Geophysical Constraints on the Subseafloor Environment Near Mid-Ocean Ridges

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Studies of the subseafloor biosphere require an understanding of the physical characteristics of oceanic crust and in particular its hydrological state. Marine geophysics provides many tools to infer subsurface structure and processes remotely, but hydrological properties are difficult to infer from other physical parameters and are best constrained by controlled, in situ experiments. Because of the lack of boreholes at unsedimented ridge-crest sites, our knowledge of hydrological processes at mid-ocean ridges is derived indirectly by using geophysical images, geological studies, measurements in vent fields, and comparisons with off-axis sites as constraints for models of hydrothermal circulation. A synthesis of results from three well-studied ridge segments shows that there have been considerable advances in our understanding of the thermal structure of young oceanic crust and the linkages between hydrothermal, magmatic, and tectonic processes. However, there are still many unanswered questions related to processes in the heat uptake zone, the geometry of circulation, the formation and fate of brines, the effect of volcanic and tectonic events on hydrothermal systems, and the critical linkages between the many processes that affect crustal permeability and drive fluid flow. We argue that geophysical investigations will contribute most to studies of the subsurface biosphere near mid-oocean ridges if they first focus on shallow processes in the vicinity of hydrothermal upflow zones. The role of geophysical studies is likely to be enhanced by advances in underwater vehicle capabilities, the move towards establishing long-term seafloor observatories, and the prospects for new techniques to drill bare rock.

INTRODUCTION

Many investigations of oceanic spreading centers and their flanks are motivated by a desire to understand the geological processes that form and age the oceanic lithosphere. Geophysical studies are a major component of such work. A variety of measurements are available to determine remotely the structure of oceanic crust and to image the melt bodies and thermal anomalies associated with crustal accretion. Seismic records and geodetic observations constrain the deformation that occurs during faulting and volcanic events. Studies of hydrothermal circulation are critical for understanding heat transport in the oceanic crust and provide a framework for studies of rock alteration and chemical exchange with the ocean [e.g., *Lowell et al.*, 1995].

The growing awareness that oceanic hydrothermal systems may support a substantial subseafloor biosphere provides a new impetus for marine geophysical studies, but also presents new challenges. At the most general level, nearly all work that constrains the structure and evolution of oceanic crust provides useful constraints for studies of the subsurface biosphere. However, the contribution of geophysics can be defined more specifically in terms of spatial and temporal constraints on the parameters that determine the subsurface environment. These include the pathways available for colonization, the temperature, the pressure, the fluxes of chemical energy and nutrients, and the geometry of pore spaces in which the microbes live. By constraining these parameters, geophysical methods have the potential not only to delineate the nature of the subsurface environment, but also to constrain the effects of geological, chemical, and biological processes on the environment.

All of the geophysical parameters of environmental interest are dependent on or at least strongly linked to hydrological processes. The fluid pathways, fluxes and temperature are determined by the patterns of hydrothermal circulation. In principle, they are fully constrained by knowledge of permeability and thermal boundary conditions. Hydrothermal flow also determines the environment for the water-rock reactions that control fluid chemistry. Dispersive mixing is a consequence of the geometry of the permeable paths and networks and, together with conductive cooling, is partly responsible for the chemical disequilibria that provide the energy to support life. The fluid pressure in an open, highly-permeable system will be close to hydrostatic but as the fluid pathways seal with depth, the pressures will evolve towards lithostatic. While there is no simple relationship between permeability and porosity, these parameters are closely linked. The scale and interconnectivity of the porosity determines the permeability and, to a first order, both parameters tend to decrease with depth in the oceanic crust.

In this paper, we first provide a brief review of some fundamental hydrological concepts, followed by an overview of the global constraints on the physical and hydrological properties of oceanic crust. We then summarize the geophysical techniques that are available to constrain the characteristics of hydrothermal circulation at spreading centers. To illustrate these techniques and highlight some unresolved issues, we discuss three, well-studied type settings: the East Pacific Rise (EPR) at 9º-10ºN, the Endeavour Segment of the Juan de Fuca Ridge, and the TAG hydrothermal field on the Mid-Atlantic Ridge. We end with some speculation as to how geophysical methods will contribute to future studies of the subseafloor biosphere, and we offer our suggestions as to what kinds of methods may provide the most useful information.

HYDROLOGICAL CONCEPTS

Hydrogeologic studies of seafloor hydrothermal systems usually focus on one or more of these topics: (1) hydrogeologic properties (transmissive and storage), (2) driving forces, (3) fluxes and fluid pathways. The first category is readily addressed with measurements on the seafloor and in boreholes [e.g., *Fisher*, 1998; *Davis et al.*, 2000], although there are considerable difficulties related to the transient nature of flow processes, heterogeneity, scaling, anisotropy, and coupling between heat, fluid, chemical and biological systems. The second and third categories have also received considerable attention in the last several decades, and progress has been made, but the techniques and interpretive methods remain largely developmental [e.g., *Baker et al.*, 1996; *Davis and Becker*, 1998; *Elderfield et al.*, 1999]. In this section, we focus mainly on measurements related to hydrogeologic properties, and to a lesser extent, driving forces and fluxes.

Fluid fluxes in porous systems are generally related to the driving forces through Darcy's law [e.g., *Bear*, 1972], which was originally developed as a one-dimensional equation that can be written in terms of the specific discharge (the volume flux per unit area) or Darcy velocity, q:

$$q = \frac{Q}{A} = -K\frac{dh}{dL} \qquad (1)$$

where Q is the volume flux, A the cross-sectional area perpendicular to the flow, K a constant of proportionality known as the hydraulic conductivity and dh/dL the spatial gradient in the head, which is a measure of potential energy. Hydraulic conductivity is an empirically-derived coefficient that describes the ease with which a fluid moves through a rock. It is dependent on both rock and fluid properties and the state of fluid saturation. When considering seafloor systems, it makes sense to consider rock and fluid properties separately, since the fluid may be in liquid and/or vapor form, the properties of which are highly variable as a function of pressure, temperature and composition. Hydraulic conductivity is related to rock permeability, k by

$$K = k\frac{\rho_f g}{\mu} \qquad (2)$$

where μ is the fluid viscosity, ρ_f is the fluid density, and g the acceleration of gravity.

In mid-ocean ridge systems the flows are generally transient and multidimensional. The definition of the head gradient breaks down for a system in which phases separate, heat and solutes are transported, and fluid properties change [e.g., *Oberlander*, 1989]. The driving force that moves fluid through rocks under these circumstances results from a combination of pressure and density gradients and can be written using a three-dimensional form of Darcy's law.

$$\mathbf{q} = \frac{k}{\mu}(\nabla p - \rho_f g\hat{\mathbf{z}}) \qquad (3)$$

where ∇p is the pressure gradient and $\hat{\mathbf{z}}$ is a unit vector pointing downwards.

For multiphase systems each phase may move in a different direction, enhancing the potential for phase separation

[*Ingebritsen and Sanford*, 1998]. The permeability for each phase must be considered separately, and they may be related in a distinctly non-linear fashion. For example, the effective rock permeability within a liquid-vapor system can be orders of magnitude less than the permeability of the same system in which only liquid or vapor is present. This occurs because individual phases may not be well connected along permeable pathways; the presence of one phase may serve to block flow of the other. At very high flow rates and in open channels, flow becomes turbulent and Darcy's law may not apply [e.g., *Bear*, 1972].

Complexities also arise from the nature of permeability itself. Permeability is an empirical construct that exists only in the context of fluid flow. Proxy methods for estimating this property from other physical properties are notoriously unreliable (inaccurate and non-transferable) and geophysical observations, independent of controlled flow experiments or well-posed modeling studies, provide only weak constraints at best. In some sedimentary systems, flow occurs primarily through the voids between sedimentary grains and permeability measurements obtained from hand samples can be reasonably extrapolated to larger scales. However, fluid flow through ocean basement is dominated by fractures and other regions of relatively high permeability, making this property heterogeneous, anisotropic, and scale dependent [*Fisher*, 1998; *Becker*, 2003]. The permeability of fractured systems may vary by orders of magnitude over distances as small as a few centimeters [*Clauser*, 1992]. As a consequence, flow is highly channelized [*Tsang and Neretnieks*, 1998]. In the context of a single fracture, channelization means that most of the fluid will move along a small part of the fracture plane, away from asperities (Plate 1a). At a larger scale, the bulk of the fluid flux will be confined to the most permeable pathways, even if the bulk rock is highly permeable (Plate 1b). In addition, because permeability is a tensor quantity and the flow must follow the most permeable pathways, the fluid will often travel in directions that are oblique to the impelling force.

Beneath mid-ocean ridges the geological processes that control permeability are complex, dynamic, and poorly known (Plate 2). Magmatic processes lead to the formation of a "layered" crust with depth dependent initial properties and provide the heat source to drive vigorous hydrothermal circulation. Ridge-perpendicular tectonic extension and the thermal contraction that accompanies cooling provide mechanisms to generate new porosity and permeability. Chemical dissolution/precipitation and rock-hydration reactions can both create and destroy porosity locally, but should have the net effect of reducing permeability overall because the density of alteration minerals is relatively low. One can speculate that microbial blooms or biologically catalyzed chemical reactions may also play significant roles in permeability evolution, but the magnitudes and mechanisms of these influences are unknown at present.

Another complexity of flow through rocks is the tendency for heat and dissolved chemicals to disperse by hydrodynamic mixing. Dispersion in the direction of flow occurs because the fluids near the center of a crack move faster than those near the walls, and in faster flows it can be enhanced by turbulent mixing. At a larger scale, along-flow dispersion is enhanced by the presence of more and less direct pathways. Dispersion perpendicular to the flow results because the flow paths are tortuous; two markers starting a given distance from each other and moving at the same velocity will not remain the same distance apart. Dispersion is generally more important for chemical than heat transport within hydrothermal systems because chemical diffusivity is about three orders of magnitude smaller than thermal diffusivity, leading to greater differences between the chemical state of fluids and surrounding rocks than between their thermal states. Dispersion is generally highest in vigorous flows through heterogeneous and fractured media [*Gelhar and Axness*, 1983] but dispersion within the subseafloor may be significant even for sluggish net flows if a tide-driven oscillatory flow is superimposed [*Davis and Becker*, 1999]. Mathematically, dispersion has the same dimensions as diffusion and can be represented by a tensor whose components are a function of the fluid velocity. Quantifying the dispersion tensor requires a combination of well-constrained field measurements (generally tracer experiments) and a good statistical understanding of the distribution and characteristics of permeable paths [e.g., *Novakowski et al.*, 1995].

In order to describe how porous systems respond to perturbations, it is also necessary to consider the ability of a porous system to store or release fluid in response to changes in pressure. For a compressible medium the expressions are quite complex [*Wang*, 2000; *Wang*, 2003], but assuming incompressible rock grains, the specific storage may be defined as:

$$S_s = \rho g \left(\frac{\Delta V_f / V_f}{\Delta P} \right) = \rho g(\alpha + \phi\beta) \qquad (4)$$

where ΔV_f is the volume of fluid stored or released per volume of rock, V_r, ΔP is the change in pressure, α is the frame compressibility, ϕ is the porosity, and β is the fluid compressibility. Fluid compressibility changes by orders of magnitude over the range of temperatures encountered within seafloor hydrothermal systems, particularly when there is a vapor phase present, but rock compressibility is also highly variable and often poorly constrained. This term may be frequency dependent and, within highly fractured systems, is

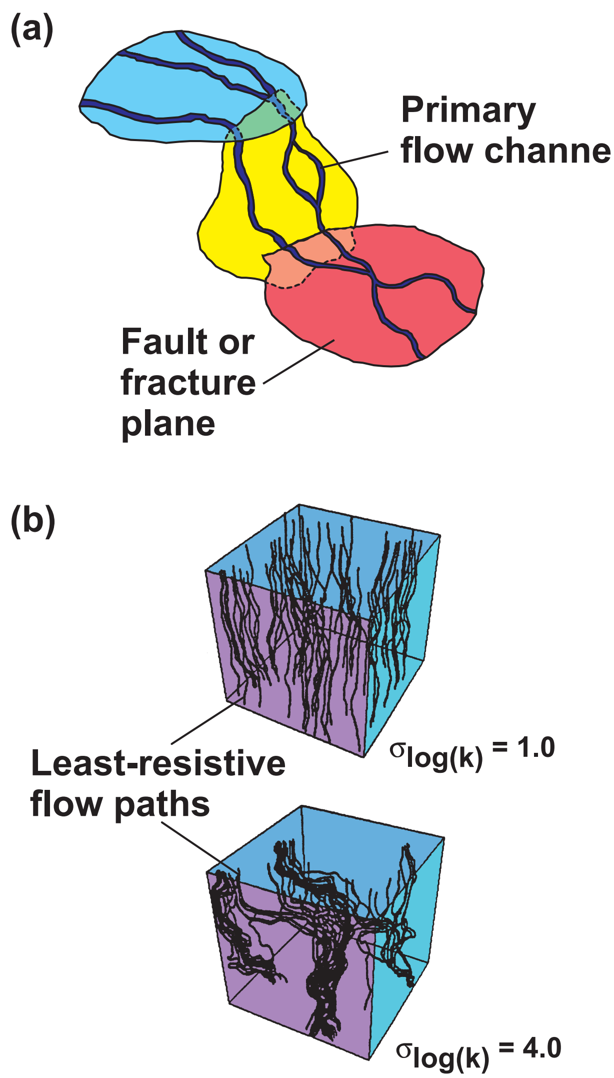

Plate 1. Examples of flow channeling at two scales. (a) At a small scale involving flow along a network of fractures, most flow occurs along a small fraction of each fracture surface. Figure modified from *Tsang et al.* [1991]. (b) Most-important solute pathways through heterogeneous rock systems having the same mean permeability, but different standard deviations in log-permeability, as shown [from *Moreno and Tsang*, 1994].

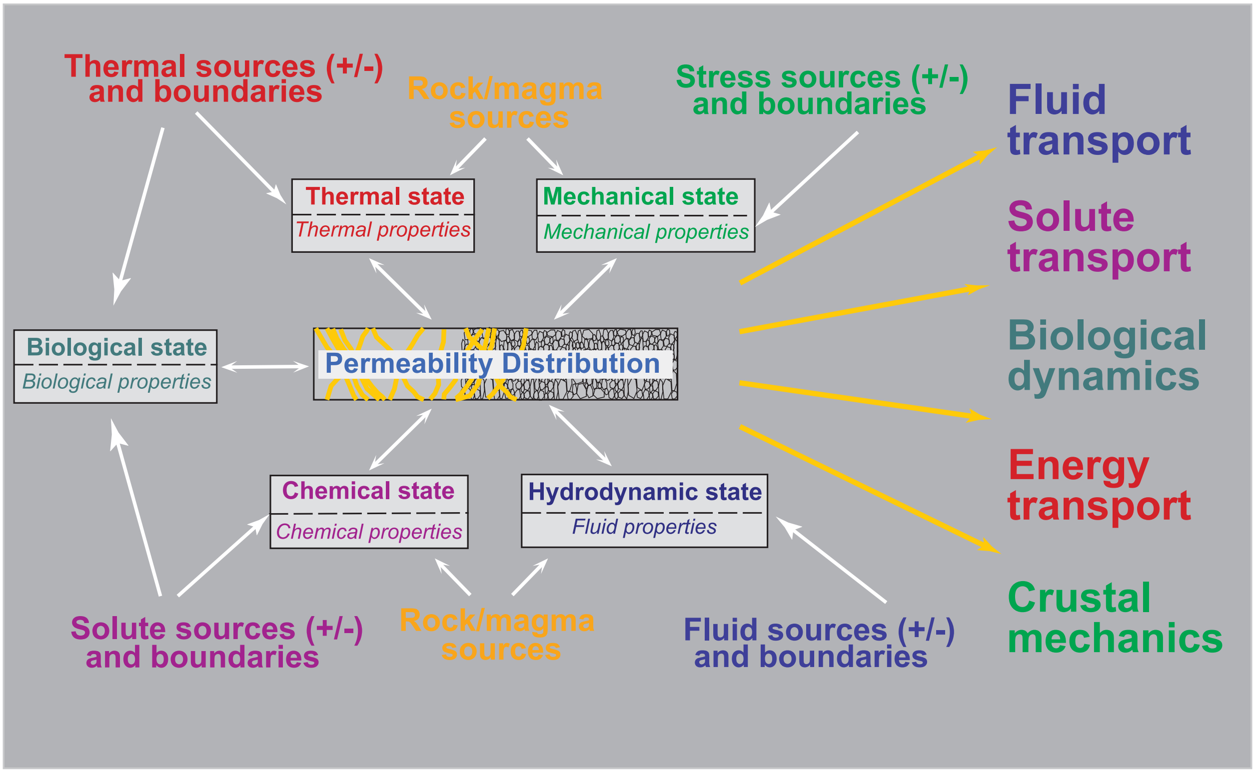

Plate 2. Conceptual illustration of relations between permeability, properties that can be measured (boxes surrounding permeability representation) and processes that are of greatest interest to scientists studying seafloor hydrothermal systems (to the right of the diagram). The properties and processes we measure cannot be related to those about which we care the most without understanding the nature (magnitude, distribution, dynamics) of permeability. Modified from *Fisher* [1998].

likely to be orders of magnitude greater than the compressibility of liquid water.

The ratio of hydraulic conductivity to specific storage is the hydraulic diffusivity, $\kappa = K/S_s$, and this is the property that determines the extent to which a perturbation in pressure propagates through a fluid-vapor-rock system [e.g., *Davis et al.*, 2000]. When the hydraulic diffusivity is high (high hydraulic conductivity and/or very rigid rock frame), a perturbation can travel a long distance quickly and with minimal attenuation. In contrast, when the diffusivity is low (low conductivity and/or very compliant rock frame), a large perturbation may not travel far from the source, will move more slowly, and its magnitude will be greatly reduced. Thus, hydraulic diffusivity is a critical parameter in determining the nature of coupling between transient processes.

GLOBAL CONSTRAINTS ON THE PHYSICAL PROPERTIES OF OCEANIC CRUST

Although permeability is difficult to infer from other physical parameters, permeability measurements in the oceanic crust are sufficiently sparse that they are best interpreted within the context of a more general understanding of the properties of oceanic crust. Geological layering and the effects of increasing overburden pressure with depth lead to large and systematic vertical variations in nearly all physical properties of oceanic crust (Figure 1). Away from spreading centers, the horizontal variations tend to have smaller amplitudes or occur at scales that are either too small to be resolved by remote geophysical techniques or too large to be detected by individual experiments. For these reasons, geophysicists frequently interpret their observations in terms of average one-dimensional structure, although it is important to remember that this is often a poor approximation near the ridge axis.

The strongest and most widespread geophysical constraints on subsurface structure come from seismological studies. The velocities of seismic waves are a function of both elastic moduli and bulk density and are affected by composition, temperature, porosity and the distribution of cracks. Experiments using artificial seismic sources are configured to look at either the reflection of near vertical waves from discontinuities or refracted waves that propagate horizontally at their depth of deepest penetration. Data from early seismic refraction experiments in the oceans were modeled by a stack of constant velocity layers that were then interpreted in terms of geological units. A legacy of these experiments is the widely used seismic nomenclature, with layers 2 (later subdivided into 2A, 2B and 2C) and 3 corresponding to the upper and lower crust, respectively [*Raitt*, 1963]. More modern refraction and reflection experiments show that systematic sharp changes in velocity are generally limited to the sediment/basement interface, the layer 2A/2B boundary and the Moho. The sediment/basement interface and Moho are petrological boundaries while the layer 2A/2B boundary marks a sharp drop in porosity at the base of highly porous lavas. Global data sets show that the compressional wave velocity and the ratio of shear wave to compressional wave velocities increase markedly in layer 2A over a few million years. These changes are attributed to the sealing of cracks and a decrease in porosity caused by hydrothermal alteration [*Houtz and Ewing*, 1976; *Jacobson*, 1992; *Carlson*, 1998].

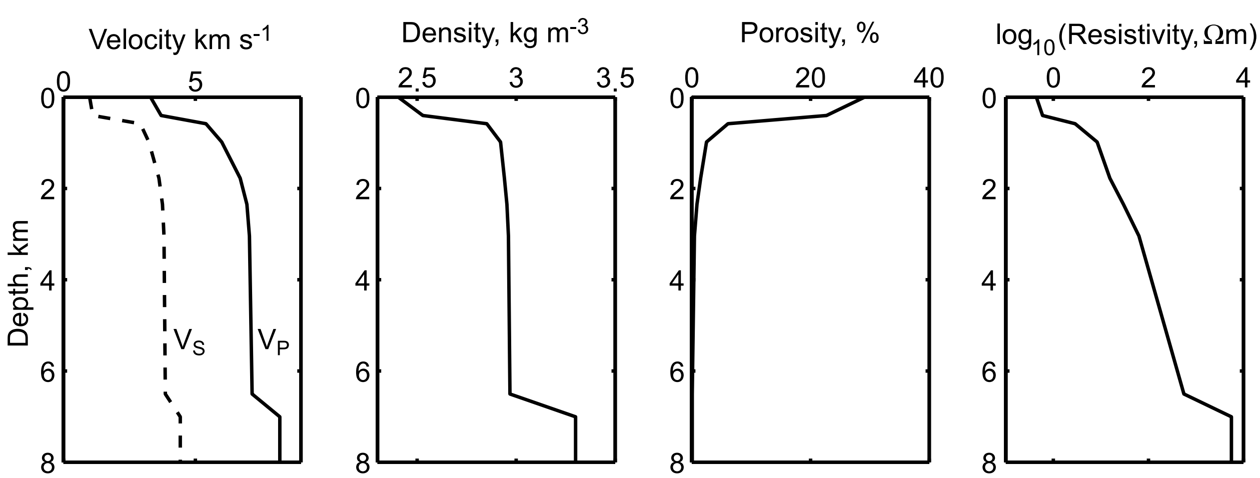

Figure 1. Illustrative vertical profiles of physical properties in young oceanic crust. The seismic velocities are a simplification of the off-axis model of *Vera et al.* [1990] for the East Pacific Rise near 9º30′N. The porosities and densities are based on a time-averaged Wyllie relationship above 2 km depth [*Evans*, 1994] and on a linear relationship between density and P wave slowness at greater depths [*Carson and Herrick*, 1990]. The resistivity is obtained from the porosity using Archie's law [*Evans*, 1994].

Several other geophysical techniques provide complementary constraints. Horizontal variations in the acceleration of gravity are sensitive to lateral changes in bulk density, which can arise from horizontal variations in lithology, temperature, porosity and the depth of interfaces. Gravitational studies can be used to infer variations in crustal thickness but the interpretation of gravitational anomalies from deep features is non-unique. Near-bottom measurements obtained over bathymetric features do provide strong constraints on the density and hence the porosity of the uppermost crust. The method shows that the average porosity of young lava flows is 20-35% but decreases quickly with age [e.g., *Luyendyk*, 1984; *Holmes and Johnson*, 1993]. Deeper penetration requires measurements over fault scarps or in fracture zones [e.g., *Johnson et al.*, 2000b]. Here, the effects of faulting may locally modify the porosity, but the technique confirms that the layer 2A/2B boundary is marked by a sharp drop in porosity.

Measurements of the earth's magnetic field are sensitive to lateral variations in the orientation and intensity of the magnetization of the oceanic crust. The primary magnetic mineral in the oceanic crust is titanomagnetite, and fine grains of this mineral in extrusive rocks record the earth's magnetic field near the time of crustal formation. Magnetic surveys are most widely used to map the isochrons associated with reversals of the earth's magnetic field, but measurements of intensity of magnetization are also sensitive to the accumulated effects of hydrothermal alteration. Titanomagnetite is susceptible to alteration even at low temperatures and the intensity of magnetization decreases with crustal age. This process is accelerated in regions that have experienced higher temperature upflow [e.g., *Tivey and Johnson*, 1989].

The electrical resistivity of oceanic crust is sensitive to both composition and the presence of conductive fluids in interconnected pores. Measurements of shallow resistivity can be made by placing electrodes into the seafloor but most techniques look at the interaction of natural or induced electromagnetic fields with the oceanic lithosphere [*Chave et al.*, 1991]. In the upper crust, vertical variations in resistivity can be interpreted using Archie's law, an empirical rule that relates the measured resistivity to that of a conductive pore fluid (seawater) and porosity [e.g., *Evans*, 1996]. The results also confirm that layer 2A is highly porous and that the porosity of the upper crust decreases with both depth and age.

The properties inferred from remote geophysical techniques are complemented by in situ measurements in a relatively small number of Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) boreholes. Logging tools can be used to measure fine scale vertical variations in the physical properties of the rock surrounding a borehole, and provide strong insights into how fine-scale geological structure is mapped into lower resolution geophysical images [e.g., *Salisbury et al.*, 1996]. Boreholes are critical for hydrological studies because they provide a means to measure subsurface permeability in situ and to determine the nature of the features that contribute to this property (Plate 3). The two most direct methods involve use of drillstring packers and temperature (fluid flow) logs [*Fisher*, 1998]. A packer is a device that hydraulically isolates part of a borehole so that fluid pressure can be modified and changes in formation conditions can be monitored. The tool and measurement techniques used in the seafloor are very similar to those used in aquifers and petroleum reservoirs, but controlled experiments in seafloor basement rocks have thus far been completed only in a "single hole" configuration, precluding accurate estimates of storage properties. Borehole temperature logs have been used to estimate fluid flow rates to and from the surrounding formation, and in combination with an estimate of driving forces, these data allow calculation of effective permeabilities [*Becker et al.*, 1983; *Becker*, 2003]. The permeability values estimated with packers and temperature data are bulk values for some assumed borehole interval, so they do not capture the true heterogeneity of the formation. In addition, because these tests are generally of relatively short duration and involve a single hole, they are indicative only of properties in the immediate vicinity of the borehole.

Despite these limitations, the borehole estimates have yielded a remarkably consistent view of formation hydraulic properties (Figure 1). The data suggest that the highest bulk permeabilities, $>10^{-13}$ m^2, can be found within seismic layer 2A and within thin, stratigraphically-defined intervals. When the global data set is considered as a whole, borehole-estimated permeabilities within the shallowest basaltic crust also decrease significantly over the first few million years in concert with the change in layer 2A properties inferred from other geophysical observations [*Becker and Fisher*, 2000]. The permeabilities measured in the upper crust below layer 2A are typically several orders of magnitude higher than found in layer 2A. While there are no in situ permeability measurements to date of hydrogeologic properties within the lower crust, it seems likely that both porosity and permeability below the upper few kilometers of basement rocks are low. An important exception to this general trend may occur near some seafloor spreading centers, as discussed later.

Further constraints on off-axis permeability come from measurements of conductive heat flow at the seafloor. Global heat flow data deviate from plate-cooling models until the seafloor is on average 65 Ma because vigorous hydrothermal circulation and advective heat loss continue within the crust [*Sclater et al.*, 1980; *Stein and Stein*, 1994].

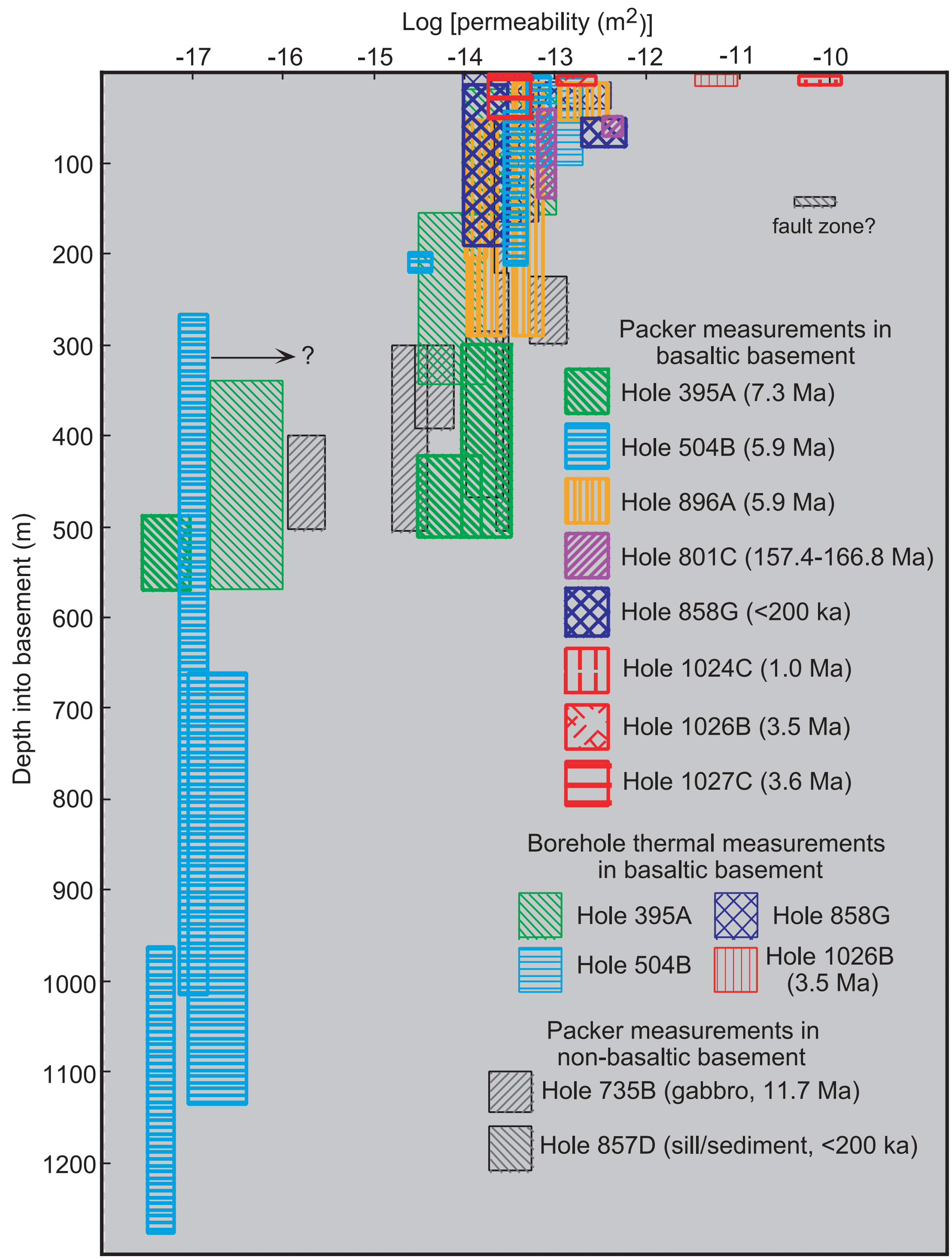

Plate 3. Summary of bulk permeabilities estimated from single-hole packer experiments and temperature logs in DSDP and ODP boreholes. Data from *Becker and Fisher* [2000] and *Fisher* [1998]. Depths are referenced to the sediment-basement interface in the case of all measurements in basaltic basement, but depths in Hole 735B (gabbro) are relative to the seafloor, and depths in Hole 857D (sediment-sill complex) are relative to the top of the first sill. Depth intervals indicate the interval of the entire open hole, except for the thin interval in Hole 857D, constrained using on geophysical logs and drilling conditions. The range of permeabilities shown for each measurement is a crude estimate of uncertainty in bulk properties, assuming that properties are evenly distributed across the indicated depth interval.

Local variations in seafloor heat flow and correlations with basement relief demonstrate that hydrothermal circulation continues in some places, confined largely to basement, at least until 105 Ma [e.g., *Noel and Hounslow*, 1988]. These results, coupled with a variety of other physical and chemical observations and modeling studies, indicate that the range of effective permeabilities overlaps the results from packer experiments and temperature logs, but extends several orders of magnitude higher [*Fisher and Becker*, 2000], a result consistent with estimates made from formation tidal response and results of numerical models [e.g., *Davis and Becker*, 2003]. The most likely explanation is that the bulk values determined in single-hole experiments are closer to background properties, whereas the larger-scale estimates are dominated by widely-spaced flow paths [*Fisher and Becker*, 2000; *Becker*, 2003].

CONSTRAINTS ON HYDROTHERMAL CIRCULATION AT SPREADING CENTERS

The emphasis of hydrological studies at spreading centers differs significantly from off-axis sites. With the exception of a few sedimented ridges, there are no boreholes suitable for hydrological measurements near spreading centers. The permeability structure beneath bare rock sites must be estimated indirectly from proxies and modeling studies or by extrapolating borehole values from sedimented sites. It is also difficult to measure conductive heat flow on bare rock [*Johnson and Hutnak*, 1997] and so detailed maps of seafloor heat flow are not available to constrain the patterns of upflow and downflow. Instead, seafloor studies at ridges focus almost entirely on sites of discharge.

Since the discovery of venting nearly a quarter of a century ago, there have been extensive efforts to locate and characterize vent sites. Along faster spreading ridges, nearly all known vents are located near the neovolcanic zone, while on slower spreading ridges they are more widely distributed. The existing observations may not be completely representative since search strategies are strongly biased towards the spreading axis. In particular, the extent of hydrothermal venting on young ridge flanks is largely unknown.

Most submersible and ROV studies in vent fields focus on characterizing the geology of sulfide structures and their setting, collecting water samples for chemical analysis, and studies of biology. Hydrological measurements are relatively sparse. Discharge temperatures are generally recorded when collecting fluid samples and there are various methods to measure spot heat and mass fluxes from both black smokers and areas of diffuse flow. There are sufficient time series observations to demonstrate that discharge temperatures and flow rates are often correlated with tides [e.g., *Schultz et al.*, 1996], but so far there have been only a few attempts to interpret such data in terms of subseafloor hydrological properties [e.g., *Dickson et al.*, 1995; *Jupp*, 2000]. Recent studies have also demonstrated correlations between subsurface fluid pressure and discharge temperature with tectonic activity [*Sohn et al.*, 1998; *Johnson et al.*, 2000a; *Davis et al.*, 2001]. The total heat flux from a vent field or a ridge segment can be estimated either by extrapolating spot measurements near the seafloor or by using water column surveys to constrain the heat budget in the thermal plumes that form above high-temperature vent sites [*Baker et al.*, 1995; *McDuff*, 1995]. One particular challenge in estimating fluid and heat output from plume studies is distinguishing between chronic and event plumes, since the former may be interpreted to result from relatively continuous heat-extraction processes (penetration of a cracking front, convective mixing), whereas the latter is thought to result from relatively short-lived processes following volcanic eruptions [e.g., *Baker et al.*, 1987; *Palmer and Ernst*, 1998].

Small-scale near-bottom surveys provide a means to map the magnetic and gravity lows produced by hydrothermal alteration in near-surface regions that have experienced upflow [*Tivey and Johnson*, 1989]. Although electromagnetic data are challenging to interpret in terms of three-dimensional electrical resistivity structure, such techniques may provide a means to image hydrothermal upflow to greater depth [*Evans et al.*, 1998; 2002]. High-temperature (~300°C) hydrothermal fluids are an order of magnitude more conductive than cold seawater and the conductivity of upflow zones may be further enhanced by the precipitation of sulfide minerals.

When a mid-crustal magma body is present, the magma chamber roof can be imaged with seismic reflection techniques. Tomographic methods can be applied to refraction and earthquake data, to image seismic velocity and attenuation anomalies within and around the magma body. Electromagnetic techniques are also sensitive to the presence of interconnected melt. The interpretation of seismic and resistivity anomalies in terms of melt content and temperature perturbations is not trivial and is dependent on the assumed topology of any partial melt, but the constraints can be quite strong [e.g., *McGregger et al.*, 1998; *Dunn et al.*, 2000].

Studies of mid-ocean ridge earthquakes constrain tectonic processes and the location of active faults that are likely to provide some of the most permeable pathways for deep circulation. Because the transition from ductile to brittle deformation is temperature controlled, the maximum depth of earthquakes also constrains the extent of hydrothermal cooling [e.g., *Toomey et al.*, 1985]. Seismic anisotropy can be inferred either by looking at systematic variations in travel times or refracted P waves as a function of propagation direction or by looking at the polarization characteristics of earthquake shear waves [e.g., *Barclay and Toomey*, 2003]. Near mid-ocean ridges, upper crustal anisotropy results because cracks are preferentially aligned parallel to the spreading axis. Thus, it is likely that the permeability structure is also anisotropic, although there is no simple means to quantitatively interpret seismic anisotropy measurements in terms of permeability.

The cumulative constraints from geophysics and complementary geological and geochemical studies are generally interpreted in terms of their implications for models of circulation. These models may be conceptual but are often based on analytical or numerical solutions of the equations for Darcy flow and the transport of heat and solutes. Numerical models fall into two general categories [*Lowell et al.*, 1995]. Single pass or pipe flow models are the simplest since they assume that the flow is confined to predetermined pathways. They are a useful tool for understanding the general behavior of a system or assessing the importance of one particular process, without considering the details of the temperature and velocity distribution. Porous medium convection models are generated by solving the appropriate conservation equations and can in principle provide a very detailed description of hydrothermal processes. However, in practice the solution of complex coupled equations presents significant numerical challenges and many processes that affect the permeability are still too poorly understood to be meaningfully described by mathematical equations. To date most models of mid-ocean ridge systems have considered only single phase flow without chemical transport in rectangular regions with simple boundary conditions and permeability distributions.

FIELD EXAMPLES

In this section, we highlight geophysical and related work at three sites that have come to be de facto type examples of hydrothermal circulation at seafloor spreading centers: the East Pacific Rise at 9-10°N, the Endeavour Segment of the Juan de Fuca Ridge, and the TAG hydrothermal field on the Mid-Atlantic Ridge. These sites are often considered to have features typical of fast-, medium- and slow-rate spreading, but it is perhaps more useful to consider them in terms of the relative balance between magmatic and tectonic activity. By definition, all spreading centers experience a continuum of magmatic and tectonic processes. But the fundamental difference between hydrothermal systems across a range of spreading rates may be the probability of experiencing different kinds of spreading events [e.g., *Wilcock and Delaney*, 1997]. Tectonic events are disproportionately likely at slower-spreading segments, and magmatic events are more common on faster-spreading segments.

East Pacific Rise near 9°50′N

The East Pacific Rise between 9° and 10°N (Figure 2) is a fast spreading ridge with a full spreading rate of 106-110 mm/yr and it is arguably the most extensively studied ridge segment in the world. Numerous seismic studies provide an unparalleled view of the axial magmatic structure; near-bottom surveys constrain the relationship between hydro-thermal outflow and seafloor morphology; and there is an ongoing time series of hydrothermal observations near 9°50′N following an eruption in 1991.

Like most fast spreading ridges, the EPR spreading axis is underlain by an upper crustal magma lens which on this segment is located about 1.5 km below the seafloor [*Detrick et al.*, 1987] (Plate 4a). The magma lens is essentially continuous along axis, about 1 km wide and only a few tens of meters thick [*Kent et al.*, 1993]. It is overlain by a lower velocity zone that is a few hundred meters thick and is interpreted as a hot boundary layer separating hydrothermal fluids from the magmatic heat source [*Vera et al.*, 1990; *Toomey et al.*, 1994]. Beneath the magma lens, a region of low velocities and high attenuation extends into the mantle [*Toomey et al.*, 1990; *Vera et al.*, 1990; *Wilcock et al.*, 1995; *Dunn et al.*, 2000; *Crawford and Webb*, 2002] (Plate 4b). Within the lower crust, a tomographic study [*Dunn et al.*, 2000] suggests that the low velocity zone is 5-7 km wide, steep-sided and contains relatively small melt fractions. The width and melt content increases significantly beneath the Moho [*Dunn et al.*, 2000; *Crawford and Webb*, 2002].

The axial morphology (Figure 2) is characterized by a broad bathymetric high that is cut by a small axial summit trough [*Fornari et al.*, 1998a; *Perfit and Chadwick*, 1998]. The rise axis is broken into 10-20 km long segments that are bounded by small offsets or subtle bends in the rise axis [*Haymon et al.*, 1991]; these segments are termed fourth order in the hierarchy of ridge segmentation [*Macdonald*, 1998]. Seismic studies show that fourth-order segmentation extends into the upper crustal magma chamber [*Toomey et al.*, 1990; *Kent et al.*, 1993]. The relative ages of lavas, the density of fissures and characteristics of axial summit trough vary systematically between segments [*Haymon et al.*, 1991]. The segment boundaries appear to define the limits of dike propagation from volcanic centers that erupt independently.

The observations of hydrothermal venting suggest a fairly simple relationship with magmatism. All known vent sites are located above the magma lens and the intensity of venting tends to decrease with increasing age of the youngest lava flows [*Haymon et al.*, 1991]. Observations at 9°50′N following the 1991 eruption [*Haymon et al.*, 1993] and elsewhere [*Delaney et al.*, 1998] show that eruptions lead to a short-term burst in hydrothermal outflow driven by the

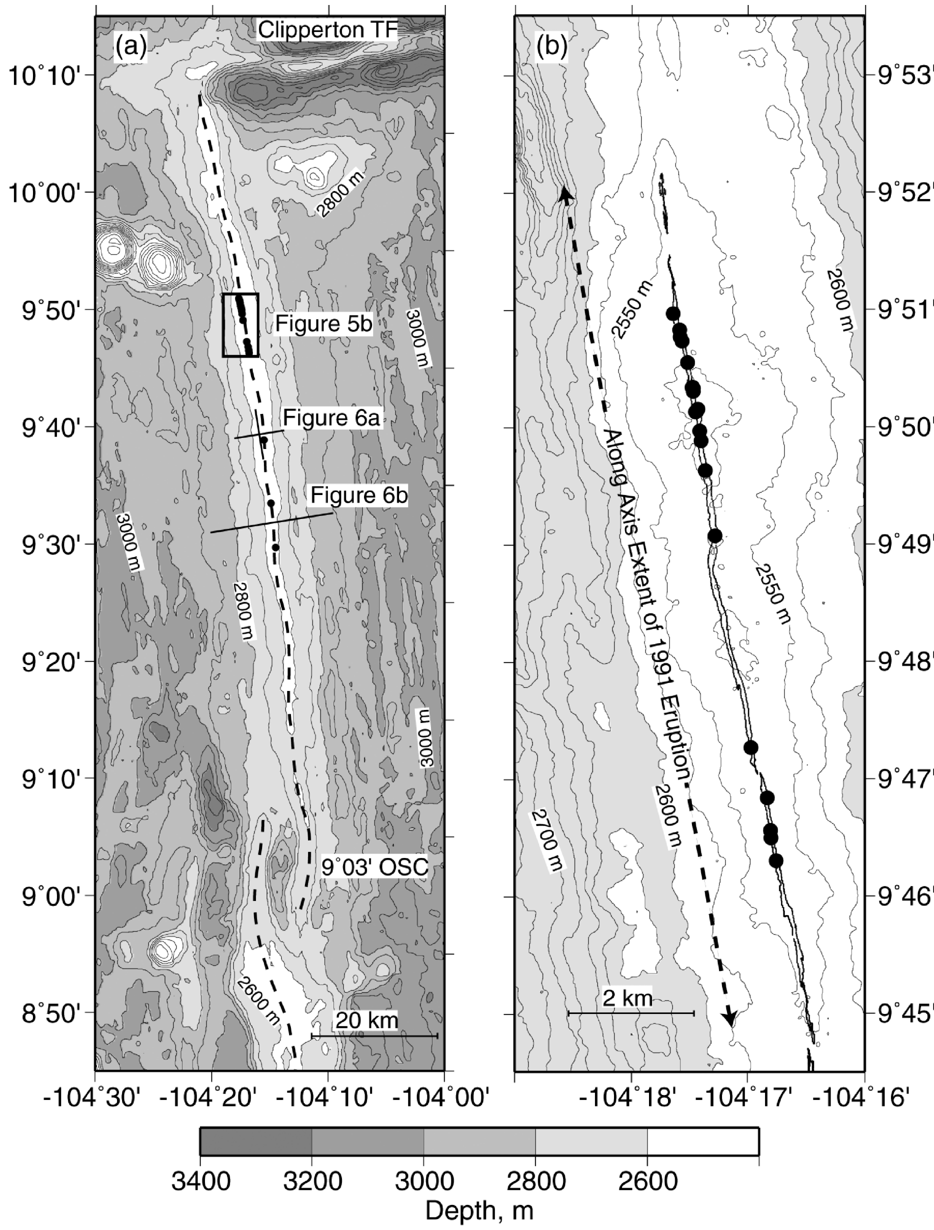

Figure 2. (a) Bathymetric map of the East Pacific Rise, 9-10°N, contoured at 100 m, showing the location of the ridge axis (dashed lines), high-temperature vents (filled circles), locations of seismic data shown in Plate 4 (solid lines) and the area covered by Figure 2b (bold box). The Clipperton transform fault (TF) and the 9°03′N overlapping spreading center (OSC) are labeled. (b) Detailed map of the 1991 eruption area contoured at 25 m showing the location of high temperature vents (filled circles) and the axial summit trough (solid line) [*Haymon et al.*, 1991; *Fornari et al.*, 1998a].

emplacement of magma in the upper crust and by the fractures that are produced by the emplacement of the dike. The 1991 eruptive event also led to a longer term increase in the intensity of venting and changes in the fluid chemistry that persist to the present (*Von Damm and Lilley*, this volume). Since the heat content of a dike or shallow intrusion will be exhausted on much shorter timescales [e.g., *Lowell and Xu*, 2000], decadal perturbations must reflect changes in the configuration of the heat uptake zone although the characteristics of this region are poorly understood.

Simple thermal calculations [e.g., *Wilcock and Delaney*, 1997], suggest that the impermeable conductive boundary layer separating hydrothermal fluids from magma or very hot rock must remain no more than ~10 m thick below

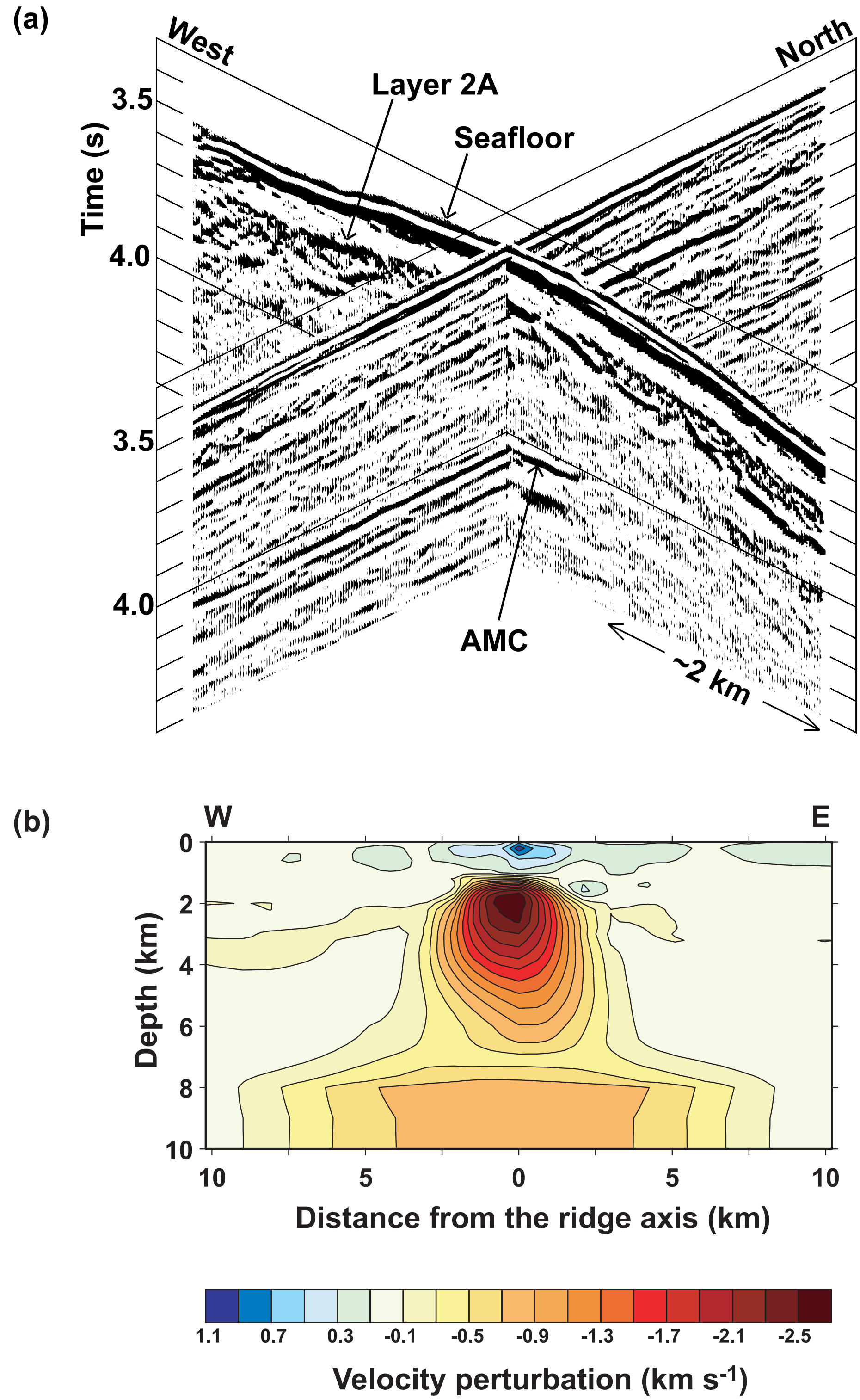

Plate 4. (a) Intersecting seismic reflection profiles oriented along and across the East Pacific Rise near 9º40′N showing reflections from the seafloor, the base of layer 2A, and the axial magma chamber (reproduced from *Phipps Morgan et al.* [1994]). (b) Velocity structure for a two-dimensional tomographic inversion on the East Pacific Rise at 9º30′N. The contour plot shows the velocity perturbations from a one-dimensional reference velocity model on the right (from *Dunn et al.* [2000]).

vigorous fields. *Lister* [1974; 1983] argued that this was infeasible for a static boundary layer and developed a model in which cracking induced by thermal contraction allows the hydrothermal system to mine heat along a downward migrating cracking front. *Lowell a nd Germanovich* [1994] present pipe model calculations that support the cracking front model as one alternative but they also argue that high heat fluxes can be maintained through a static boundary layer above a magma chamber by continuous magma replenishment and by the latent heat from crystals that settle to the base of the magma chamber. Support for the cracking front model at 9°50′N comes from observations of a swarm of small earthquakes in 1995 that was followed 4 days later by an increase in venting temperatures [*Fornari et al.*, 1998b; *Sohn et al.*, 1998].

Lister [1974; 1983] envisioned a model in which the cracking front only migrated downwards and penetrated deep into the crust, but this model cannot be correct for the axis of the East Pacific Rise since the axial magma lens is a steady state feature. It seems most likely that the fracturing during diking events thins the conductive boundary layer [*Lowell and Germanovich*, 1994]. Between magmatic events periodic cracking events due to thermal contraction or tectonic extension will also thin the boundary layer, but these effects are counteracted by mineral clogging and magma solidification on the roof of the magma lens. The shape and evolution of the boundary layer is likely be quite complex but over time its effective thickness and thus the hydrothermal heat fluxes will show a net decline until a fresh intrusion and/or a new eruption resets the clock.

Another topic of uncertainty at the East Pacific Rise is the geometry of hydrothermal recharge. Seafloor observations show that axial fissures are nearly all aligned parallel to the ridge axis. The density of fissuring increases with the relative age of lavas but the widest fissures are found on the youngest and most hydrothermally active ridge segments [*Wright et al.*, 1995a]. These extend at least to the base of layer 2A and appear to be eruptive features that tap melt during an eruption and then focus the high-temperature hydrothermal discharge that follows [*Wright et al.*, 1995b]. Seismic refraction data suggest that anisotropic crack distributions extend to about 2 km depth [*Dunn and Toomey*, 2001]. *P* wave anisotropy of 4% at 0.5-1 km and 2% at 1-2 km is consistent with vertical cracks aligned along the rise axis. Such observations lead many researchers to conclude that the circulation cells that cool the axial magma lens are aligned along axis [*Haymon*, 1996]. While these arguments are very plausible, they remain to be tested. The density of anisotropic cracks required to explain the seismic observations accounts for less than 10% of the total porosity [*Dunn and Toomey*, 2001]; in order for this porosity to dominate rock permeability, most of the fluid flow would have to be highly focused within a very small volume of rock. Without information on the connectivity of ridge parallel cracks or the geometry of the remaining porosity, the level of permeability anisotropy is poorly constrained.

Lowell and Yao [2002] argue that anhydrite precipitation which occurs when seawater heats up to ~150°C will clog a recharge zone unless it covers a large area extending well off axis. Seismic observations suggest that the lower crust cools substantially within a few kilometers of the ridge axis [*Dunn et al.*, 2000; *Crawford and Webb*, 2002] and this is only feasible if hydrothermal circulation penetrates the lower crust to the sides of the axial magma chamber [*Chen*, 2001; *Cherkaoui et al.*, 2003]. There is isotopic and chemical evidence that fluids circulate in the lower crust at ≥700ºC [e.g., *Gregory and Taylor*, 1981; *McCollom and Shock*, 1998; *Manning et al.*, 2000] although many would argue that the volume of fluid involved is too limited to play a significant role in heat transport. At such high temperatures, it is not clear if circulation occurs in a network of brittle fractures at hydrostatic pressures. It is also unclear whether the fluids that cool in the lower crust feed into the axial systems. However, unless there is extensive and as yet undiscovered venting just off axis, deep circulation must couple thermally into the axial cells and this requires a component of across axis flow.

Endeavour Ridge

The Endeavour (Figure 3) lies near the northern end of the Juan de Fuca Ridge and has an intermediate spreading rate of 61 mm/yr. The tectonics of the region are complex because the plate boundaries are reorganizing as the subduction of the Explorer plate to the north slows [e.g., *Rohr and Furlong*, 1995]. The Endeavour is offset at either end by overlapping spreading centers and a large portion of the segment appears to be a failing rift [*Karsten et al.*, 1986]. However, the morphology of the central third of the Endeavour is similar to other intermediate-spreading ridge segments and is a major focus of ongoing hydrothermal studies on the Juan de Fuca Ridge.

The Endeavour axis is characterized by a 0.5-km-wide, 100-m-deep axial valley that is highly fissured and devoid of recent lava flows [*Delaney et al.*, 1992]. Hydrothermal discharge is concentrated in at least five high-temperature vent fields (Figure 3b) with large sulfide structures that are spaced 2-3 km apart and located either at the foot of the west wall or near the center of the axial valley. The Endeavour segment is seismically active and microearthquakes extend to ~3.0 km depth beneath the axial valley [*Wilcock et al.*, 2002]. Early seismic imaging experiments [*Rohr et al.*, 1988; *White and Clowes*, 1990] were inconclusive as to the presence of an axial magma chamber. *Wilcock and Delaney*

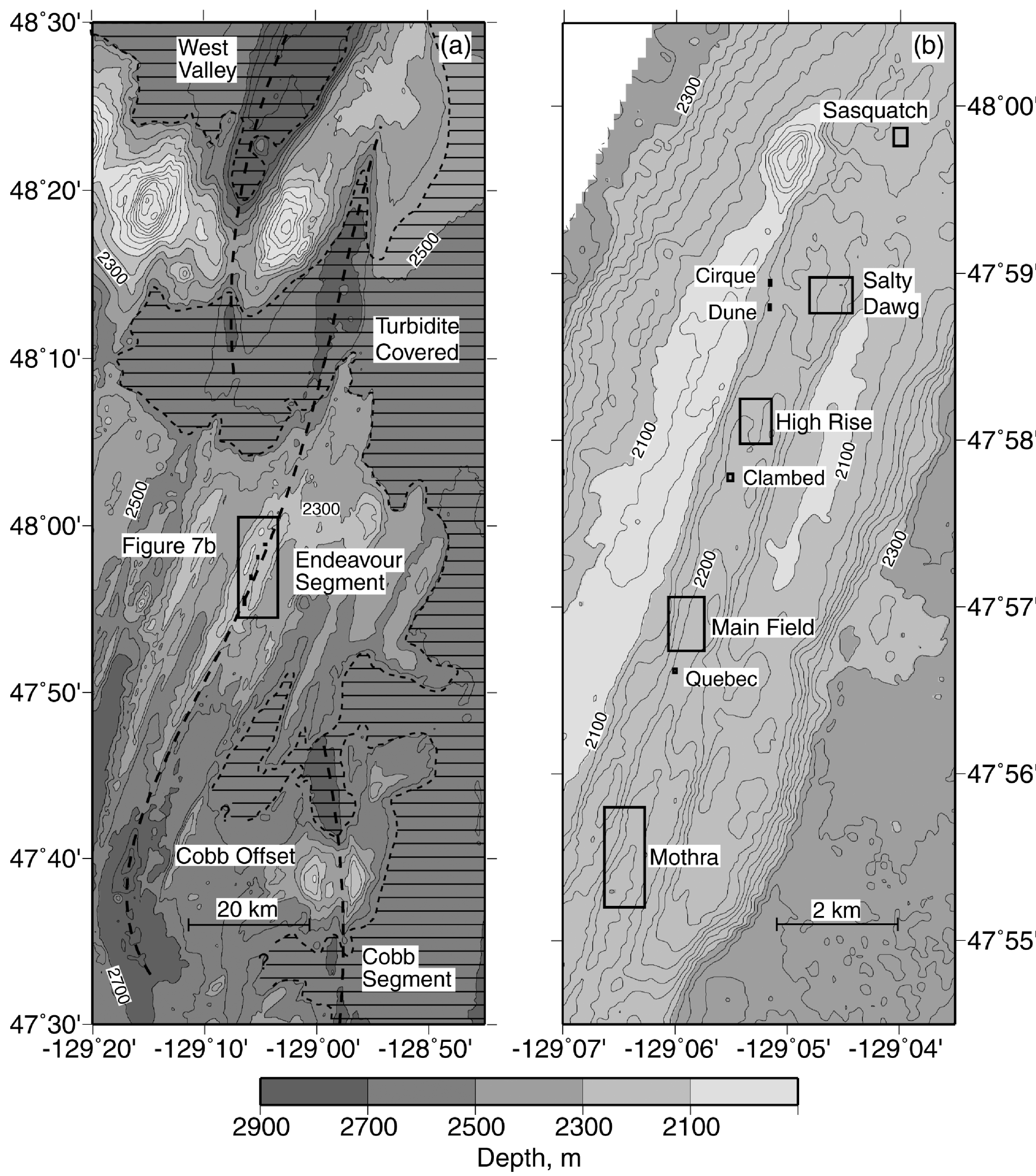

Figure 3. (a) Bathymetric map of the Endeavour segment of the Juan de Fuca Ridge, contoured at 100 m, showing the location of the ridge axis (dashed lines) and the area covered by Figure 3b (box). The hachured region shows the approximate limits of thick turbidite sediments based on a simple interpretation of the bathymetric features. (b) Detailed map of the central portion of the Endeavour contoured at 25 m showing the location of vent fields (bold boxes). The five known high-temperature vent fields are indicated by labels with a larger font size; smaller fonts are used for low temperature fields.

[1997] argued that hydrothermal heat extraction on the Endeavour occurs along an irregular cracking front that is actively penetrating hot rock at mid-crustal depths. A recent seismic reflection survey [*Carbotte et al.*, 2002] shows that the Endeavour is underlain by a magma chamber at 2.3-2.6 km depth and this result is interpreted as evidence that the hydrothermal systems here and on the East Pacific Rise are driven by heat conduction from the magma lens [*Detrick et al.*, 2002].

In 1999, the hydrothermal systems were invigorated and their chemistry perturbed by a large seismic swarm [*Johnson et al.*, 2000a; *Lilley et al.*, 2003]. This led to transient changes in pressures at boreholes up to 100 km away some of which occurred synchronously with the largest earthquake [*Davis et al.*, 2001]. However, temperature perturbations measured at existing vent sites on axis were delayed several days [*Johnson et al.*, 2000a] just as for the 1995 event on the East Pacific Rise [*Sohn et al.*, 1998]. Since earthquakes on land often have instantaneous impacts on continental hydrothermal systems, *Dziak and Johnson* [2002] infer that marine hydrothermal systems are hydrologically unique. However, this inference may be premature given the relative paucity of observations. Earthquakes are likely to lead to near instantaneous changes in pressure and hence hydrothermal discharge rates on the ridge axis but no observations have been made. Earthquakes may also create new pathways to shallow regions of hot fluid and this will also produce a near instantaneous response. However, within established vents, the thermal anomalies created by changes in the configuration of the heat uptake zone will propagate to the surface at a speed intermediate between the fluid velocity and the Darcy velocity depending on the proportion of wall rock that remains in thermal equilibrium with the fluid. A full suite of temperature, chemical and flow measurements and seafloor observations are required to fully characterize the impact of earthquakes on mid-ocean ridge hydrothermal systems.

There have been many attempts to determine hydrothermal heat fluxes from the Endeavour. All but one are based on summing sources on the seafloor or averaging the heat in the neutrally buoyant hydrothermal plume as it is transported by ocean currents. The resulting estimates vary from about ~100 MW to over ~10,000 MW and all have uncertainties similar to the estimates themselves [*McDuff*, 1995]. Most recently, *Stahr et al.* [2003] have developed a precise technique that employs an autonomous underwater vehicle to measure the fluxes carried by the buoyant plume across the boundaries of a control volume. They estimate a hydrothermal heat flux of 600 ± 100 MW for the Main Vent field.

Knowledge of the heat flux estimates places strong constraints on the large scale permeability. If the permeability is assumed uniform, most of the hydraulic resistance is in the upflow zone because hydrothermal discharge is focused. By further assuming that the cross-sectional area of the upflow zone is invariant with depth and thus equal to the area of the vent field, it is straightforward to estimate permeability [*Lowell and Germanovich*, 1994; *Wilcock and McNabb*, 1995]. Calculations for the Main Endeavour field [*Wilcock and McNabb*, 1995] updated with the latest heat flux estimate yield a permeability of 4×10^{-13} m^2 with a formal uncertainty of about a factor of two. If the permeability is heterogeneous, the permeability will exceed this value in regions of concentrated flow. One can conclude that the average permeabilities beneath vigorous high-temperature systems are similar to those measured in layer 2A off-axis and much higher than those measured off axis at greater depths (Plate 3).

As along the East Pacific Rise, the configuration of the circulation cells is poorly known. *McDuff et al.* [1994] argue that spatial variations in fluid temperatures and compositions in the Main field are most simply explained by a model in which downflow is concentrated between vent fields (Figure 4a). *Wilcock and McNabb* [1995] interpret the elongated shape of vent fields as evidence for an anisotropic permeability that would favor circulation cells oriented predominantly along axis. However, the chemistry of vent fluids on the Endeavour requires an organic source that is most likely sedimentary [*Lilley et al.*, 1993]. Unless there is a substantial buried sediment layer beneath the ridge axis, recharge must occur outside the axial valley. On the basis of heat flow measurements made on the ridge flank, *Johnson et al.* [1993] argue that convection cells are oriented across axis with recharge down inward-dipping listric faults that intersect the seafloor several kilometers off axis (Figure 4b). Another intriguing explanation is that the recharge occurs well to the north in a region where the turbidite sediments overflow the ridge axis (Figure 4c). This third explanation may seem unlikely at first, but there is good evidence on the eastern flank of the Juan de Fuca Ridge that crustal fluids travel tens of kilometers in the shallow subsurface between points of recharge and discharge [*Davis et al.*, 1999; *Elderfield et al.*, 1999; *Fisher et al.*, 2003].

Another topic of uncertainty in mid-ocean ridge hydrothermal systems is the fate of brines produced by phase separation. *Bischoff and Rosenbauer* [1989] have argued that high-temperature systems organize into a two-layer double diffusive system in which a single-pass seawater cell overlies a high-temperature brine cell. Numerical modeling studies suggest that such a configuration may be unstable [*Schoofs and Hansen*, 2000]. On volcanically active segments, diking-eruptive events are clearly associated with a progression from vapor-dominated vent fluids immediately after the eruptions to brine-dominated counterparts months to a few years later [*Butterfield et al.*, 1997], but this only requires temporary brine storage and not necessarily the formation of an organized two layer system. On the Endeavour, the Main field has been venting fluids with salinities below seawater for over 15 years. Assuming a steady heat flux of 600 MW [*Stahr et al.*, 2003] and vent fluids that average 360ºC with a salinity that is 75% of seawater [*Butterfield et al.*, 1994], the rate of subsurface brine accumulation is ~10 kg s^{-1} for a 25 wt% NaCl brine. Assuming a mid-crustal porosity of 2%, a brine layer occupying an area of 1 km^2 would thicken at 15 m/year. Either a substantial volume of brine is accumulating beneath the Main vent field or brine is being transported

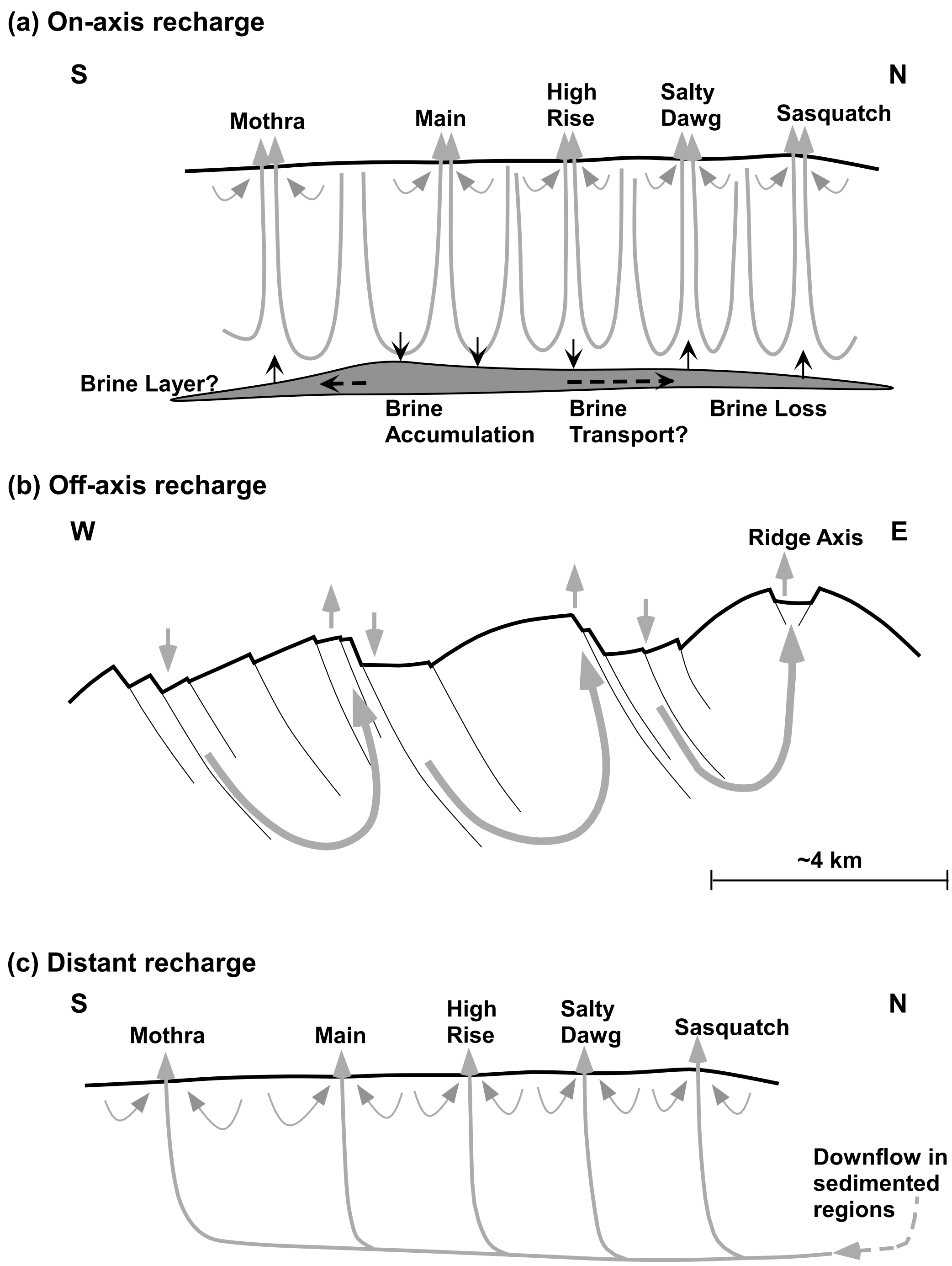

Figure 4. Schematic diagrams illustrating alternative models of hydrothermal circulation beneath the central portion of the Endeavour segment. Circulation cells may be oriented (a) along-axis with recharge between high-temperature vent fields [*McDuff et al.*, 1994], (b) across-axis with recharge along inward-facing normal faults [*Johnson et al.*, 1993] or (c) may involve recharge well to the north where sediments overlap the ridge axis. The fluids venting from the Main field have salinities below seawater and brine is either accumulating in a growing brine layer beneath or migrating laterally to vent in the Mothra, Salty Dawg and Sasquatch fields. This process is illustrated in (a) but is also compatible with the patterns of circulation shown in the other diagrams.

laterally to the Mothra, Salty Dawg, and Sasquatch fields, all of which have vent fluids with salinities above seawater (Figure 4a). Either explanation suggests the presence of an organized and at least quasi-stable brine layer beneath the ridge axis.

TAG Site

The Trans-Atlantic Geotraverse (TAG) hydrothermal site (Figure 5), near 26°N 45°W on the Mid-Atlantic Ridge (MAR), has become a type example of hydrothermal activity on a slow-spreading (tectonically-dominated) mid-ocean ridge. The TAG site contains several massive sulfide deposits, the largest being at the main, active mound (Figure 6), that were emplaced on and within volcanic rocks, making the site particularly interesting because of what it may tell us about economically important, ophiolite-hosted sulfide deposits on land.

The TAG segment is a 40-km-long segment trending NNE and is bounded at either end by non-transform discontinuities (Figure 5) [*Purdy et al.*, 1990]. The spreading rate is 24 mm/yr and slightly asymmetric with higher rates to the east [*McGregor et al.*, 1977]. Seafloor morphology of this segment is consistent with that along other segments between the Kane and Atlantis Fracture Zones. Long spreading segments north and south of TAG are characterized by classic "bulls-eye" mantle Bouguer gravity anomalies interpreted to indicate focused magmatic accretion, but the TAG segment is short and the associated gravity low has a relatively small amplitude [*Fujimoto et al.*, 1996]. A study of microseismicity along the TAG segment revealed that most activity is concentrated below the axial high [*Kong et al.*, 1992], but there was no seismicity detected immediately below the active, high-temperature mound. Analysis of TAG basaltic glasses recovered by dredging suggests that primary magmas experienced crystallization at relatively great depths below the seafloor, and then rose rapidly to the surface, indicating a deep-seated heat source for hydrothermal circulation [*Meyer and Bryan*, 1996]. This interpretation is consistent with the relatively large depths of microearthquakes along this ridge segment [*Kong et al.*, 1992]. Surface surveying revealed a broad region of low magnetic intensity interpreted to indicate widespread hydrothermal alteration of basaltic crust [*Wooldridge et al.*, 1992].

The TAG hydrothermal field is located between the central axis of spreading and the eastern wall of the median valley. The field (Figure 5b) comprises the main TAG mound and three other hydrothermal regions, the Alvin Zone which contains one active and six inactive low- temperature mounds, the MIR zone which comprises discontinuous inactive sulfides, and the low temperature zone which may or may not be active [*Rona et al.*, 1993; *Humphris and Tivey*, 2000]. The main TAG mound is circular in plan view, 200 m

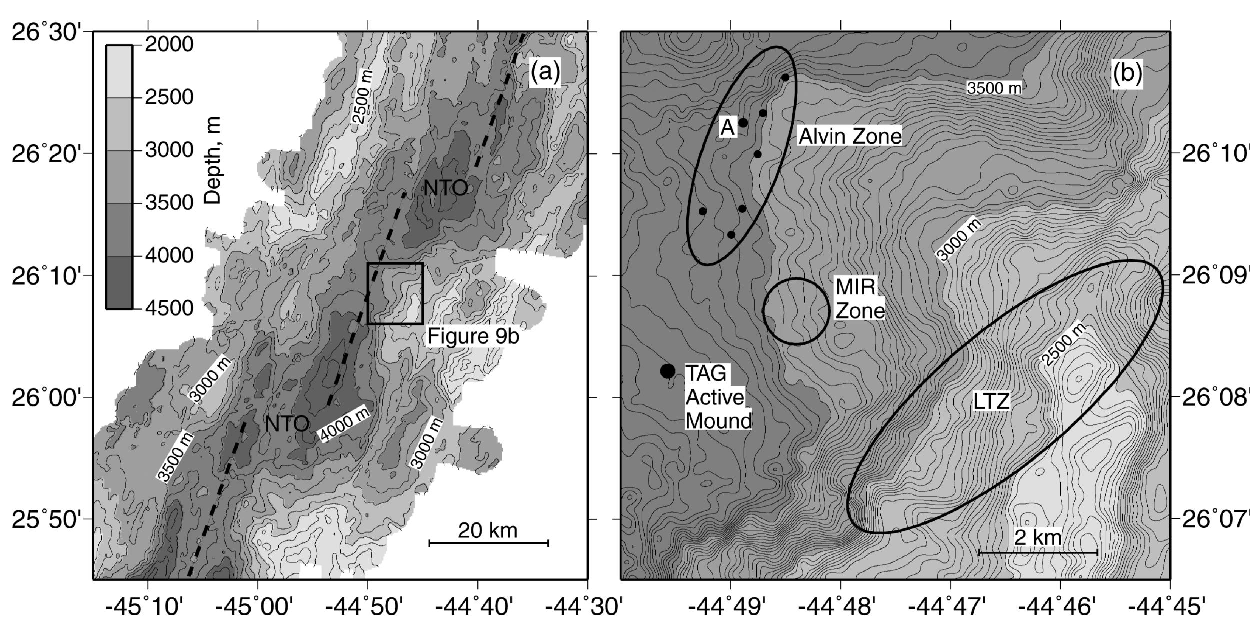

Figure 5. (a) Bathymetric map of the TAG region of the Mid-Atlantic Ridge, contoured at 250 m, showing the approximate location of the spreading center (dashed lines), the non-transform offsets (labeled NTO) which delimit the TAG segment [*Tucholke et al.*, 1997] and the area covered by Figure 5b (bold box). (b) Detailed map of the central portion of the TAG segment (based on Figure 1 of *Humphris and Tivey* [2000]), contoured at 25 m, showing the location of all identified hydrothermal mounds (filled circles) and zones (bold lines) including the low-temperature zone (LTZ). Within the Alvin zone only one mound marked "A" is known to be active.

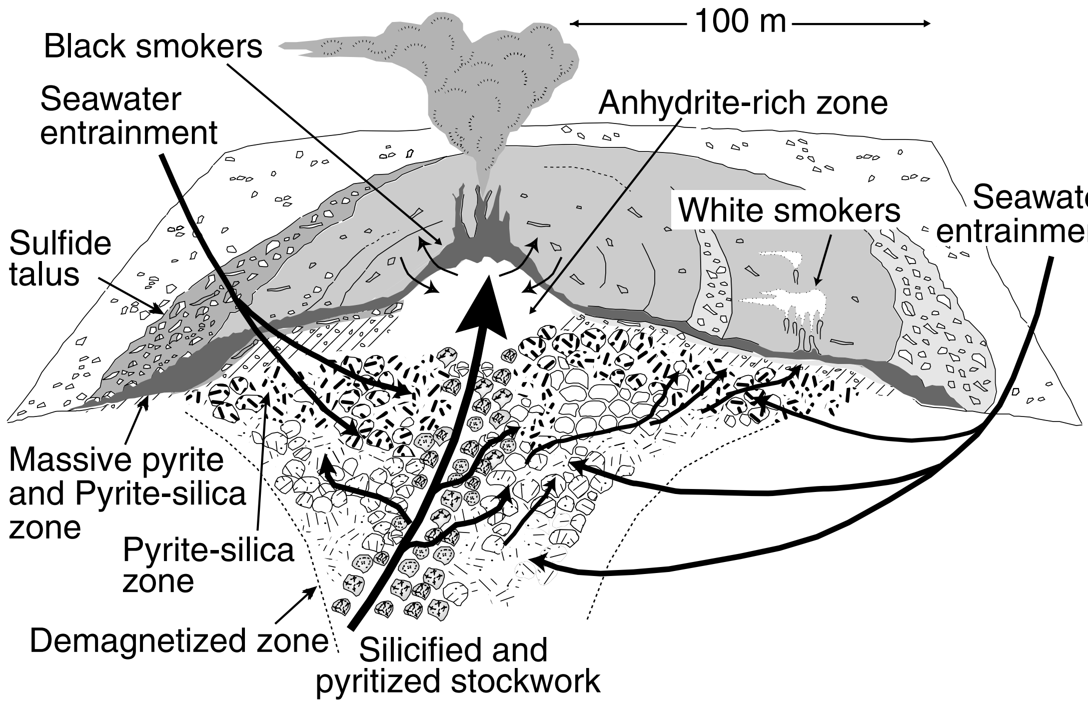

Figure 6. Schematic model of the TAG active mound illustrating surface and subsurface structures and likely patterns of fluid flow (modified from *Humphris and Tivey* [2000]). Note the narrowness of the stockwork zone below the mound, and the anhydrite-rich zone that made drilling conditions difficult and limited recovery during ODP Leg 158. Seawater entrainment close to the mound, as shown with arrows, is a form of secondary convection, but the primary flow system that extracts heat and mass from the host rock extends much deeper below the mound and could be recharged a considerable distance away. The cartoon shows locations of black and white smoker vent areas, but there are also regions along the mound margins where warm, shimmering water discharges across the seafloor.

in diameter, and rises 50 m above the surrounding seafloor. The mound is composed mainly of massive sulfides and anhydrite [*Herzig et al.*, 1998]. Analysis of hydrothermal deposits from the TAG field indicates that the area has been hydrothermally active (on and off) for about the past 140 kyr [*Rona et al.*, 1993], making this the longest-lived seafloor hydrothermal deposit on record. The main mound includes black-smoker vents near the crest with fluid temperatures of 350-360°C, another vent area towards the SE side of the mound with white smokers and temperatures of 260-300°C, and a region along the eastern edge of the mound through which warm, shimmering water emerges from the seafloor.

The main TAG mound (Figure 6) was drilled during ODP Leg 158. Penetration was limited (maximum subseafloor depth = 125 m) largely because of difficulty in establishing and maintaining an open hole within unstable sulfide and fractured basalt. Anhydrite has a retrograde thermal stability; it dissolves at temperatures close to bottom water temperatures but is stable at elevated temperatures. This made penetration by drilling very difficult, since cold seawater pumped as a drilling fluid caused dissolution and collapse of the borehole; these processes also led to limited recovery of samples during Leg 158. Recovery was sufficient to define the lateral extent of the shallow stockwork zone below the mound; it is remarkably narrow, probably not more than 30-50 m in diameter [*Hannington et al.*, 1998], a result consistent with near-bottom gravity profiling [*Evans*, 1996]. Basement alteration below the mound is not pervasive, indicating that flow and water/rock interaction are heterogeneous [*Herzig et al.*, 1998]. On the other hand, a detailed survey of seafloor heat flow suggests that heat is advected within the shallow subsurface to distances of 20-50 m away from the primary discharge areas [*Becker et al.*, 1996]. Heat flow is also extremely variable in the immediate vicinity of the most vigorous vent sites, suggesting that secondary circulation may result in cycling of bottom water through shallow sediments and sulfide.

Several studies have noted the association of the main TAG mound with two distinct structural trends, ridge-parallel faults and ridge-oblique faults [*Karson and Rona*, 1990; *Kleinrock and Humphris*, 1996]. It was suggested that the mound was located where the intersection of these fault groups enhanced permeability and allowed rapid extraction of crustal heat and dissolved minerals. There is little evidence for recent shallow magmatism near the TAG mound and some workers have argued that the heat source for this system comes from the downward migration of a cracking front into the lower crust [*Wilcock and Delaney*, 1997]. *Humphris and Cann* [2000] examined the energy and chemical budgets of venting and sulfide deposition at the active TAG mound. Their calculations indicate that the TAG mound formed during a few short episodes of vigorous venting whose cumulative duration was a few hundred to at most thousands of years. They note that similar massive sulfide deposits in ophiolites are underlain by narrow stockworks and an alteration pipe that roots in a leached reaction zone near the dike-gabbro boundary. By analogy they conclude that the primary energy source for the formation of the mound is the release of latent heat associated with magmatic intrusions, and they proposed that previous intervals of intense hydrothermal activity likely corresponded to individual magmatic episodes.

There is no way to trace the ultimate source of hydrothermal heat, but *Humphris and Cann* [2000] argue that observations in ophiolites preclude the pervasive water/rock interaction at depth required to extract sufficient heat along a cracking front. This interpretation may be reconciled with the observation that deep fluid penetration seems to be required at several spreading centers in order to explain the depth of seismic activity and strong vertical gradients in seismic velocities close to the ridge. If deep fluid flow is not pervasive, then the regions close to the channels through which most of the fluid moves will experience high water-rock ratios, while alteration of the bulk of the crust is dominated by diffusion. At the time scales of interest (~10,000 yr), conduction of heat is sufficiently efficient over length scales less than ~1 km that we would not be able to distinguish between focused flow and a broader distribution of flow paths.

FUTURE DIRECTIONS

The preceding discussion illustrates how geophysical studies have often focused on large-scale patterns of hydrothermal circulation; relations between hydrothermal, magmatic and tectonic processes; and the ultimate nature of the heat source(s) responsible for observed fluxes. These studies have contributed towards a qualitative understanding of crustal accretion and have helped constrain quantitative models of the large-scale thermal and mechanical structure of young oceanic lithosphere [e.g., *Chen and Morgan*, 1990; *Phipps Morgan and Chen*, 1993; *Cochran and Buck*, 2001]. But there remain many unanswered questions regarding the geometry and distribution of hydrothermal flows, the formation and fate of brines, and critical linkages between crustal architecture, magma input, tectonic modification, heat extraction and fluid circulation. Without the ability to make in situ observations at depth, our understanding of deep hydrological processes is likely to be limited, particularly at finer scales.

We believe that geophysical observations can contribute most to hydrological studies of mid-ocean ridge systems if they first focus on shallow processes in the vicinity of upflow zones. Such work is particularly relevant to studies of the subseafloor biosphere because these regions are close to the sites where fluids are sampled and are likely the most hospitable to life. Layer 2A has the highest porosity and is most likely to host the physical processes that promote chemical disequilibria. The thermal gradients which cool hydrothermal fluids conductively are highest near the seafloor. The high and heterogeneous permeability of the shallow crust will promote dispersive mixing between hydrothermal fluids and seawater that circulates in secondary cells. In addition, the loading of ocean tides combined with local variations in (1) fluid properties due to temperature (including phase separation), (2) elastic properties due to variable compaction and cementation, and (3) permeability due to complex fracture geometries will lead to time-dependent pressure gradients in shallow regions, and hence an oscillatory component of flow that will further enhance dispersive mixing [e.g., *Davis et al.*, 2000]. Geophysical studies focused on the shallow crust have the additional advantage that they allow comparison of interpretations based on independent methods, and they may be tested through direct observation.

There are at least three developments that will significantly enhance the role of geophysical studies in constraining shallow hydrological processes. First, advances in remotely operated and autonomous underwater vehicles coupled with improvements in seafloor navigation have enhanced capabilities to survey vent fields. Applications include imaging the seafloor [*Bradley et al.*, 1999], mapping hydrothermal plumes [*Stahr et al.*, 2003], and collecting high resolution potential field data to constrain the subsurface dimensions of upflow [*Tivey and Johnson*, 2001]. Seismic refraction experiments with well-navigated on-bottom sources have so far been limited to linear profiles with one or two recorders [e.g., *Christeson et al.*, 1994], but the capability exists to deploy more recorders and shots to obtain high-resolution seismic images of hydrothermal upflow zones. Ongoing improvements in seafloor electromagnetic techniques may eventually lead to high-resolution three-dimensional images of shallow resistivity structure [*Evans et al.*, 1998].

Second, there is a movement within the mid-ocean ridge research community towards the development of long-term hydrothermal observatories at which many processes are monitored simultaneously. On mid-ocean ridges this has already led to (1) coordinated response efforts to monitor the evolution of hydrothermal output in the aftermath diking events [e.g., *Fox*, 1995; *Delaney et al.*, 1998]; (2) observations of earthquake triggered perturbations in hydrothermal outflow [*Sohn et al.*, 1998; *Johnson et al.*, 2000a]; and (3) measurements of the effects of tides on the temperatures, fluxes, and chemical compositions of hydrothermal discharge [e.g., *Tivey*, 2001]. Observatories will dramatically improve the quality and size of data sets available to constrain subsurface hydrological processes and these should motivate more sophisticated modeling efforts. Quantitative models will be required to address many problems including (1) the effects of two-phase flow on the geometry of hydrothermal circulation; (2) the spatial and temporal relationships between diffuse and high temperature upflow; the relative importance of dispersive mixing and conductive heating in diffuse flows; (3) the poroelastic and thermal response of hydrothermal systems to geological events and tidal loading; and (4) the long term relationships between hydrothermal fluxes and the physical and chemical processes that modify permeability.

Finally, although progress in drilling, sampling, and borehole experiments at bare-rock sites has been slower than desired, continuing technological development has created new opportunities for working in these dynamic environments. Drilling on bare rock has provided some of the most persistent challenges of scientific drilling in the oceans, but it remains a high priority of the new Integrated Ocean Drilling Program, scheduled to begin in 2003. Promising technologies include active heave compensation (allowing better control of weight applied to the formation during drilling), hammer-in and drill-in casing systems (allowing establishment and maintenance of open holes within fractured formations), and multilevel borehole observatory systems. Boreholes drilled in and around hydrothermal upflow zones will have many benefits. They will facilitate the sampling of subsurface fluids and a host of in situ experiments. Single boreholes will allow direct measurements of permeability and provide a means to introduce chemical tracers that can be used to infer flow paths and dispersive mixing. Experiments between adjacent boreholes will facilitate hydraulic measurements over longer length scales and will be required to estimate storage properties.

Geophysical techniques have yielded critical views of seafloor hydrothermal systems during the past several decades. Our challenge in the coming decades is to exploit conventional and new geophysical techniques, and to combine and reconcile the results with data from geological, geochemical and biological experiments so as to create a more complete, accurate, and nuanced view of seafloor hydrothermal processes. To be successful, we must take advantage of existing opportunities, continue to push technological development, and risk working in one of the most complex, dynamic environments on the planet.

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