The Thermal Structure of the Oceanic Crust, Ridge-Spreading and Hydrothermal Circulation: How Well Do We Understand Their Inter-Connections?

Christopher R. German

Southampton Oceanography Centre, Southampton, UK

Jian Lin

Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, USA

Understanding the complex interplay between the geological processes active at mid-ocean ridges and the overlying ocean through submarine hydrothermal circulation remains a fundamental goal of international mid-ocean ridge research. Here we reflect on some aspects of the current state of the art (and limits thereto) in our understanding of the transfer of heat from the interior of the Earth to the ocean. Specifically, we focus upon the cooling of the upper oceanic crust and its possible relationship to heat transfer via high-temperature hydrothermal circulation close to the ridge axis. For fast- and intermediate-spreading ridges we propose a simple conceptual model in which ridge extension is achieved via episodic diking, with a repeat period at any given location of ca. 50 years, followed by heat removal in three stages: (a) near-instantaneous generation of "event plumes" (~5% of total heat available); (b) an "evolving" period (ca. 5 years) of relatively fast heat discharge (~20%); (c) a decadal "quiescent" period (~75%). On the slow-spreading Mid-Atlantic Ridge, our understanding of the mechanisms for the formation of long-lived, tectonically-hosted hydrothermal vent fields such as TAG and especially Rainbow are more problematic. We argue that both hydrothermal cooling which extends into the lower crust and heat release from serpentinisation could contribute to the required heat budget. Slow-spreading ridges exhibit much greater irregularity and episodic focussing of heat sources in space and time. This focussing may sustain the high heat flow required at the Rainbow and TAG sites, although the detailed processes are still poorly understood.

1. INTRODUCTION

Mid-ocean ridges play an important role in the plate-tectonic cycle of our planet. Extending some 50–60,000 km across the ocean-floor, the global mid-ocean ridge system is the site of the creation of oceanic crust and lithosphere at a rate of ~3.3 km^2 yr^{-1} [*Parsons*, 1981; *White et al*., 1992]. Approximately 75% of the total heat flux from the interior of the Earth (ca. 32 of 43 TW) occurs through this oceanic crust, which covers some two-thirds of the surface of the planet. Much of this heat is released along mid-ocean ridges through complex processes of magma solidification and rapid cooling of the young oceanic lithosphere. While the majority of this heat loss occurs through conduc-

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10.1029/148GM01

tion, approximately one third of the total heat loss at mid-ocean ridges is effected by a convective process termed hydrothermal circulation [e.g., *Stein and Stein*, 1994].

Hydrothermal circulation plays a significant role in the cycling of energy and mass between the solid Earth and the oceans (see syntheses by *Humphris et al*. [1995] and *German and Von Damm* [2003] for detailed reference-lists). This circulation, which is driven by the heat loss at mid-ocean ridges, affects the composition of the Earth's oceans and, indeed, atmosphere. Hydrothermal circulation occurs when seawater percolates downward through fractured oceanic crust along the volcanic mid-ocean ridge system: the seawater is first heated and then undergoes chemical modification through reaction with the host rock as it continues downward, reaching maximum temperatures that can exceed 400°C . At these temperatures the fluids become extremely buoyant and rise rapidly back to the seafloor, where they are expelled into the overlying water column. Active sites of hydrothermal discharge provide an extreme ecological niche that is home to a variety of quite unique chemosynthetic fauna, previously unknown to science [*Van Dover*, 2000]. The first identification of submarine hydrothermal venting and their accompanying chemosynthetically-based communities in the late 1970s remains one of the most exciting discoveries in modern science. A quarter of a century later, however, many of the processes that control the distributions of hydrothermal activity around the global ridge crest, the detailed circulation path for hydrothermal fluids within young oceanic crust and how such processes vary with time, all remain only incompletely understood. The purpose of this volume, which arose from the first InterRidge Theoretical Institute (IRTI) held in Italy in September 2002, is to bring together expertise from two discrete communities, studying mid-ocean ridge geological processes and submarine hydrothermal systems, respectively; these communities had been progressing their research largely independently over the preceding decade.

The first of these groups had been tasked through an InterRidge Working Group to investigate the 4-D (space and time) Architecture of Mid-Ocean Ridges. This research community focused on fundamental geological processes of the building of oceanic crust and lithosphere at ridges through a combination of acoustic and geophysical mapping of the ocean floor, seismic/electro-magnetic imaging, near- and on-bottom geological observations, rock sampling and geochemical/petrological analyses, ocean drilling, laboratory experiments, and theoretical and numerical modelling and syntheses. The highlights of the achievements of this community in the last decade include the direct measurements of the depth and dimension of magma lenses beneath fast-spreading, intermediate-rate, and hotspot-influenced ridges [*Sinha et al*., 1998; *Detrick et al*., 2002a; *Sinha and Evans*, this volume]; complete mapping of volcanic and tectonic fabrics of selected ridge spreading segments at all spreading rates, including the ultra-slow spreading Gakkel Ridge under the Arctic ocean [*Michael et al*., 2003] and the Southwest Indian Ridge [*Sauter et al*., 2002; *Lin et al*., 2002; *Dick et al*., 2003]; discovery of previously unrecognized forms of seafloor morphological features that indicate inherent and complex magmatic/tectonic cycles of seafloor spreading, including amagmatic ridges in ultra-slow spreading ridges and long-lived detachment faults in slow and other magma-starved ridges [*Tucholke and Lin*, 1994; *Cann et al*., 1997; *Tucholke et al*., 1998; *Searle et al*., 2003]; and direct sampling of the mantle rocks along a segment of the Mid-Atlantic Ridge near the Fifteen–Twenty fracture zone [*Kelemen et al*., 2003]. Land-based laboratory and theoretical studies have likewise produced valuable models and testable hypotheses on the focusing of 3-D mantle and melt flows beneath mid-ocean ridges [*Parmentier and Phipps Morgan*, 1990; *Lin and Phipps Morgan*, 1992; *Magde et al*., 1997]; the thermal state of the oceanic crust as a function of magma supply and hydrothermal circulation at ridge crests [*Phipps Morgan and Chen*, 1993; *Shaw and Lin*, 1996; *Chen and Lin*, 2004, *Chen*, this volume]; integrated petrological models on the compositional variations among mid-ocean ridge crustal and mantle rocks [*Langmuir et al*. 1992; *Michael and Cornell*, 1998]; predictions of the varying styles of seafloor faulting and earthquakes and their dependence on the degree of tectonic extension and rock types [*Searle and Escartin*, this volume]; the causes of the dramatic differences in ridge crest topography between fast- and slow-spreading ridges; and the potentially important role of water and/or serpentinites in controlling the rheology and, thus, the tectonic deformation of the oceanic lithosphere [*Hirth et al*., 1998; *Searle and Escartin*, this volume].

The second community involved in the development of both the IRTI workshop and the ideas presented in this volume were co-ordinated through the InterRidge Working Group studying Global Distributions of Seafloor Hydrothermal Venting. Previously, a state-of-the-art volume detailing many of the interacting subdisciplines of modern hydrothermal research had been compiled by *Humphris et al*. [1995] based on a US Ridge Theoretical Institute addressing physical, chemical, biological and geological interactions in seafloor hydrothermal venting. More recently, *German and Von Damm* [2003] have synthesised core information available at the time of publication of that earlier volume, together with new information that has only been obtained during the intervening decade. There have been two key discoveries in this time that have quite revolutionised our assumptions about seafloor venting to the oceans. The first of these has been the recognition that temporal evolution of any given seafloor hydrothermal system can vary profoundly [e.g., *Von Damm*, this volume] while the second is that hydrothermal activity may be far more widespread than had previously been recognised occurring along ridges of all spreading rates and, hence, being present in all ocean basins [e.g., *Baker and German*, this volume]. For the decade following discovery of high-temperature venting, what was remarkable was the constancy in temperature and composition of erupting vent fluids even though individual sites might exhibit quite distinct compositions. The best example of this constancy was the first high-temperature system discovered at 21°N on the East Pacific Rise, which appears to have remained unperturbed for ~20 years [*Von Damm et al*., 2002]. Another example is the TAG hydrothermal field at the Mid-Atlantic Ridge which has thus-far maintained a near-constant end-member fluid composition from 1986 until 2003, even after being subject to ODP drilling [*Edmonds et al*., 1996]. What has become apparent in the past decade, however, is that such constancy may not be the norm. Both the temperature and composition of fluids exiting any given vent site can change significantly with time, apparently in direct response to volcanic extrusion and/or dike intrusion beneath the seafloor [e.g., *Butterfield and Massoth*, 1994; *Butterfield et al*., 1997; *Von Damm et al*., 1995, 1997]. Perhaps the best-studied such site, to date, is that at 9°50'N on the East Pacific Rise, which last underwent a volcanically eruptive episode in April 1991 [*Haymon et al*., 1993]. A key question of active current debate is: what has controlled the continuing evolution of fluid temperatures and compositions at this site over the subsequent decade?

A major discovery in the past ten years of hydrothermal research has been the recognition that hydrothermal activity can occur in reasonable abundance along slow-spreading ridges [e.g., *German et al*., 1996b] as well as along faster ridges that are supplied by significantly greater magmatic heat fluxes. Previously, *Baker et al*. [1996] had predicted that the abundance of venting along any section of ridge crest should scale directly with the available magmatic heat flux and, hence, with ridge spreading rate. The discovery of hydrothermal activity along some of the world's slowest spreading ridges – notably the SW Indian Ridge [*German et al.*, 1998; *Bach et al*. 2002] and the Gakkel Ridge in the Arctic Ocean [*Edmonds et al*., 2003] – appears to contradict, or at least modify, this hypothesis. Where does this "additional" hydrothermal flux originate? Recent discoveries on the Mid-Atlantic Ridge have revealed a new class of tectonically-hosted vent-sites including the low-temperature Saldanha and Lost City sites [*Barriga et al*., 1998; *Kelley et al*., 2001] and the high-temperature Logatchev and Rainbow hydrothermal fields [e.g., *Charlou et al*., 2002]. The latter, in particular, appears to impart exceptionally high fluxes of heat and chemicals into the surrounding ocean [*Thurnherr & Richards*, 2001; *German et al*., submitted].

Considering the above, the timeliness of this volume becomes self-evident. While the important role of hydrothermal circulation in regulating ocean (hence, atmospheric) compositions may be well established, what remains unclear is how that hydrothermal circulation is regulated itself. This is not something that can become understood just by studying hydrothermal flow alone. Instead, an integrated approach is required in which the transfer of heat throughout the ocean-ridge system is taken as the key master variable that drives all other aspects of ridge and hydrothermal interactions during its transfer from the interior of the Earth. That is the overarching theme of the chapters that follow. In the remainder of this chapter, however, we use some key case studies to review our current understanding of the processes that sustain these hydrothermal systems and to highlight key problems where our level of understanding can be described as incomplete, at best.

2. CASE STUDY I: THE FAST-SPREADING NORTHERN EAST PACIFIC RISE AND THE INTERMEDIATE-RATE JUAN DE FUCA RIDGE.

2.1. The East Pacific Rise at 9°–10°N

The East Pacific Rise (EPR) between 9° and 10°N has become one of the most intensively studied sections of ridge crest anywhere on Earth due primarily to the attention received during the US RIDGE programme over the past decade or more – a commitment renewed with the selection of this area as an integrated study site that will continue to be studied in particular detail during the new RIDGE 2000 initiative (http://r2k.bio.psu.edu). Of interest to the discussion we present below is that this region also lies within the northern most limits of a NOAA Acoustic Hydrophone Array (AHA), which continuously monitors seismic activity along the entire ca. 2,000 km of the EPR between 10°S and 10°N [*Fox et al*., 2001; http://www.pmel.noaa.gov/vents/acoustics.html].

The ridge at this location has a full spreading rate of 110 mm/yr [*Carbotte and Macdonald*, 1992] and is underlain by a seismic low-velocity zone, interpreted to represent a melt lens at a depth of ca. 1.5 km below the seafloor [*Detrick et al*., 1987]. The area attracted particular attention from hydrothermal researchers in the early 1990's after it was recognised that an episode of dike emplacement and volcanic extrusion had occurred toward the northern end of the previously surveyed region at 9°09'–9°54'N in April 1991 [*Haymon et al*., 1991, 1993]. Time series studies of vent-fluid compositions at this site (near 9°50'N) have subsequently continued to show evolving chemical compositions, in direct response to this volcanic episode, over more than a decade [e.g., *Von Damm et al*., 1995, 1997; *Von Damm*, this volume].

2.2. Magmatic Heat Release Is Episodic

When considering the steady-state magmatic heat flux available to this section of ridge crest, one approach is to take that described by *Sinha and Evans* [this volume], which predicts a total magmatic thermal flux of 5.0×10^{10} Watts along the 600 km of global ridge crest that are spreading at a full rate of ~110 mm/yr. This yields a unit heat flux at the EPR 9–10°N of 8.3×10^{7} W/km. However, the value calculated using that approach represents the total heat emplaced magmatically within the full 6 km thickness of fresh-formed oceanic crust. If we only consider the heat emplaced by diking above an EPR magma lens (at a depth of 1.5 km), the heat flux delivered immediately below the seafloor and right at the ridge axis reduces to ca. 25% of the total predicted by *Sinha and Evans* [this volume], i.e., ca. 2.1×10^{7} W/km.

Let us now consider a situation in which all of this upper oceanic crust heat flux (conveniently expressed as 21 MW/km or 210 MW/10km) was dissipated by high-temperature hydrothermal circulation that is focussed close to the ridge axis. Reported heat fluxes from individual high-temperature focussed-flow vent-sites typically fall in the range 10–100MW [e.g., *Speer and Rona*, 1989; *Bemis et al*., 1993; *Lupton*, 1995; *Rudnicki et al*., 1994; *German et al*., 1996a; *Rudnicki & German*, 2002]. While integrated heat-fluxes including diffuse flow may often exceed this value [e.g., *Lowell & Germanovich*, this volume] our focus, initially, is solely upon high-temperature "black smoker" flow. Here, therefore, we assume that a representative heat-flux from any given "black smoker" hydrothermal field is of the order 100 MW. In that case, the steady-state heat flux associated with dike emplacement would appear sufficient to sustain one such hydrothermal field approximately every 4–5 km along the ridge axis. While identification of individual vent sites is not straightforward at areas recently perturbed by volcanic eruptive processes [e.g., *Von Damm*, this volume], such a high frequency of venting appears greater than what is observed along more "quiescent" sections of the EPR where a mean spacing of ~8–15 km between adjacent vent sites appears to be closer to the "norm" [e.g., *Fornari et al*., this volume]. Thus, what is observed at nominal "steady-state" heat flux on the northern EPR appears to be a factor of ~2–3 lower than the above prediction. It would appear, therefore, that the "quiescent" mode of black smoker hydrothermal flow—as perhaps best exemplified by that witnessed at the EPR 21°N vent sites throughout the past 23 years—may only account for some ca. 33–50% of the steady-state heat flux available to support hydrothermal circulation through the upper oceanic crust, close to the axis of fast-spreading ridges.

What, then, is the fate of the "missing" hydrothermal heat flux not accounted for in the above calculations? One possibility is that it is manifest, at steady state, in the form of diffuse hydrothermal flow that is not incorporated within the entrainment process close to black smoker hydrothermal plumes and, hence, remains unsampled by "plume-integration" heat-flux studies [e.g., *Converse et al*., 1984]. What also merits consideration, however, is whether there is a significant component of hydrothermal flux that occurs in close association with periods of dike intrusion and/or eruptive activity at the ridge axis that is not well represented – indeed, perhaps completely overlooked – in estimates based on steady-state "quiescent" stages of hydrothermal discharge. A major barrier to evaluating the possible importance of this issue, to date, has been our lack of understanding of the episodicity with which dike emplacement and/or volcanic eruption should be expected to recur at any given location along the global ridge crest. While the time-averaged spreading rates of all mid-ocean ridges are reasonably well constrained [e.g., *DeMets et al*., 1994], what remains elusive is any clear understanding of the size and frequency of the "quantum events" by which mid-ocean ridges achieve this time-averaged growth.

2.3. Dike Emplacement: The "Quantum Events" of Fast-Spreading Ridges?

An important new development in understanding mechanisms of plate-spreading at mid-ocean ridges has come from the use of the NOAA Acoustic Hydrophone Array (AHA), which has now provided data monitoring episodes of seismic activity along the EPR from 10°S to 10°N since 1996 [*Fox et al*., 2001]. During the five years of 1996–2001, for example, "clusters" of seismic activity have been observed on five separate occasions on the EPR axis centred at latitudes of ~8°39'N, 7°17'N, 3°25'N, 2°46'S, and 5°46'S (marked by yellow ellipsoids in Plate 1). Each of these hydroacoustically determined event clusters was located toward the centre of a different second-order volcanic segment of the EPR rather than on the transform faults like most of the hydroacoustic events. The seismic swarm near the 8°39'N area occurred on March 2, 2001, where evidence for a hydrothermal plume was subsequently identified through a "rapid response" effort reported by *Bohnenstiehl et al*. [2003]. Because these seismic swarms occurred right at the axis of a fast-spreading ridge, where the brittle thickness of the lithosphere is expected to be thin, we hypothesize that these seismic swarms are associated with magmatic emplacement rather than tectonic faulting—such interpretations have yet to be confirmed from independent observations, except at 8°39'N.

If our interpretation is correct, however, it would imply that each of these seismic swarms is indicative of a fresh episode of dike intrusion. This dike intrusion may also be accompanied by volcanic extrusive activity at the seafloor, although the lat-

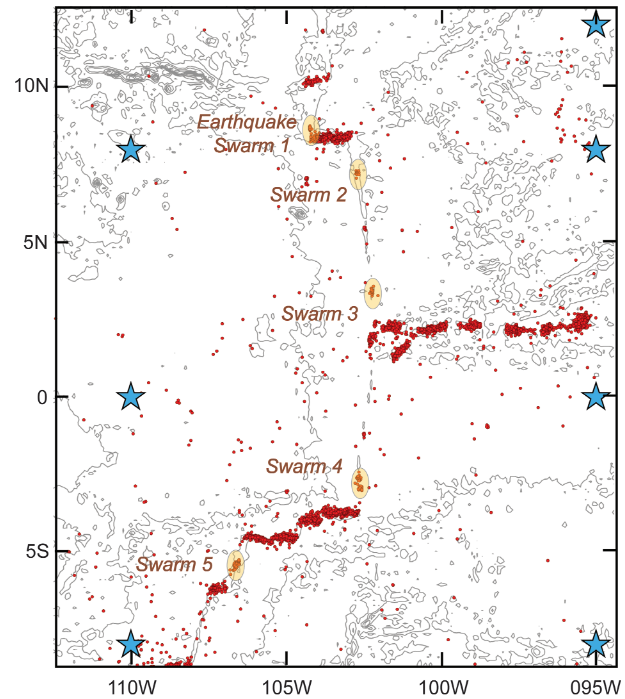

Plate 1. Map showing earthquakes on the East Pacific Rise during 1996–2001. Black thin lines show bathymetric contours. Blue stars mark the location of 7 NOAA hydroacoustic sensors. Small red dots show earthquakes determined by the NOAA hydroacoustic array during 1996–2001. Original earthquake data are from Fox et al. [2001] and the plot is modified from Gregg et al. [2003]. Yellow ellipsoids denote locations of five clusters of earthquake swarms on the EPR axis away from major transform faults. We interpret these earthquake swarms to be potentially associated with diking events on the EPR, although only the event cluster near 8°39'N has been confirmed to be associated with hydrothermal discharge [Bohnenstiehl et al., 2003].

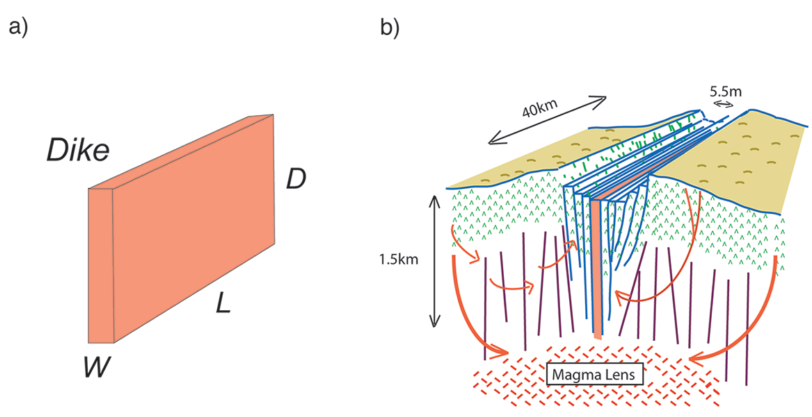

Plate 2. a) Simple schematics of a "model" dike intrusion measuring W = 5.5 m, D = 1.5 km and L = 40 km as used in our calculations; b) illustration [after Alt, 1985] showing possible routes of hydrothermal circulation (arrows) through a section of fast- or medium-spreading ridge crest and an idealised dike emplacement (vertical orange slab) which rises 1.5 km above an axial magma lens, measures 5.5m in width and extends 40km along-axis. The heat release from crystallization and cooling of a dike above an axial melt lens may be sufficient to generate "event plumes" (days–weeks), release heat rapidly through "evolving" hydrothermal systems (months–years) and yet still maintain more "quiescent" high temperature hydrothermal flow over timescales of decades.

ter is not required to have occurred. The interpretation is based on a comparison of this recent seismic activity along the EPR 10°S–10°N section with that reported from ground-truthed seafloor volcanic eruption events witnessed previously, both by submersible and using the SOSUS array, along the Juan de Fuca Ridge [e.g., *Dziak et al*., 1995; *Fox and Dziak*, 1998; *Dziak and Fox*, 1999]. The fact that the ground-truthed seismic swarm at 8°39'N was, indeed, accompanied by hydrothermal discharge not previously reported [*Bohnenstiehl et al*., 2003] encourages us in the validity of our approach. Another important calibration point for the NOAA hydroacoustic data set is obtained from the recognition that no significant seismic swarm activity has been reported from the segment at 9–10°N on the EPR throughout the past 5 years – an area where repeat Alvin dives can confirm that no further volcanic eruptions have, indeed, occurred.

Within the AHA survey area, an average occurrence has been recorded of one seismic swarm per year, with each such swarm—assumed here to be indicative of volcanic eruptions—appearing to extend over some ca. 30–50 km along axis. For the whole surveyed area (from 10°S to 10°N, ca. 2,200 km in total distance) to undergo one such "quantum event" of ridge spreading, a series of 45–75 episodes, each measuring 50–30 km in extent, would be required. This suggests that, on average, a repeat event for dike-intrusion/volcanic-extrusion at any one vent site on the fast-spreading EPR should only occur approximately once every 50 years. To a first-order approximation, this time scale does not appear inconsistent with much that is already known from the northern East Pacific Rise: first, that hydrothermal activity can continue unperturbed over decadal time-scales, e.g., at 21°N EPR over 20–25 years [*Von Damm et al*., 2002]; second, that recently perturbed sites (e.g., 9°50'N) can continue to evolve/relax back toward relatively stabilised conditions over a period of approximately one decade following dike emplacement/volcanic eruption [*Von Damm*, this volume; *Von Damm et al*., submitted].

If episodic spreading, as described above, does dominate the growth of oceanic crust at ridge axes and, further, if such events do only recur once every ~50 years (range 45–75 years) then, for a full spreading rate of 110 mm/yr [*DeMets et al*., 1994], each such spreading event on the northern EPR should lead to an "instantaneous" extension, across the ridge axis, of ~5–8 m. This figure is also in reasonable agreement with what has previously been observed. For example, *Wright et al*. [1995] conducted a detailed study of seafloor images at the EPR 9°50'N site and showed that active hydrothermal venting occurred in close association with cracking at the fourth-order segment scale. These features were interpreted as "eruptive" cracks because they were sited directly within fresh lava-flows. Maximum crack widths in that study measured up to 7 m and the majority (even after partial infilling of individual fissures by lava flows) fell into a range of ca. 2–4 m extension across strike. In the extreme, those authors argued for up to 20% extension across the axial zone of the ridge crest, effected through eruptive fissuring, during the 1991 episode of volcanic extrusion.

A final consideration is the volume of magma delivery that would be required to generate a single diking event, should this represent the "quantum event" of plate-tectonic spreading at fast-spreading ridge axes. *Gregg et al*. [1996] have estimated that the volume of extrusive flow alone at EPR 9°N associated with the 1991 eruption was ca. 1 x10^6 m^3. Assuming a depth of 1.5 km to the top of the underlying magma lens along this section of the EPR [*Detrick et al*., 1987], emplacement of a dike measuring 5–8 m wide, striking along the ridge axis would require a much greater volume of magma that measured 5–8 m x 1.5 km x 1 km = 7.5–12 x10^6 m^3 to be supplied for every 1 km of along-axis extent of that dike (Plate 2). When compared with the dimensions of an axial magma lens, which measures ca. 100 m thick and some 2 km wide across axis [*Kent et al*., 1990; *Singh et al*., 1999], however, we see that this dike volume could readily be supplied by the extraction of just ca. 5% of the volume of a magma lens in a single eruptive episode. This would be followed by some decades of magma chamber replenishment before any further magmatic extraction would be expected to recur. A final calculation shows that this volume of lava could be supplied from the magma chamber to the dike over a time frame of ca. 35 days assuming comparability with observed magma supply rates of 3–3.5 m^3/sec, as reported previously from Kilauea, Hawaii [*Tilling and Dvorak*, 1993] and Krafla, Iceland [*Einarsson*, 1978; *Trygvasson*, 1980]. It does not seem unreasonable, therefore, to conclude that the dike-emplacement mechanism should indeed serve as a reasonable approximation for mechanisms of ridge growth on fast- and intermediate-spreading ridges, wherever an axial magma lens has been identified to occur.

2.4. Heat Released by "Quantum" Diking Events

Having established the frequency with which dike-emplacement/volcanic eruptions might be expected to recur at any given ridge crest location, we can now return to the problem of whether a significant hydrothermal heat flux might be associated with such perturbations. In the previous section we estimated that emplacement of a single dike, extending 30–50 km along axis, should occur approximately once every 50 years at any given section of the fast-spreading northern EPR. Further, for the full-spreading rate of 110 mm/yr at the NEPR, each such diking event should lead to a broadening of the ridge axis by ca. 5.5 m (with a range of 5–8 m). For simplicity, we initially consider the case for an idealised dike that strikes 40 km along-axis, exhibits a constant across-axis width of 5.5 m, and extends 1.5 km down to the top of the underlying magma lens (Plate 2). The heat that would be released through cooling of this dike from its emplacement temperature to a reference "hydrothermal" temperature of 350°C [c.f., *Sinha and Evans*, this volume] would consist of two components: the latent heat released due to crystallization of the basaltic magma and the heat released during cooling of that solidified basaltic material from 1,250°C to the reference temperature of 350°C [e.g., *Sinha and Evans*, this volume].

2.4.1. Latent heat released due to crystallization of basaltic magma. The heat that would be released from crystallization of the lavas within our idealised dike at *in situ* emplacement temperatures, $E_{crystallization}$, can be calculated as the simple product of the dike volume V_{dike}, the density of the magmatic material undergoing crystallization ρ_{crust}, and the latent heat of fusion for basaltic material H_f. Thus:

$$E_{crystallization} = V_{dike} \times \rho_{crust} \times H_f$$

For the idealised dike described above, V_{dike} = 5.5 m x 1.5 km x 40 km = 3.3 x 10^8 m^3. Then, using values of ρ_{crust} = 2,750 kg m^{-3} and H_f = 5 x 10^5 J kg^{-1} [*Cannat et al*., this volume; *Sinha and Evans*, this volume], we obtain a total heat release predicted to accompany crystallization, $E_{crystallization}$ = 4.54 x 10^{17} J, which equates to $E'_{crystallization}$ =1.14 x 10^{16} J/km along axis.

2.4.2. Heat released during cooling of solidified basaltic material, 1,250°C to 350°C. To determine the heat released during cooling of the solidified basaltic dike, a separate calculation is required in which the total energy available $E_{cooling}$ is a product of the dike volume V_{dike}, the density of the basaltic matter ρ_{crust}, the specific heat of basaltic crust $C_{p\ crust}$, and the temperature change ΔT for which the calculations are being conducted. Thus:

$$E_{cooling} = V_{dike} \times \rho_{crust} \times C_{p\ crust} \times \Delta T$$

Again, the assumed dike volume for these calculations is V_{dike} = 3.3 x 10^8 m^3 and the appropriate basalt density ρ_{crust} = 2,750 kg m^{-3}. Then, for a value of $C_{p\ crust}$ = 1,050 J kg^{-1} K^{-1} and taking ΔT = 900 K, which is the temperature change associated with cooling the magmatic body from 1250°C to the 350°C reference temperature [*Sinha and Evans*, this volume], we calculate a "cooling" heat release that totals $E_{cooling}$ = 8.58 x 10^{17} J, which equates to ca. $E'_{cooling}$ = 2.15 x 10^{16} J/km along axis. Thus, in sum, the total heat released from this idealised diking event should be $E_{dike} = E_{crystallization} + E_{cooling}$ = 1.31 x 10^{18} J, which is equivalent to E'_{dike} = 3.23 x 10^{16} J/km along the full 40 km strike of this dike.

2.5. The CoAxial Segment of the Juan de Fuca Ridge

The same calculations illustrated above can also be repeated for a differing, ~40 km idealised dike emplacement on the Juan de Fuca Ridge with a full-spreading rate of 70 mm/yr. There, the representative "quantum" dike width for the equivalent of 50 years' extension would be reduced to 3.5 m, consistent with reported estimates based on studies of the CoAxial segment of the Juan de Fuca Ridge [3–5 m: *Cherkouai et al*., 1997; 2–4m: *Baker et al*., 1998]. Another important difference is that for the Juan de Fuca Ridge calculations we increase the depth to the top of the axial magma chamber to 2.5 km. This assumption is based on a recent multi-channel seismic reflection experiment showing that the depth to the top of the magma lens varies in the range 2.3–2.8 km for various ridge segments of the Juan de Fuca Ridge [*Canales et al*., 2002; *Carbotte et al*., 2002; *Detrick et al*., 2002b]. With these variations in our idealised dike volume, the equivalent total amount of heat available remains remarkably similar (presumably coincidentally so) with a value of E_{dike} = 1.39 x 10^{18} J. This value for the Juan de Fuca Ridge is of particular interest because *Baker et al*. [1998] carried out extensive hydrothermal heat-flow determinations during the 2–3 year period following the June 1993 seafloor dike-intrusion event in the CoAxial segment. We now have a framework within which to compare our model with independent studies of the hydrothermal flow apparently stimulated by that event.

2.6. Three Stages of Heat Release Associated with a Diking Event

2.6.1. The "Event-Plume" Period. In their study at the Juan de Fuca Ridge, *Baker et al*. [1998] were able to calculate the heat released into "event plumes", immediately after the dike emplacement event. They then compared those fluxes with the flow of heat (and chemicals) released during a phase of rapidly diminishing "chronic" (black-smoker type) high-temperature hydrothermal plume discharge, which continued over a further 2–3 years. Although *Baker et al*. [1998] calculated that the total heat content of the three event plumes detected within days to weeks of dike emplacement was high, $E_{event\ plume}$ = ca. 2 x 10^{16} J, they also concluded that this contribution to the total flux at the CoAxial segment was small when compared to the integrated heat flux released through the diminishing "chronic" hydrothermal discharge monitored from plume surveys over the subsequent 2–3 years. The estimated heat fluxes from this region dropped dramatically from values in the range 20–40 GW during the first few weeks after the diking event (July –August 1993) to values significantly less than ~1 GW by June 1995. Integrating under the curve for measured decreasing heat flow over this time period, *Baker et al*. [1998] calculated a total heat flow from their 2–3 year study of decreasing hydrothermal plume discharge to represent a total energy release of ca. $E_{evolving}$ = 3.5 x 10^{17} J. This is approximately one order of magnitude greater than was calculated for the initial event plume systems and led *Baker et al*. [1998] to conclude that event plume discharge, released at the instant that a dike is emplaced into the oceanic crust, could not account for more than ca. 5% of the total rapidly-released (≤2 years) heat flow associated with such a diking event.

2.6.2. The "Evolving" Period. Baker et al. [1998] found that the heat and fluid flow at their two study sites on the CoAxial segment not only diminished but ceased almost completely over 2–3 years. This is in contrast to the observations from the immediately adjacent Cleft segment of the Juan de Fuca Ridge, where sustained venting has been observed over more than 11 years [e.g., *Baker*, 1994]. We believe that this discrepancy can be resolved by the observation that the rapidly-diminishing "chronic" or "evolving" plume discharge monitored by *Baker et al*. [1998] at the CoAxial segment between 1993 and 1996 ($E_{evolving}$ = 3.5 x 10^{17} J) was significantly less than the total heat available for cooling the entire dike to the top of the magma chamber (E_{dike} = 1.39 x 10^{18} J). The amount of heat reported by *Baker et al*. [1998] to have been released rapidly, in the 2–3 years following dike emplacement, equates to no more than ca. 25% of the total heat available from crystallizing and cooling of the entire volume of the dike to hydrothermal temperatures of 350°C. From one perspective, therefore, this rapid cooling phase following dike emplacement might be seen as equivalent to the heat released from cooling of the uppermost ~25% of the dike in volume, i.e., to a depth of ca. 600–650 m. Again, this is in good agreement with *Baker et al*. [1998]'s preferred explanation that they witnessed cooling of a dike to a depth of ~500 m. The remaining ca. 75% of the heat from the diking event ($E_{quiescent}$ = 1.04 x 10^{18} J), then, must be released much more slowly, over subsequent decades.

2.6.3. The "Quiescent" Period. Comprehensive plume survey data for "event plumes" and "chronic" heat release are only available for the CoAxial segment of the Juan de Fuca Ridge. If the results of these studies are taken to be representative of the partitioning associated with all dike-emplacement events we can now calculate what further amount of heat should also be available that is not released through vigorous, but rapidly evolving venting in the years immediately following a dike event. Here we estimate the potential for longevity of such heat flow if it is extracted over a decadal time scale in the form of less-vigorous, relatively "quiescent" black-smoker hydrothermal flow. For the Juan de Fuca Ridge, our calculations estimate that ca. 25% of the total heat available from a dike emplaced above an axial magma lens (E_{dike} = 1.39 x 10^{18} J) was released by rapid heat removal (including event plumes) over a time scale of a few years ($E_{event\ plume\ +\ evolving}$ = 3.5 x 10^{17} J). For the EPR, the same 25% proportion would represent a similar total heat content of $E_{event\ plume} + E_{evolving}$ = 0.25 x 1.31 x 10^{18} J = 3.3 x 10^{17} J. Subtracting this from the total heat emplaced by a 5.5 m wide, 40 km long dike at the EPR would yield a residual heat content, $E_{quiescent}$ = 9.8 x 10^{17} J. If this residual heat is extracted in the form of prolonged, but less vigorous, high-temperature hydrothermal venting, it would be sufficient to sustain a total hydrothermal heat flux of 800 MW along a 40 km section of the EPR for approximately 50 years, which is the estimated time period between repeat diking events. We note that such "quiescent" high-temperature hydrothermal heat flow has been observed along the northern EPR for ≥20 years [e.g., 21°N: *Von Damm et al*., 2002]. At a mean spacing of one hydrothermal vent site every 8–15 km along axis (see earlier), this predicted long-term high-temperature hydrothermal heat flux of 800 MW along a 40 km section of the ridge crest could easily be accommodated by, for example, 4 vent sites each with a power of 200 MW persisting over 50 years, or 5 such sites at 160 MW each persisting over the same time interval.

Thus we suggest a conceptual model (Plate 3) in which dike emplacement is accompanied by a minor proportion (ca. 5%) of the total available heat being released quasi-immediately (days) into "event plumes" followed by a more significant rate of energy release associated with rapidly-diminishing flow (order 10s of GW to < 1GW) through "chronic" or "evolving" hydrothermal plumes, in which some further ~ 20–30% of the total heat emplaced by diking might be released on the order of 5 years. This would then be followed by a much more "quiescent", extended period of heat flow (ca. 50 years) during which the remaining ca. 60–80% of the heat emplaced by diking was discharged through long-term high-temperature hydrothermal flow. Such hydrothermal flow could be emitted through relatively low-power hydrothermal vent systems spaced at regular intervals, ca. 10 km apart, along the ridge axis. Encouragingly, this conceptual model appears to be reasonably consistent with field data for both long-term and rapidly evolving high-temperature hydrothermal systems along the axes of the fast-spreading northern EPR and the intermediate-spreading Juan de Fuca Ridge. At sites where dike emplacement is known to have occurred (EPR 9°50'N and CoAxial segment of the Juan de Fuca Ridge), there has been rapid evolution of the hydrothermal system after volcanism has ceased, with flow diminishing rapidly over a matter of a few years and certainly less than a decade [*Baker et al*., 1998; *Von*

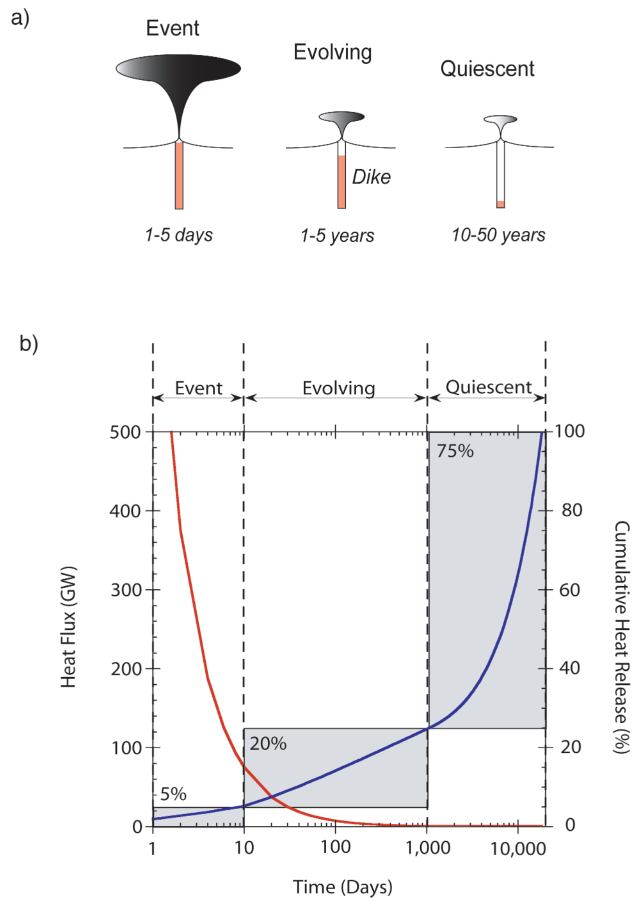

Plate 3. a) Cartoon illustration of our conceptual model, and b) graphical representation of the same model for the heat flux (red curve) and corresponding cumulative heat release (blue curve) as a function of time following dike emplacement at a fast- or medium-spreading ridge crest. Here we identify three time-periods. During the "event" period (measured in days–weeks) event plumes form releasing large quantities of heat into the water column. The "evolving" period (months–years) is characterized by continuing rapid discharge of heat via vigorous hydrothermal venting which evolves over a period of years, up to a decade. This is followed by the "quiescent" period (more than one decade), during which sustained, less-vigorous venting dominates the cumulative heat-flux from the system. Here we discuss only heat fluxes associated with such a progression from "event" to "evolving" to "quiescent" flow. A more detailed treatment of the changes in chemical compositions of the fluids during these different stages is presented by Von Damm [this volume] for the case of EPR 9°N. More general discussions are presented by German & Von Damm [2003] and references therein.

Damm, this volume]. At sites where no such dike emplacement has been observed, however, relatively low-power hydrothermal vent sites (ca. 100 MW each) have apparently continued relatively unperturbed over periods extending over decadal time-scales [e.g., Cleft segment of the Juan de Fuca Ridge: *Baker*, 1994; EPR 21°N : *Von Damm et al*., 2002].

2.7. Potential Deficiencies in This Simplified "Unifying" Model

Above, we have used a thermodynamic approach to generate a conceptual heat-balance model that can plausibly explain the inter-relationship between heat supply from magmatic emplacement at ridge crests and heat removal through high-temperature hydrothermal flow. In the next section we use the same approach to highlight where difficulties arise when one attempts to apply this same simplistic model to slower-spreading ridge crests. Before addressing that point, however, we must point out two key issues that the above, simplified model has overlooked.

First, to construct a simple heat-balance alone and yet not consider the mechanisms by which that heat might be extracted from crust to ocean, is to ignore a large body of work conducted into the mechanisms by which hydrothermal flow takes place through a porous/permeable medium. This work was commenced more than 20 years ago [e.g., *Strens and Cann*, 1982; *Lowell and Rona*, 1985] and is reviewed in detail by *Lowell and Germanovich* [this volume]. To briefly illustrate the point, the availability of the appropriate amount of heat in a system does not necessarily imply a mechanism by which fluid can then be extracted at a particular preferred temperature (e.g., the 350°C of a high-temperature "black smoker") over any designated amount of time. For example, while extracting heat through hydrothermal circulation from the top of a freshly-emplaced dike, the rest of that dike will also be cooling by conduction into the host surrounding rock. The standard conductive time scale for this process is $\tau \sim W^2/\kappa$ where W is dike width and thermal diffusivity $\kappa \sim 10^{-6} m^2/sec$. For the extreme case of conduction into cold rock (i.e., maximum heat removal not involving hydrothermal circulation) this yields $\tau \sim 1$ year for a 5.5m-wide dike (see above) and implies virtually complete cooling to background temperatures in just one decade – a time scale much less than the 50 years predicted by our calculations. [See *Lowell & Germanovich*, this volume, for more detailed discussions on this topic].

A second important issue is that the calculation in the previous paragraph, for dike-cooling by conduction, also appears to converge with the most recent field data from the northern East Pacific Rise. Although the EPR 21°N site has remained stable over more than a decade of measurements [cf., *Von Damm et al*. 1985, 2002] the most recent report indicates that some change in composition may finally be occurring. Even more relevant, *Von Damm* [this volume] concludes from inter-comparison of fluid compositions and temperatures at the Bio9, Bio9' and P vents at EPR 9°50'N that fresh magmatic replenishment may have already recommenced at this location and fresh eruptions may be imminent, within little more than ten years of the prior eruptions in 1991. Clearly, such activity would be completely inconsistent with the model we have presented above: we have constructed an eminently testable hypothesis which the EPR 9°N Integrated Study Site of the Ridge 2000 programme should be uniquely well placed to test in the decade ahead. If *Von Damm* [this volume] is correct, then we may not have long to wait to see the results.

3. CASE STUDY II: THE SLOW-SPREADING NORTHERN MID-ATLANTIC RIDGE

If we set aside the mechanistic shortcomings of the "quantum event" model discussed above, it remains the case that our heat-budget approach, which could explain the links between crustal growth and hydrothermal cooling at fast- and intermediate-spreading ridges, appears to bear little or no relevance to slow-spreading ridges, particularly in the instances of large discrete hydrothermal fields such as those found at the TAG [e.g., *Rona et al*., 1986] and Rainbow [*Charlou et al*., 2002] sites on the northern Mid-Atlantic Ridge. In this section we examine processes that may be active along such slow-spreading ridges.

3.1. The TAG Hydrothermal Field: An Unusually Large High-Temperature Vent System

To date, a total of seven sites of high-temperature hydrothermal venting have been identified along the slow-spreading Mid-Atlantic Ridge (MAR) together with the low-temperature Lost City hydrothermal field [*Kelley et al*., 2001]. Three of the high temperature fields have all been found in close proximity to one another close to the Azores Triple Junction: Menez Gwen (38°N), Lucky Strike (37°N) and Rainbow (36°N). To the south are the Broken Spur (29°N), TAG (26°N) and Snake Pit (23°N) sites, which are followed at 15°N by the Logatchev hydrothermal field [e.g., *Baker and German*, this volume]. The first site of high-temperature hydrothermal venting to be located on the MAR was the TAG hydrothermal field at 26°N [*Rona et al*., 1986], which was long thought to be particularly anomalous. The next sites of venting to be discovered, the Snake Pit and Broken Spur vent fields, are both directly associated with axial summit fracturing along active volcanic ridges. These two vent fields reveal both hydrothermal deposits and volcano-tectonic settings [e.g., *Fouquet et al*., 1993; *Murton et al*., 1995] that are quite reminiscent of high-temperature hydrothermal vents along the fast-spreading EPR [e.g., *Wright et al*., 1995]. The TAG hydrothermal field, in contrast, is much larger than the majority of seafloor hydrothermal systems and is hosted in tectonically-controlled terrain close to the base of the east wall of the rift valley, i.e., away from the active ridge axis [*Kleinrock and Humphris*, 1996]. The site of active venting at TAG is atop a 50 m tall, 150 m diameter mound of hydrothermal sulfide and sulfate deposits. This mound is much larger than most seafloor hydrothermal settings but comparable to numerous volcanic-hosted massive sulfide deposits found on land including some of the larger sulfide deposits from the Troodos ophiolite, Cyprus [e.g., *Hannington et al*., 1998].

Humphris and Cann [2000] have calculated the total energy that would be required to construct the TAG hydrothermal mound based primarily upon the Fe budget of the TAG system. (A similar although simpler, estimate of this type is also presented by *Lowell & Germanovich* [this volume]). Those calculations indicate that formation of the TAG hydrothermal field requires a total input of $E_{TAG} = 1–2 \times 10^{19}$ J. Further, *Humphris and Cann* [2000] speculate that such deposition could be achieved through a brief period of high heat flow, for example, from the precipitation of Fe from high-temperature fluids flowing at a rate of 650 kg/s (with an associated power output of ca. 1,000 MW) over a period of 100–1,000 years. Because this rate of heat supply was significantly greater than can be provided from steady-state flux along a short section of slow-spreading ridge crest, the preferred explanation presented for these calculations was that hydrothermal circulation at TAG was driven by short periods of intense activity driven by episodic magmatic supply lasting tens of years, separated by thousands of years of inactivity [*Humphris and Cann*, 2000]. Such a timeline for TAG would be entirely consistent with the chronology of *Lalou et al*. [1998] and *You and Bickle* [1998].

3.2. The Rainbow Hydrothermal Field: An Even Larger High-Temperature Vent System

The Rainbow hydrothermal field, like TAG, is a fault-controlled hydrothermal system that is significantly larger than most typical volcanically-hosted vent sites. This system, which is associated with a non-transform discontinuity, occupies an area on the seafloor measuring ca. 300 m along axis and 100 m across axis. It hosts at least 10 discrete clusters of active black smoker (≤364°C) chimneys. The current output for the site, estimated from hydrothermal plume heat-flux studies, is 2.3 GW [*Thurnherr & Richards*, 2001; *Thurnherr et al*., 2002], i.e., significantly stronger than TAG [*Humphris and Cann*, 2000]. What is also important to note is that these fluxes at Rainbow represent a direct like-for-like comparison with the high-temperature fluxes considered for the EPR (above). At Rainbow, although fluxes are calculated from an integrated plume study, field observations reveal a distinct absence of diffuse flow at the Rainbow vent site [Desbruyeres et al., 2001] and this is confirmed by the negligible concentrations of dissolved Rn-222 measurable in the Rainbow non-buoyant plume – even directly above the vent site [*Cooper,*1999]. In more typical hydrothermal settings, high Rn-222 concentrations are observed in hydrothermal plumes and these fluxes are dominated by contributions from extremely enriched diffuse flow "ground waters" which are entrained into the turbulent buoyant plumes rising above individual black smokers. At Rainbow, such diffuse flow is essentially absent and all the measured heat-flux, according to He:heat, CH_4:heat, Mn:heat and Fe:heat ratios appears to be dominated by high-temprature "black-smoker" type fluid flow [*German et al*., submitted].

Furthermore, this level of output from the Rainbow hydrothermal field has been sustained throughout the past 8,000–12,000 years, based on a comparison of modern plume signals, 12-month sediment-trap fluxes, and long-term records preserved in sediment cores raised from beneath the dispersing non-buoyant plume [*Cave et al*., 2002]. If all the above field observations are correct, the apparent energy release implied over the past 10,000 years would be ca. $E_{Rainbow} = 7.3 \times 10^{20}$ J, i.e., more than an order of magnitude greater than that estimated for the TAG hydrothermal mound (although 6 such mounds, the others inactive, have also been reported for the TAG ridge segment). As we will see later, supply of this amount of energy is extremely problematic. Is it valid, therefore, to use the values listed above which have been derived from oceanographic and biogeochemical, rather than geophysical observations? First, we have confidence in the heat-fluxes calculated from hydrothermal plume studies because of the close convergence between values calculated from "instantaneous" plume observations made during on-station operations in 1997 [*Thurnherr and Richards*, 2001] and those calculated from long-term current meter moorings deployed between 1997 and 1998 [*Thurnherr et al*., 2002]. Second, we observe that geochemical fluxes from the present day plume (as collected in sediment traps) [*Cave*, 2002] agree directly with the average fluxes calculated for the surface mixed layer of underlying sediments [*Cave et al*., 2002]. Reassuringly, the sediment trap data were collected during the same time interval as the long-term oceanographic data used to calculate heat flux [*Thurnherr et al*., 2002] and so those two data sets are certainly directly related. Of course, the average flux into the sediments' surface mixed layer does not necessarily reflect a constant flux. Rather, during the time to accumulate this much sediment (7–10 cm, 3–4 ky) fluxes may have waxed and waned to values both higher and lower than those associated with the 1997–98 study. Even if such were the case, however, the time-averaged flux during this period would still have been indistinguishable from that of a constant flux identical to the modern day. On the assumption that chemical:heat fluxes at Rainbow have also remained unchanged throughout this period, the inference must be that the time-averaged heat flux during accumulation of the surface mixed layer (3–4ky) has also persisted at a value indistinguishable from the present. Finally, below the surface mixed layer, C-14 dating shows a steady accumulation of sediments at 2.5–3.5 cm/kyr beneath the Rainbow plume with no distinguishable change in hydrothermal plume fluxes (c.f. surface mixed layer, sediment traps) extending back to 8–12 kyr [*Cave et al*., 2002]. On this basis we surmise that hydrothermal activity at the Rainbow vent-site has been sustained at, or close to, modern-day values over a time scale of ca.10,000 years. How might such large amounts of energy be supplied through hydrothermal circulation within the MAR rift valley?

One first approach is to apply the same considerations that we used previously for the fast- and intermediate-spreading ridges. At a full spreading rate of 26 mm/yr, the total crustal accretion within the past 10,000 years is ~260 m. The mean to the depth of the axial magma lenses on the EPR and Juan de Fuca Ridge is ~2 km, although no steady-state magma lens has so far been imaged along any section of the northern MAR that is not hotspot influenced [*Detrick et al.*, 1990]. If we apply our previous methodologies and initially assume that hydrothermal circulation can only access heat in the upper crust to a depth of ~2 km, we calculate that the heat available from the upper ocean crust sums to $E'_{upper\ crust} = E'_{crystallization}$ (7.15 x 10^{17} J/km, crystallization of basaltic magma) + $E'_{cooling}$ (1.35 x 10^{18} J/km, cooling of solidified basalt from 1,250 to 350°C) = 2.06 x 10^{18} J/km for each 1 km section along axis. Interestingly, this calculation suggests that it would be possible to provide all the heat required to form the TAG mound (E_{TAG} = 2 x 10^{19} J) by integration of the entire available heat release associated with formation of the upper 2 km of the oceanic crust along 10 km of the MAR axis, over a period of 10,000 years or, perhaps more relevantly, over just 2–2.5 km along axis over the 40,000–50,000 year lifetime of the TAG mound [*Humphris and Cann*, 2000]. In the case of the Rainbow hydrothermal field, by contrast, the situation is much more problematic. To provide all the heat calculated to have been released through the Rainbow hydrothermal system would appear to require an integration of all the available heat emplaced by formation of the uppermost ocean crust along ca. 350 km of the slow-spreading MAR axis. This seems clearly unrealistic.

3.3. Alternative Heat Sources at Slow-Spreading Ridges: Serpentinisation and Deeper Cooling?

From their distinctive end-member vent fluid compositions, it is already well-established that the Rainbow hydrothermal fluids are fed not only by some magmatic heat source at depth but that they must also be interacting with serpentinising ultramafic rocks [*Charlou et al.,* 2002; *Douville et al*., 2002]. Our next calculation, therefore, considers whether such serpentinisation reactions might play a significant role in balancing the heat budget calculated for Rainbow. For a latent heat $H_{serpentinisation}$ = 2.5 x 10^5 J/kg for the serpentinisation of peridotites [*Fyfe and Lonsdale*, 1981; *Lowell and Rona*, 2002] the heat budget for Rainbow could be entirely balanced by the serpentinisation of $M_{serpentinisation}$ = 2.92 x 10^{15} kg of ultramafic rocks or a volume of $V_{serpentinisation}$ = 8.85 x 10^{11} m^3 for mantle density of $\rho_{peridotite}$ = 3,300 kg m^{-3}. The thickness of the serpentinised mantle layer along the MAR must be quite variable but is not well known. For illustration purposes, however, if one assumes that mantle serpentinisation could proceed down into the top ca. 2 km of the upper mantle, the required volume of serpentinisation calculated above would be equivalent to extraction of all the heat available from serpentinisation beneath a 260-m wide (10,000 yrs of crustal growth) mantle block that extends ~1,700 km along axis. Even at such an extreme (~0.1 km^3/yr serpentinisation rate), the predicted average heat release associated with such serpentinisation would be ca. $E'_{serpentinisation}$ = 4.12 x10^{17} J/km over this 10,000 year period, i.e., only ~20% of the heat flux that could be sustained, over the same length scale, by cooling of the upper 2 km of the overlying oceanic crust. Thus we argue that while serpentinisation could contribute to the heat budget, it would be unlikely to provide the main source of heat required to drive the Rainbow hydrothermal system. (NB: By contrast, much lower rates of serpentinisation would be required to sustain the low-temperature ultramafic-hosted Lost City hydrothermal field: 5x10^{-3} to 1x10^{-4} km^3/yr [Früh-Green et al., 2003].)

One possible explanation for Rainbow is that hydrothermal circulation at the ridge axis could penetrate deeper at slow-spreading ridges than at fast- and intermediate-spreading ridges. On the EPR and Juan de Fuca Ridge, it seems unlikely that fluid circulation could penetrate deeper than the upper 1.4–2.5 km of the oceanic crust because that is the depth to the axial magma lens. On slow-spreading ridges, by contrast, microearthquake studies have shown brittle failure to a depth of 5–8 km along the MAR [*Toomey et al*., 1988; *Kong et al*., 1992; *Wolfe et al*., 1995]. If we repeat our earlier calculations for 260 m of extension across the MAR but allow heat extraction by hydrothermal circulation to extend to the base of a 6-km thick crust, the amount of heat potentially available through crystallization and cooling of the whole crust layer is increased three-fold to $E_{whole\ crust}$ = 6.18 x 10^{18} J/km over a 10,000 year period. To that can be added the amount of heat available from serpentinisation of the upper 2 km of the underlying upper-mantle lithosphere over the same period (1.18×10^{18} J/km), yielding a potential energy budget over the time frame of the Rainbow hydrothermal field's activity that totals $E_{whole\ crust\ +\ serpentinisation} = 7.36 \times 10^{18}$ J/km. Of course, it remains debatable whether serpentinisation can indeed generate significant amounts of heat at Rainbow [*Allen & Seyfried*, 2004]. Nevertheless, because even the maximum contribution from such a process is small it does not affect our overall outcome, that at such a heat flow rate, ~100 km of the MAR axis would be required to match the total heat budget for the Rainbow site ($E_{Rainbow} = 7.3 \times 10^{20}$ J) over the past 10,000 years.

We conclude this section, therefore, with the observation that hydrothermal circulation at the Rainbow vent field could have continued uninterrupted throughout the past 10,000 years IF it were supplied by all the heat available from cooling the entire 6 km section of the ocean crust emplaced within this time (± that released from serpentinisation of the underlying lithospheric mantle to a depth of 2 km), along ~100 km of the MAR ridge axis; this is essentially the full length of both ridge segments immediately north and south of the non-transform discontinuity that hosts the Rainbow vent field. While this continues to appear to be a rather extreme solution, we do already know that hydrothermal fluids can circulate laterally over some tens of kilometres at depth, as revealed from recent vent-fluid compositions along the Endeavour segment of the Juan de Fuca Ridge [*M.D.Lilley*, pers. comm., 2003]. Furthermore, while the preferred modern heat flux value for the Rainbow vent field is high (2.3 GW), the error margins associated with the calculations are rather broad [0.5–3.1 GW: *Thurnherr et al*., 2002]. At the highest extreme, of course, the above calculations would require an increase of >33% in the length of ridge axis affected, to ~150 km. The lower bound for that same heat flux estimate, by contrast, would reduce our calculated values to <25% of the current requirements, indicating thorough cooling along no more than 24 km of axial ridge crest, i.e., to 12 km along axis, north and south of the vent site.

What remains problematic, however, is that—as for the EPR—all we have identified here is a plausible mechanism to reconcile total heat budgets at Rainbow and TAG, for crustal production and hydrothermal cooling. Once again, however, the detailed mechanisms by which such heat extraction could take place have not been explained. At Rainbow, in particular, vent-fluid compositions provide clear evidence for phase-separation at depth which, if circulation does indeed penetrate to 5–6 km, implies temperatures in excess of 500°C are likely at depth [*Douville et al*., 2002]. Fluid circulation models cannot readily explain how such sustained high temperatures could be maintained both at depth and in exiting high-temperature fluids over the 10 ky timescales we have inferred [see *Lowell & Germanovich*, this volume].

3.4. Is There Missing Heat Flux Associated With Megamullion Detachment Fault Formation?

While the focused heat flow required to sustain Rainbow as an uninterrupted hydrothermal system with a power of 2.3 GW over 10,000 years at one single site appears large compared to steady-state heat flow available along axis [*Sinha and Evans*, this volume], it is relatively small considering the spatial and temporal focusing of magma accretion along slow-spreading ridges. The formation of "megamullion" detachment fault structures [*Cann et al*., 1997; *Tucholke et al*., 1998] is an extreme example of such focusing. During megamullion formation, magmatic crustal thickness is greatly reduced or is absent (Fig. 1) implying that the magma supply that would otherwise be delivered to the megamullion location may be diverted to other locations along the ridge axis. Megamullion features can be continuous along axis over distances of ca. 20 km and in time over periods of up to 2 million years [*Tucholke et al*., 1998]. At steady state, a slow-spreading ridge such as the central northern MAR (full spreading rate of 26 mm/yr) should extend by 26 km/Myr. The potentially missing magmatic crust from the formation of a megamullion detachment fault that lasts 2 Myr and extends 20 km along axis, therefore, could have a maximum volume of $V_{megamullion} = 6$ km x 52 km x 20 km $= 6.24 \times 10^{12}$ m^3. The minimum amount of heat required to be preferentially extracted from along this section of ridge crest, therefore, can be calculated from the same two components discussed earlier; crystallization of the basaltic/gabbroic oceanic crust and cooling of that material from emplacement temperatures of ca. 1,250°C to a standard reference temperature of 350°C. Repeating those calculations described previously but with the much larger volume of $V_{megamullion} = 6.24 \times 10^{12}$ m^3, we calculate that the missing heat from the formation of such a megamullion is $E'_{megamullion} = 2.5 \times 10^{22}$ J/km. For comparison, this much heat could only be extracted from the oceanic crust by a vent site of the 2.3 GW power observed at Rainbow if that vent site continued uninterrupted for a period of ~350 kyr, which is much longer than has been reported from the sedimentary record [*Cave et al*., 2002]. This argues for the importance of considering non-steady state magma and heat supply at slow-spreading ridges.

4. SUMMARY AND CONCLUSIONS

In this opening chapter to the volume we have emphasised the importance of understanding the interplay between episodic volcanic/tectonic processes at mid-ocean ridges and their interactions with the overlying ocean through convective hydrothermal circulation.

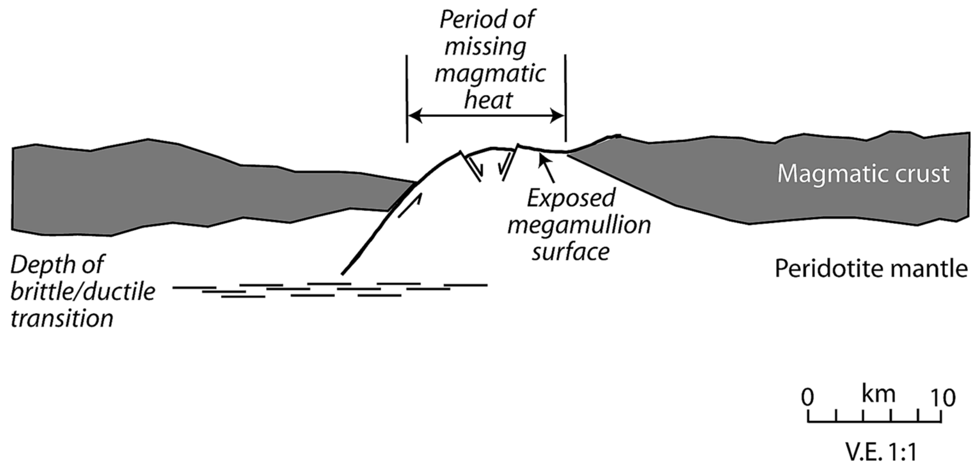

Figure 1. Schematic diagram of a megamullion detachment fault surface [modified from Tucholke et al., 1998]. During the formation of a megamullion detachment fault, the magmatic oceanic crust (shaded area) is significantly reduced or absent and, thus, much less heat release is expected from crystallization and cooling of this material. However, serpentinisation of the exposed peridotite mantle (white area) could contribute a small amount of heat flux.

For fast- and intermediate-spreading ridges we have proposed a simple conceptual model in which ridge extension in the upper crust is achieved via episodic diking events with a quantum extensional dimension. We argue that the heat available from emplacement of such a dike, which extends from the seafloor down to the roof of an axial magma chamber, could be matched by hydrothermal cooling in three stages: (a) instantaneous heat loss during a period in which “event plumes” are formed; (b) an “evolving” period of relatively fast discharge of heat through rapid cooling of the upper part of the dike in the form of diminishing but vigorous high-temperature hydrothermal flow; and (c) a decadal “quiescent” period, in which the residual heat available from cooling of the lower section of the dike is sufficient to sustain high-temperature flow for decades, albeit in the form of discrete and less-vigorous “black smoker” vent fields. While this approach matches thermal budgets and appears consistent with field data, to date, it cannot readily be reconciled with current fluid-circulation models. Further, our hypothesis may prove to be in need of significant refinement/re-evaluation if (for example) the EPR 9°N site suffers an eruption event comparable to 1991 within the next few years.

On the slow-spreading Mid-Atlantic Ridge, the apparent heat fluxes required to sustain the long-lived, tectonically-hosted TAG and Rainbow hydrothermal vent fields appear to be at least one order of magnitude greater than can readily be explained by steady-state cooling of the upper crust, alone. Hydrothermal cooling of the lower crust and heat release from serpentinisation could both contribute to the required heat budget. Furthermore, slow-spreading ridges exhibit much greater irregularity and episodic focussing of heat sources in space and time, as exemplified by the formation of megamullion detachment faults that can last ca.1–2 Ma. While focussing of heat sources, both spatially and temporally, along the MAR could be large enough to sustain high heat flow at the Rainbow and TAG sites, the mechanisms by which such heat could be extracted from the underlying lithosphere and delivered to such long-lived, highly-localised, vent-sites remains poorly understood.

The episodic nature of mid-ocean ridge and hydrothermal processes at ridges of different spreading rates is a theme that recurs throughout the chapters that follow. Much of the complexity arises from the inherent “quantum” nature of volcanic/tectonic events at mid-ocean ridges. What we have illustrated here is that tectono-magmatic and hydrothermal fluxes can become quite decoupled one from another – especially along slow-spreading ridges. This may explain how sections of even the slowest-spreading ridges can become magmatically robust and/or host vigorous hydrothermal systems, beyond what could be predicted based on long-term steady state models.

We urge that future investigations should strive to employ integrated geophysical and hydrothermal approaches to further our understanding of mid-ocean ridge systems. By recognising geophysical and hydrothermal approaches that cannot be reconciled at the present time, we identify those processes that most urgently need to be better understood.

Acknowledgments. This manuscript has benefited from many discussions with participants of the InterRidge Theoretical Institute (IRTI). Particular thanks are due to Karen Von Damm (UNH) and our reviewers, Bob Lowell (Georgia Tech) and Bill Seyfried (U.Minnesota)—not least for help in clarifying quite how much remains to be understood. We are grateful to Trish Gregg for creating a draft version of Plate 1. Significant funding for the IRTI was provided by the US RIDGE-2000 Program and the European Science Foundation. Additional support was provided by the InterRidge Office and the University of Pavia. CRG is funded through NERC's Core Strategic Research at SOC, JL is supported by NSF grant OCE-0129741 and a grant from the WHOI Deep Ocean Exploration Institute. This is Woods Hole Oceanographic Institution Contribution Number 11092.

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C. R. German, Southampton Oceanography Centre, European Way, Southampton SO14 3ZH, UK.

J. Lin, Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts 02543.