suggested by *Lister* [1974, 1982] or as a result of enhanced permeability associated with reactivated faults. Finally, I will briefly discuss heat transfer from cooling lithosphere during cellular convection.

3.2.1. Fluid circulation in fractures and heat transfer from crustal rocks. The problem of fluid circulation in fractures and heat transfer from crustal rocks has been discussed by *Bodvarsson and Lowell* [1972], *Lowell* [1975], *Strens and Cann* [1982], and *Lowell and Rona* [1985] in the context of seafloor hydrothermal systems and by *Bodvarsson* [1969], *Gringarten et al.* [1975], and *Lowell* [1976] in the more general context of a hot dry rock geothermal system.

The simplest way to understand this problem is to consider cold fluid at temperature T_{f0} entering a thin isolated crack of length l and height h embedded in hot rock at some initial temperature T_{r0} (Figure 1) and flowing toward the surface at some mean velocity u_f. Heat is transferred by conduction from the rock to the moving fluid, which exits the rock at some height h at a temperature $T_f(h,t)$. The thermal problem consists of solving for heat conduction in the rock together with convective heat transport by the fluid moving in the fracture. Because the rate of conductive heat

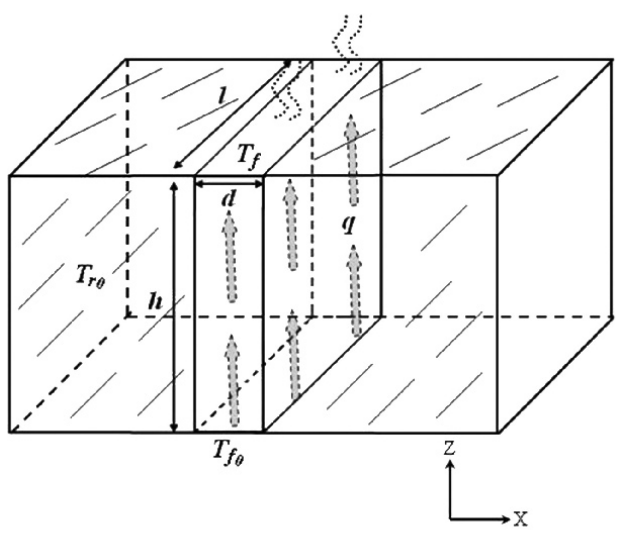

Figure 1. Schematic of heat extraction from crustal rock by upward flowing fluid in a single thin fracture of length l and width d. Fluid enters the fracture at temperature T_{f0} and exits at height h at a temperature T_f. The initial temperature of the impermeable rock is T_{r0}. Heat is transferred by conduction from the rock to fluid moving upward at a constant mass flow rate q per unit length of the fracture.

flow perpendicular to the crack is much greater than conductive heat flow parallel to the crack, the heat conduction problem in the rock is given by [e.g., *Bodvarsson*, 1969; *Lowell*, 1976]

$$\frac{\partial T_r(x,t)}{\partial t} = a_r \frac{\partial^2 T_r(x,t)}{\partial x^2}, \tag{3}$$

where a_r is the thermal diffusivity of the rock. Equation (3) is solved with the initial condition

$$T_r(x,0) = T_{r0}. \tag{4}$$

For fluid moving in a thin fracture of width d, the rate at which heat is advected by the moving fluid is equal to the rate at which heat is conducted across the two walls of the fracture. That is [e.g., *Bodvarsson*, 1969; *Lowell*, 1976],

$$\rho_f c_f d u_f \frac{\partial T_f(z,t)}{\partial z} = -2\lambda_r \left.\frac{\partial T_r(x,t)}{\partial x}\right|_{x=0}, \tag{5}$$

where ρ_f, c_f, and λ_r are the density of the fluid, specific heat of the fluid, and thermal conductivity of the rock, respectively. The conductive heat flux from the rock is solved at $x = 0$ because the fracture is thin. The solution to equation (5) is subject to the boundary condition

$$T_f(0,t) = T_{f0}. \tag{6}$$

It is also assumed that the temperature of the fluid and rock are equal at the wall of the fracture. Hence,

$$T_f(z,t) = T_r(0,t). \tag{7}$$

The solution to this set of equations is [e.g., *Bodvarsson*, 1969; *Lowell*, 1976]

$$T_f(z,t) = T_{f0} + (T_{ro} - T_{fo})\,\mathrm{erf}\left[\frac{\lambda_r z}{2 c_f q \sqrt{a_r t}}\right], \tag{8}$$

where erf is the error function and $q = \rho_f c_f u_f d$ is the mass flux per unit length of the fracture.

If we identify the height $z = h$ with the seafloor, then $T_f(h,t)$ gives the vent temperature as a function of time for given initial temperature, flow rate, etc. As an example, suppose $h = 10^3$ m, $T_{r0} = 400°C$, $T_{f0} = 100°C$. From equation (8) the vent temperature is initially equal to 400°C and decreases with time. To find the time for $T_f(h,t)$ to fall to 300°C, for example, equation (8) gives

$$\mathrm{erf}\left[\frac{\lambda_r h l}{c_f Q \sqrt{at}}\right] = \frac{T_f(h,t) - T_{f0}}{T_{ro} - T_{f0}} = 0.67, \tag{9}$$

where $Q = ql$ is the total mass flow rate in kg s^{-1}. For values of the error function <0.7, the error function is approximately equal to its argument. Hence, as the vent temperatures declines to <300°C, it will decay as $t^{-1/2}$. One can further see from equation (9) that to reach a given vent temperature, there is a simple trade-off between the lifetime of the system and the flow rate. The longer the desired lifetime, the smaller the flow rate needs to be. But for a given vent system, the flow rate is constrained by the observed heat output. For example, the vent fields listed in Table 1 typically have discharge temperatures of ~350°C and total heat outputs of $H_f \sim 10^9$ W. The mass flow Q rate through such a system is

$$Q = \frac{H_f}{c_f \Delta T} \sim \frac{10^9}{6 \times 10^3 \times 350°\mathrm{C}} \sim 500 \text{ kg/s}. \tag{10}$$

Substituting this value for the flow rate into equation (9) and assuming a fracture length $l = 10^3$ m and thermal properties listed in Table 2, we find that the time for the temperature to decay from 400°C to 300°C is ~10^6 s. This result is similar to those of *Strens and Cann* [1982] and *Lowell and Rona* [1985], who emphasized that black smoker flow cannot be maintained by mining heat from the surrounding rock by flow in a single fracture or fault zone.

For this simple model, one can lengthen the lifetime by simply lowering the mass flow rate in the fracture and then adding the flow through a number of independent fractures to get the total observed flow rate. To increase the time to 30 years, for example, equation (9) gives a flow rate of ~16 kg s^{-1} for an individual fracture. If 30 independent fractures were each transporting hydrothermal fluid at this rate, the hydrothermal temperature would decrease from 400°C to 300°C over 30 years. But for these fractures to be thermally independent, they would have to be spaced far enough apart to not thermally interfere with each other. This spacing is controlled by the thermal diffusion time scale τ, which is given by

$$\tau \sim X^2 / a_r, \tag{11}$$

where X is the distance between fractures. For τ = 30 years, $X \sim 30$ m. Hence, for a typically hydrothermal system on a slow spreading ridge to be driven by crustal heat alone, heat would be mined from ~1 km^3 of crust as the vent temperature decreased from 400°C to 300°C over a 30-year time period.

The model described above would constitute a highly idealized system because the flow rate is fixed, hydrothermal recharge is not accounted for, and no mechanism of focusing the flow from disparate fractures occupying a kilometer of crust across axis to focused discharge in a much narrower region has been discussed. In a natural convection system, flow results from buoyancy differences between the warm and cold fluids. If the permeability of the system remained constant, the flow rate would decrease as the temperature of the venting fluid decreased. In this case, both the temperature and the hydrothermal heat output would decay as $t^{-1/4}$ [e.g., *Lowell et al.*, 1995; *Lowell and Germanovich*, 1994, 2004]. Moreover, the recharge zone would have to be situated so as not to thermally interfere with the discharge zone. The fundamental point of this exercise is to emphasize that high-temperature, high-heat output hydrothermal circulation driven by mining heat from crustal rocks cannot be maintained in steady state. Such a system would always decay in a relatively predictable manner.

3.2.2. Heat transfer from downward fluid migration. *Lister* [1974] developed a theory for heat extraction from a set of downward propagating thermal contraction cracks. This theory has been widely accepted, at least conceptually, and the idea of a "cracking front" is commonly used in the mid-ocean ridge hydrothermal community. This model has been appealing for several reasons. First, it has been viewed as an alternative to subsurface magma as a heat source to drive hydrothermal circulation [e.g., *Wilcock and Delaney*, 1996]. Such a different mode of heat transfer appeared to be necessary in regions of hydrothermal activity where magma appeared to be absent, such as the Endeavour segment of the Juan de Fuca Ridge and slow spreading ridges in general. Second, vent fluids exhibit relatively constant concentrations of incompatible trace elements such as Li and B over periods of years [*Campbell et al.*, 1988]. Because these elements are readily leached from basalt, *Seyfried et al.* [1991] argue that hydrothermal fluids are continuing to react with relatively unaltered basalt, perhaps in connection with a downward migrating fracture front [*Seyfried and Ding*, 1995].

There are flaws in the *Lister* [1974, 1982] model, however, so it must be viewed with considerable caution. First, the model assumes that the cracks are tensile. The compressive stresses stemming from the overburden are neglected even as the cracks propagate to depths of kilometers into the crust. Second, the model predicts fluid penetration through macroscale cracks with a spacing ~0.1–1 m. Recent studies of crustal rocks from Hess Deep and the Oman ophiolite indicate that the first cracks to form are grain boundary microcracks and that larger macrocracks occur later [*Manning and MacLeod*, 1996; *Manning et al.*, 2000]. Moreover, the Lister model predicts that the cracking temperature should decrease with depth, which is counter to observations [*Manning et al.*, 2000]. Finally, the Lister model predicts that the cracking front would migrate downward at ~1–10 m a^{-1}, resulting in full penetration of the crust on a time scale that is essentially instantaneous relative to seafloor spreading. In this scenario, cracking would occur across a substantial temperature range, which is inconsistent with the assumption that cracking occurs over a relatively small temperature range [*Lister*, 1974, 1982; *Mével and Cannat*, 1991]. If the cracking front is nearly isothermal, or has a slight increase in temperature with depth, then the effective rate of crack propagation is approximately centimeters per year and would only penetrate the crust several kilometers off axis [*Manning et al*., 2000]. In this case the rate of heat transport associated with the crustal cooling associated with a downward migrating cracking front would be orders of magnitude less than suggested by *Lister* [1974, 1982].

With the *Lister* [1974, 1982] model, it is also difficult to account for observed rock alteration, which occurs in association with microscale cracks [*Manning and MacLeod*, 1996] and with the relative constancy of incompatible trace elements in vent fluids. If hydrothermal circulation occurs through cracks separated by approximately meters, the fluid has little direct access to rock, and hence, the extraction of trace elements is problematical. On the other hand, if cracking occurs at the microscale, the fluid has access to essentially the entire rock volume, and trace elements may be extracted, even if the fluid is not vigorously circulating through the microcrack network [*Lowell et al.*, 2008].

Another way to consider fluid migration associated with downward crack propagation is to associate it with diking events or reactivation of preexisting faults. *Bodvarsson* [1982] develops a model of downward crack propagation near the boundary of a dike. *Lowell and Germanovich* [1994] suggest that cracks may propagate laterally near the top of a magma chamber as a result of hydrothermal cooling near a dike. Fluid may also migrate downward and extract heat from deeper crustal rocks as a result of fault reactivation. Thermal stresses associated with fluid circulation in the crust may cause a preexisting fault to be reactivated by a slip episode. The slip zone may then propagate downward aseismically, resulting in enhanced permeability at depth. The hydrothermal fluid would then migrate downward into this zone of enhanced permeability. Similar processes are known to occur during production of petroleum reservoirs [e.g., *Segall et al.*, 1994; *Germanovich et al.*, 1999]. Finally, *Mével and Cannat* [1991] argue that the seafloor spreading rate may impact permeability generation, noting that at slow spreading ridges, high-temperature ductile shear zones predate cracking.

A model for heat extraction as a result of downward fluid migration associated with downward propagating cracks or a reactivated fault zone (Figure 2) can be constructed using an approach similar to that of *Bodvarsson* [1982]. Consider a fluid-saturated vertical fault zone imbedded in a rock volume with temperature T_r, density ρ_r, and specific heat c_r as shown in Figure 2. Suppose fluid is circulating with a mean temperature T_f within a fault zone of height h and length l. Assuming that heat is transferred from the rock to the fluid in the fault as though by conduction from a half-space [*Carslaw and Jaeger*, 1959], then the instantaneous heat flux from the rock to the fluid is

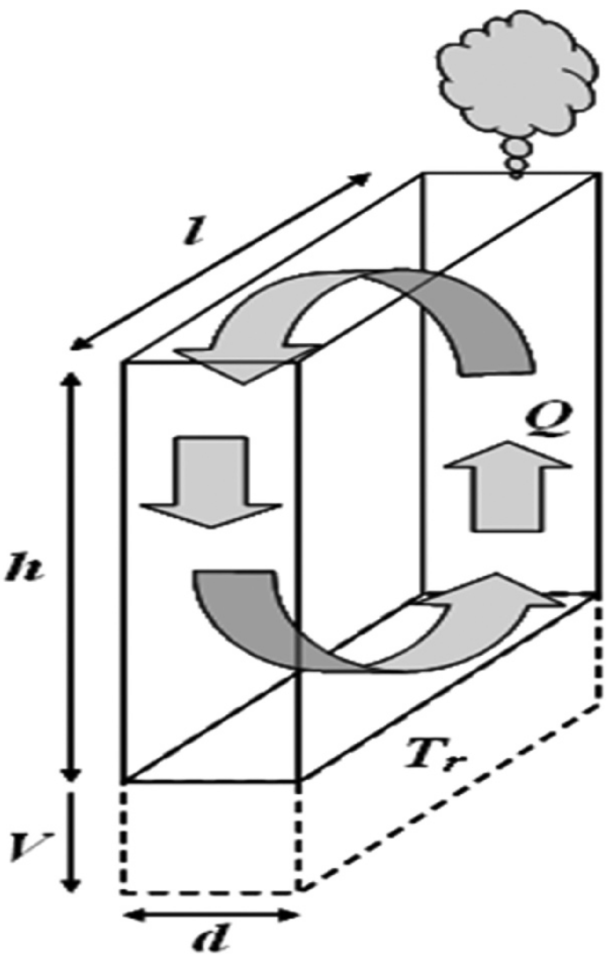

Figure 2. Schematic of fluid circulation in downward propagating fracture, or downward migration of fluid in a reactivated fault zone, of height h, width d, and length l. As the fluid migrates downward at velocity $\langle v \rangle$, heat is transported by conduction from the rock at temperature T_r to the circulating fluid. Q represents the total mass flux of fluid within the fault or fracture zone.

$$H_r = \frac{2\lambda_r (T_r - T_f)}{\sqrt{\pi a_r t}}, \tag{12}$$

where λ_r is the thermal conductivity of the rock. The factor of 2 arises because heat is transferred through both walls of the fault. Assuming that the crack migrates downward, or fluid migrates downward along a reactivated fault zone at a constant rate $\langle v \rangle$, the average rate $\langle H_r \rangle$ at which heat transferred to the fluid per unit length of the fault or fracture surface over a time interval τ is found by integrating equation (12) over the interval $0 \le t \le \tau$. The result is

$$\langle H_r \rangle = 4\lambda_r (T_r - T_f)\langle v \rangle \left(\frac{\tau}{\pi a_r} \right)^{1/2} = H_f. \tag{13}$$

One can express equation (13) in terms of the depth of the crack or fluid migration h by writing $\tau = h/\langle v \rangle$. Figure 3 plots $\langle H_r \rangle$ as a function of h for various propagation rates, assuming T_r = 700°C, the mean temperature of the fluid T_f = 200°C, and rock parameters given in Table 2.

Figure 3 shows that the mean rate of heat transfer per kilometer of fault or fracture length from downward migrating fluid is <100 MW km^{-1}, even for deep crack propagation or fault reactivation, unless the rate of propagation is high (~10 m a^{-1}). For rapid rates of crack or fluid propagation, the heat output exceeds a few hundred MW per kilometer at

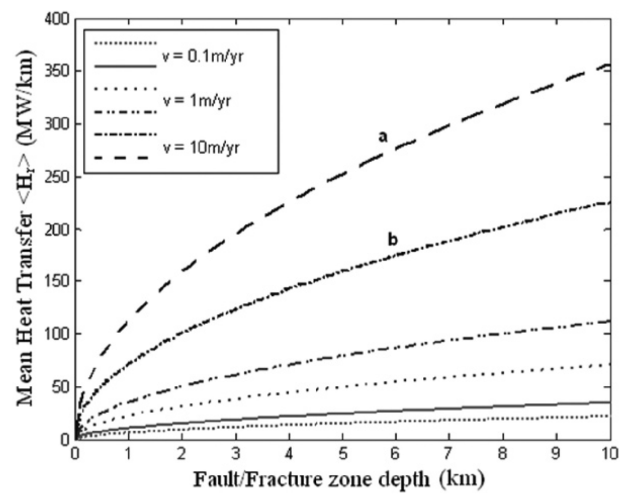

Figure 3. Mean heat flux from the rock to downward migrating fluid within a propagating fracture or reactivated fault zone as a function to total fracture or fault zone depth for various values of migration velocity. For each velocity, the upper curve (curve a) includes the effect of serpentinization, whereas the lower curve of the pair (curve b) does not.

depths of a few kilometers. This would require crack propagation or fault reactivation to occur at this rate for several hundreds of years. Moreover, *Manning et al.* [2000] argue that the rate of crack propagation may be only centimeters per year. It thus seems unlikely that high-temperature, high-heat output hydrothermal systems such as TAG, Lucky Strike, or Rainbow (see Table 1) could derive their heat in this manner. It may be possible, however, for lower-temperature, lower-heat output hydrothermal systems, such as Lost City or Broken Spur (Table 1), to derive their heat from hot rock near downward propagating fluid-filled cracks or downward migration of fluids in reactivated fault zones.

3.2.3. Cellular convection and heat transfer from cooling lithosphere. A standard approach to modeling hydrothermal circulation has been to consider convection in a horizontal porous layer heated from below. This approach stems from the classical stability analysis [*Lapwood*, 1948; *Nield*, 1968] and has often been applied to studies of oceanic spreading centers [e.g., *Ribando et al.*, 1976; *Fehn and Cathles*, 1979; *Cherkaoui et al.*, 1997; *Wilcock*, 1998; *Fontaine et al.*, 2001; *Coumou et al.*, 2006, 2008; *Lowell et al.*, 2007]. In most cases, the models are two-dimensional with constant temperature maintained at the base. Models constructed by *Rabinowicz et al.* [1999] and *Fontaine et al.* [2008] consider the effect of a sloping bottom boundary.

In particular, *Fontaine et al.* [2008] argue that at slow spreading ridges, hydrothermal circulation mines heat from the lithosphere, resulting in a sloping lithospheric boundary along the ridge. At slopes greater than 15°–20°, circulation tends to be focused into one large cell, with broad recharge mining heat from the mantle and hydrothermal discharge focused at the segment center.

A key assumption of *Fontaine et al.* [2008] is that the sloping lithospheric boundary is maintained at constant temperature on time scales corresponding to the development of hydrothermal circulation cells. That is only possible, however, if the rate of cooling of the lithosphere by hydrothermal circulation is equal to the rate at which heat is transported into the base of the lithosphere. In the absence of magmatic input, either conductive cooling or vigorous hydrothermal circulation will cool the crust and upper mantle, resulting in downward migration of the lithospheric isotherm. If only conductive cooling occurs, the rate of migration of an isotherm is given by the thermal diffusion time scale (equation (11)). Hence, the depth D to the 600°C isotherm is given by

$$D \sim (a_r t)^{1/2}. \tag{14}$$

To achieve a slope of 15° over a lateral distance along the ridge axis of 20 km, $D = 5.4$ km relative to the lithospheric thickness at the segment center. From (14), this would occur after a time of ~10^6 years. If such an along-axis lithospheric slope occurred as a result of thermal conduction alone, it would imply that no magma had been input along the segment in ~10^6 years. In this time the ridge would have undergone ~10 km of spreading at a half rate of 0.01 m a^{-1}. On the other hand, if hydrothermal circulation is actively mining heat from the lithosphere, then the isotherm could migrate downward significantly faster, and the quasi-steady state convection regime envisioned by *Fontaine et al.* [2008] might not become established. Rather than simply model hydrothermal circulation assuming a steady state sloping lithosphere, downward migration of the boundary should be an integral part of the model.

3.3. Heat of Serpentinization

Serpentinization of peridotite is an exothermic reaction that releases ~2.5×10^5 J kg^{-1} [*Macdonald and Fyfe*, 1985]. Because the heat of serpentinization is of the same order of magnitude as the latent heat of crystallization of magma [*Maclennan*, 2008], it is potentially an important source of heat for driving hydrothermal circulation. There are several factors, however, that inhibit the effectiveness of this reaction as a driver of hydrothermal circulation. First, the reaction occurs at temperatures between approximately 200°C and 400°C [*Martin and Fyfe*, 1970] as hydrothermal fluid is taken up into the rock matrix. Consequently, heat release requires intimate contact between the circulating fluid and the rock mass being reacted. *Lowell and Rona* [2002] developed a simple thermal balance model to estimate the rate of reaction and reaction volumes necessary to drive the recently discovered Lost City vent field on the Mid-Atlantic Ridge [*Kelley et al.*, 2001]. *Lowell and Rona* [2002] argued that if the Lost City vent field were driven by exothermic serpentinization reactions, the rate of reaction would have to range between 10 and 100 kg of peridotite per second. Because of limited fluid access to large rock volumes, *Lowell and Rona* [2002] argued that the lifetime of the current hydrothermal venting episode at Lost City was likely to be between 10^2 and 10^4 years. Subsequently, *Allen and Seyfried* [2004] argued from geochemical data that the heat source for Lost City was likely to be heat mined from crustal rocks rather than heat of serpentinization. Second, nearly all serpentinization reactions involve a substantial volume expansion ranging between 25% and 45% [*Coleman*, 1971]. The effect of such volume changes is difficult to determine. Volume expansion would tend to close fractures at the site of local serpentinization thus restricting access of the fluid to fresh rock. This would slow the rate of serpentinization. But the deformational stresses on nearby rock that would result from volume expansion could promote cracking or faulting that would provide fluid access and serpentinization of nearby rock masses. The overall role of volume expansion is difficult to quantify, but intuitively, the effect would seem to limit the rate of serpentinization and the effectiveness of this reaction as a heat source to drive high-temperature hydrothermal circulation.

One way to quantify the role of exothermic serpentinization reactions is to link these reactions to downward migration of fluid-filled cracks or downward migration of hydrothermal circulation in a reactivated fault zone. This can be done using the model in section 3.2.2 by adding the effective temperature change resulting from the heat released during serpentinization to (12) and (13). Letting ζ denote the heat released during serpentinization, equation (13) then becomes

$$\langle H_r \rangle = 4\lambda_r \left(T_r + \frac{\zeta}{c_r} - T_f \right) \langle v \rangle \left(\frac{\tau}{\pi a_r} \right)^{1/2}. \tag{15}$$

Assuming that $\zeta = 2.5 \times 10^5$ J kg^{-1} [*Macdonald and Fyfe*, 1985], equation (15) increases $\langle H_r \rangle$ by ~50%. This effect is shown in Figure 3. The effects of serpentinization during downward fluid migration do not significantly change the results from section 3.2.2, however.

3.4. Magmatic Heat Transfer

Although hydrothermal circulation at slow spreading ocean ridges may, in principle, be driven by a variety of heat transfer mechanisms, heat transfer from subsurface magma bodies is still the most likely driver of high-temperature hydrothermal venting at slow spreading ridges. It is likely that magma exists beneath the Snake Pit hydrothermal field south of the Kane fracture zone [*Calvert*, 1995; *Canales et al.*, 2000], and a large magma body has been imaged beneath Lucky Strike at 37°N [*Singh et al.*, 2006]. Although recent seismic data did not image a shallow or midcrustal level magma chamber beneath the TAG field (26°N, MAR), the data do not preclude magma at the base of the crust or within the upper mantle [*Canales et al.*, 2007]. Even at the ultraslow spreading Gakkel Ridge in the Arctic Ocean, hydrothermal activity appears to be more abundant than expected and is linked to recent volcanism [*Michael et al.*, 2003; *Baker et al.*, 2004].

Magma-hydrothermal processes can be understood in a basic manner by constructing a simple heat balance model using observed values of T_f, discharge area A_d, heat output H_f, and heat uptake area at the top of the magma chamber A_m from Table 1 as constraints [e.g., *Lowell and Germanovich*, 2004; *Lowell et al.*, 2008]. Consider a single-pass system in which cold seawater near 0°C enters the crust and circulates downward toward the top of a subsurface magma body, flows quasi-horizontally in a heat uptake zone near the top of the magma, and ascends toward the seafloor where the high-temperature vent fluid T_f discharges to the ocean (Figure 4). Heat is conducted from magma initially at its liquidus temperature $T_L \approx 1200$°C across a thin thermal boundary layer of thickness δ. This heat is taken up by the circulating fluid and transported to the seafloor without loss giving rise to the observed hydrothermal heat output H_f. The thermal balance can therefore be written [*Lowell et al.*, 2008]

$$\frac{\lambda_r (T_L - T_f/2) A_m}{\delta} = c_f Q T_f = H_f, \tag{16}$$

where c_f is the specific heat of the fluid and Q is the total mass flux through the deep circulation limb of the hydrothermal cell. By assuming that no conductive heat loss occurs during fluid ascent, the required heat flux from the underlying magma is a lower estimate, and the boundary layer thickness δ is an upper estimate. Even at slow spreading ridges, where the depth to magma is typically greater than at fast and intermediate ones, conductive losses tend to decrease with time as the ascending fluid heats the surrounding rocks and self-insulates the discharge channel [e.g., *Lowell and Rona*, 2002]. This mass flux is driven by the buoyancy difference between the hot discharging fluid and the cold recharge. Assuming that the majority of the flow resistance occurs in the discharge zone, then [*Lowell et al.*, 2008]

$$Q = \frac{\rho_f \alpha_{fd} g k_d T_f A_d}{\nu_d}, \tag{17}$$

where α_f is the thermal expansion coefficient, g is the acceleration due to gravity, k is the permeability, and ν is the kinematic viscosity of the fluid, respectively. The subscript d refers to properties of the discharge zone.

To apply equations (16) and (17) to the hydrothermal systems in Table 1, the value of H_f needs further discussion. Essentially, the heat output of hydrothermal systems has been determined in two ways. One is an "integrated" measurement of the heat output on the vent field scale made in the water column plume above the vent field; the other involves "point measurements" at individual vents that are summed to get the contribution of all the vents in the vent field. The latter measurements often involve extrapolation because heat output measurements are not made at every orifice or on every vent [e.g., *Ramondenc et al.*, 2006]. *Baker* [2007] notes that "integrated" measurements yield higher estimates than "point-wise" measurements. This most likely occurs because some low-temperature diffuse flow that mostly occurs in the extrusive layer 2A is entrained into the water col-

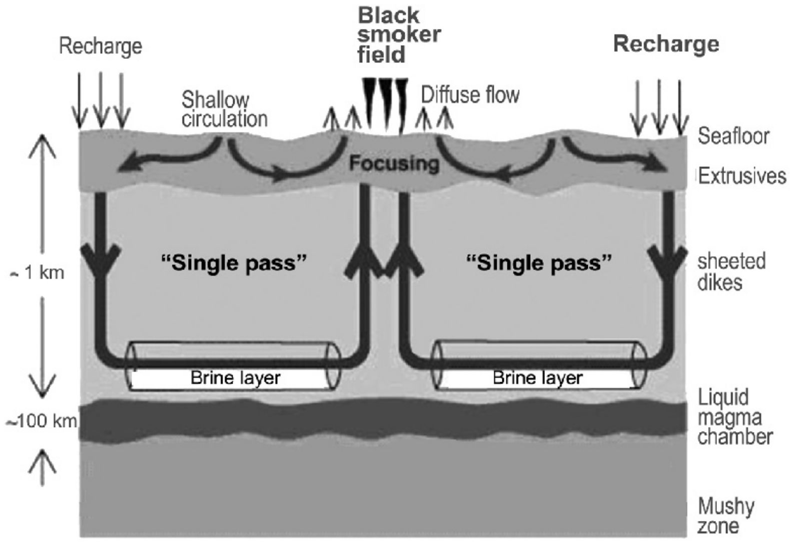

Figure 4. Cartoon of a single-pass hydrothermal circulation cell at a mid-ocean ridge. The single pass refers to the deep circulation system in which fluid circulates downward into the ocean crust, flows more or less horizontally near the top of the subaxial magma chamber at the base of the sheeted dikes, and ascends back to the surface. Heat conducted from the magma is taken up by the fluid in the horizontal limb, where high-temperature water-rock reactions and phase separation occur. Phase separation is denoted by the brine at the base of the circulation cell. Focused high-temperature flow is thought to occur in the deep single-pass limb; diffuse flow may occur as a result of induced flow driven by the high-temperature discharge and mixing of the deep circulation with shallower circulation in the extrusives [from *Lowell et al.*, 2008].

umn plume. Estimates of the ratio of diffuse to focused flow range from 0.5 to 0.9 [*Schultz et al.*, 1992; *Veirs et al.*, 2006; *Ramondenc et al.*, 2006; *Baker*, 2007].

For simplicity, we assume that H_f in equation (16) is either 50% of the integrated heat flux measurement where available or equal to measurements of point-wise heat flux for sites where integrated measurements are not available in Table 1. From equations (16) and (17), one can obtain the values of Q, k_d, and δ for each of the hydrothermal systems in Table 1. The results are given in Table 3. Table 3 shows that the values of Q, k_d, and δ fall within the range of the generic model calculations performed by *Lowell and Germanovich* [2004] and *Lowell et al.* [2008], with the exception of Broken Spur. Broken Spur has exceptionally low high-temperature heat output compared to the other systems. This system may be in a dying phase, or it may be a system that is not driven by magmatic heat input.

A key feature of the simple magma-hydrothermal model presented here is that for high-temperature venting and heat output to be maintained on decadal time scales the rate of heat extraction from the underlying magma must remain nearly constant. Then the conductive thermal boundary layer at the base of the hydrothermal system will remain thin. But as heat is transferred from the convecting magma, it is gradually crystallizing and cooling. Consequently, the Rayleigh number Ra of the convecting magma is gradually decreasing. Because Ra is related to the rate of heat transfer through the relationship $Nu \approx 0.1Ra^{1/3}$, where the Nusselt number Nu

Table 3. Calculated Hydrothermal Parameters for Vent Fields on Slow Spreading Ridges Based on Observables From Table 1

Vent Field	Q (kg s^{-1})	$k \times 10^{-13}$ (m^2)	δ (m)	$u_m \times 10^{-8}$ (m s^{-1})	τ_{2x} (years)
TAG	390	5	24	8.5	38
Broken Spur	6.4	0.26	150	1.4	230
Rainbow	410	6	22	9.1	36
Lucky Strike	950	2	23	9.0	36
Kairei	56	8	17	12	27