Borehole-to-borehole hydrologic response across 2.4 km in the upper oceanic crust: Implications for crustal-scale properties


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Subseaflow hydrologic observatories (CORKs) were installed in four boreholes in young seafloor on the eastern flank of the Juan de Fuca ridge to evaluate the hydrogeology of the upper oceanic crust. Two CORKs installed at Site 1301 were incompletely sealed, allowing cold bottom water to flow into basement at 2–5 L/s and causing a pressure perturbation in a preexisting sealed CORK at Site 1027, which was 2.4 km away. The pressure perturbation at Site 1027 is analyzed using conventional aquifer test methods, yielding transmissivity of $T = 0.5$ to $1.2 \times 10^{-2}$ m$^2$/s and bulk permeability of $k = 0.7$ to $2 \times 10^{-12}$ m$^2$ in the upper 300 m of basement. Storativity (a parameter that includes fluid and aquifer compressibilities, porosity, and layer thickness) is $S = 1$ to $3 \times 10^{-3}$, corresponding to a crustal aquifer (matrix) compressibility of $\beta_m = 3 \times 9 \times 10^{-10}$ Pa$^{-1}$. The inferred basement permeability is consistent with, but at the low end of, permeabilities calculated from single-hole packer experiments at this and other young oceanic crustal sites; it is much less than values estimated from numerical models, analyses of formation response to tidal pressure oscillations, or pressure responses to coseismic strain events. The relatively low permeability indicated by the cross-hole response may result from the basement aquifer in this area being azimuthally anisotropic, with a preferential flow direction oriented subparallel to the abyssal hill topography and tectonic structural fabric created at the ridge axis; this hypothesis will be tested during future experiments.


1. Introduction

[2] Fluid flow within the volcanic oceanic crust influences the thermal and chemical state and evolution of oceanic lithosphere and lithospheric fluids; establishment and maintenance of subseaflow microbial ecosystems; diagenetic, seismic, and magmatic activity along plate-boundary faults; creation of ore and hydrate deposits both on and below the seafloor; and exchange of fluids and solutes across continental margins [e.g., Alt, 1995; Huber et al., 2006; Parsons and Sclater, 1977; Peacock and Wang, 1999]. The global hydrothermal fluid mass flux through the upper oceanic crust rivals the global riverine fluid flux to the ocean, and effectively passes the volume of the oceans through the crust once every 10$^6$–10$^8$ [Elderfield and Schultz, 1996; Johnson and Pruis, 2003; Mottl, 2003]. Most of this flow occurs at relatively low temperatures, far from volcanically active, seafloor spreading centers where new ocean floor is created.

This “ridge flank” circulation can be influenced by off-axis volcanic or tectonic activity, but is driven mainly by the rise of lithospheric heat from below the crust. Although the average maximum age at which measurable heat is lost advectively from oceanic lithosphere is 65 Ma [Parsons and Sclater, 1977], many sites remain hydrologically active for tens of millions of years beyond this age, with circulation largely confined to basement rocks redistributing heat below thickening sediments [Fisher and Von Herzen, 2005; Von Herzen, 2004].

[3] Despite its importance, many of the fundamental parameters that control and are influenced by fluid flow in the oceanic crust remain largely unquantified, including permeability [e.g., Fisher, 2004; Ingebritsen and Sanford, 2006; Lister, 1974; Lowell, 1991; Wilcock and Fisher, 2004]. Permeability is commonly considered to be an intrinsic rock property, an empirically derived coefficient of proportionality that relates driving forces to rates of fluid flow. In fact, permeability is a highly variable (spatially and, in some cases, temporally) tensor property, the values of which differ depending on the measurement method, spatial scale, and dimensionality of measurement or representation [e.g., de Marsily et al., 2005; Guéguen et al., 2006; Illman, 2006; Neuman and Di Federico, 2003; Proce et al., 2004; Sanchez-Vila et al., 2006]. Resolving spatial and temporal aspects of the permeability distribution and flow dynamics in the oceanic crust requires tests in which system responses are

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monitored at simultaneously at multiple locations. These tests can also help to resolve the hydrogeologic storage properties of the ocean crust, which are important for understanding the penetration and influence of fluid pressure perturbations and other transient phenomena.

In the present study we evaluate the hydrologic properties of the upper oceanic crust using an inadvertent cross-hole experiment. This experiment was initiated during installation of a subseafloor observatory network [Fisher et al., 2005a]. The full suite of planned experiments awaits completion of the observatory network, but the observed cross-hole response allows quantitative assessment of transmissive and compressive properties of the upper oceanic crust, and will help to guide future observational and modeling experiments.

2. Background: Geological Setting and Borehole Observatories

2.1. Eastern Flank of the Juan de Fuca Ridge

Numerous studies summarize geology, geophysics, and basement-fluid chemistry and hydrogeology within young seafloor on the eastern flank of the Endeavour segment of the Juan de Fuca Ridge (JFR) [e.g., Becker et al., 2000; Davis et al., 1992b; Elderfield et al., 1999; Hutnak et al., 2006; Mottl et al., 1998; Wheat and Mottl, 1994]. Topographic relief associated with the JFR axis and abyssal hill bathymetry on the ridge flank has helped to trap turbidites flowing from the continental margin to the east (Figure 1). This has resulted in burial of young oceanic basement rocks under thick sediments. Sediment cover is sparse and oceanic basement is exposed near the spreading center. The sediment becomes thicker and more continuous to the east, with basement exposed at only a few, isolated outcrops. Basement relief is dominated by linear ridges and troughs oriented subparallel to the spreading center and produced mainly by faulting, variations in magmatic supply at the ridge, and off-axis volcanism. Low-permeability sediment limits advective heat loss across most of the ridge flank, resulting in strong thermal, chemical, and alteration gradients in basement.

Ocean Drilling Program (ODP) Leg 168 drilled a transect of eight sites on 0.9 to 3.6 Ma seafloor east of the JFR [Davis et al., 1997a]. ODP Leg 168 also installed four long-term borehole observatories in the upper crust, two of which are located on 3.5–3.6 Ma seafloor near the eastern end of the drilling transect, in Holes 1026B and 1027C (Figure 1) [Davis et al., 2000; Davis and Becker, 2002]. Integrated Ocean Drilling Program (IODP) Expedition 301 returned to this area in 2004, replaced the borehole observatory in Hole 1026B, and drilled two additional boreholes and installed observatories in Holes 1301A and 1301B, 1 km to the south of Hole 1026B (Figure 1). Operations during IODP Expedition 301, and subseafloor pressure observations from the 13 months following this expedition, provide the basis for the current study.

2.2. Borehole Observatories

Borehole observatories installed during ODP and IODP ("CORKs") were designed to seal open holes so that thermal, pressure, chemical conditions could equilibrate following the dissipation of the drilling disturbance; to facilitate collection of fluid and microbiological samples and temperature and pressure data using autonomous samplers and data logging systems; and to serve as long-term monitoring points for large-scale crustal processes [Becker and Davis, 2005; Davis et al., 1992a; Fisher et al., 2005b]. CORKs installed to monitor conditions in oceanic basement generally include a seafloor reentry cone and casing hanger(s); two to four concentric (nested) casing strings that penetrate through sediments and allow access to underlying basement; a series of seals (both between casing strings and between casing and the formation) that hydraulically isolate the open crustal interval at depth from the
overlying ocean; downhole and seafloor instrumentation for collection of samples and data; and a seafloor wellhead that includes valves, fittings, electrical connections and a landing platform so that the observatory can be serviced by submersible or remotely operated vehicle (ROV), allowing samples and data to be retrieved without recovery of the complete observatory assembly.

[8] Expedition 301 CORKs and the preexisting CORK in Hole 1027C were visited with the ROV ROPOS soon after the drilling expedition in September 2004, and again in September 2005 with the submersible Alvin. Data recovered during those dives showed that the Hole 1026B observatory was sealed and operating as intended, but the observatories in Holes 1301A and 1301B were not sealed properly. Incomplete sealing allowed cold ocean-bottom water to flow into the formation following CORK installation. The Hole 1301B CORK was at least partly sealed in summer 2007 with pressure data recovered from Hole 1026B during Leg 168 included a data logger, pressure sensors, thermistors at multiple depths, and a fluid sampler, all of which were recovered in 1999. As of the start of Expedition 301 the observatory was open, and warm (65°C) reacted basement fluid vented freely through the top of the wellhead because shallow crustal fluids are overpressured with respect to ambient hydrostatic conditions [e.g., Davis et al., 1997b; Davis and Becker, 2002; Fisher et al., 1997]. In fact, the original Hole 1026B CORK was never completely sealed after being installed in 1996, but discharged fluid for years until it was replaced during Expedition 301. Pressure data recovered from Hole 1026B following installation of a new CORK system on Expedition 301 indicate that the new observatory is properly sealed. Pressure in Hole 1026B is recovering slowly from the

cdrilled, and there was no geophysical logging in basement. Expedition 301 included greater basement penetration and collection of geophysical logs, but core recovery was again highly incomplete (Figure 2). Low core recovery and difficulty with geophysical logging are common problems in uppermost basement rocks, which typically comprise rubbly and highly fractured basalt pillows and flows [e.g., Anderson et al., 1985a; Bartetzko et al., 2001; Bartetzko and Fisher, 2008; Moos, 1990].

[10] Hole 1026B was drilled to 295 meters below seafloor (mbsf), cased across the sediment-basement interface, and extended to 48 m subbasement (msb) during Leg 168 [Shipboard Scientific Party, 1997]. Site 1026 is located above a buried basement high, the same abyssal hill upon which basement outcrops are located ~8 km to the north and south of the drill site (Figure 1). The CORK installed in Hole 1026B during Leg 168 included a data logger, pressure sensors, thermistors at multiple depths, and a fluid sampler, all of which were recovered in 1999. As of the start of Expedition 301 the observatory was open, and warm (65°C) reacted basement fluid vented freely through the top of the wellhead because shallow crustal fluids are overpressured with respect to ambient hydrostatic conditions [e.g., Davis et al., 1997b; Davis and Becker, 2002; Fisher et al., 1997]. In fact, the original Hole 1026B CORK was never completely sealed after being installed in 1996, but discharged fluid for years until it was replaced during Expedition 301. Pressure data recovered from Hole 1026B following installation of a new CORK system on Expedition 301 indicate that the new observatory is properly sealed. Pressure in Hole 1026B is recovering slowly from the

| Table 1. Hole and Subseafloor Observatory Specifications |
|-----------------------------|-----------------------------|-----------------------------|-----------------------------|-----------------------------|
| Hole 1026B | Hole 1027C | Hole 1301A | Hole 1301B |
| Longitude | 47° 45.757' N | 47° 45.387' N | 47° 45.210' N | 47° 45.228' N |
| Latitude | 127° 45.548° W | 127° 43.867° W | 127° 45.833° W | 127° 45.827° W |
| Seafloor depth, mbfl | 2658 | 2656 | 2667 | 2668 |
| Sediment thickness, m | 247 | 576 | 262 | 265 |
| Total depth, mbfl | 295° | 632 | 370 | 382 |
| Total depth, mbfl | 48° | 56 | 108 | 318 |
| Thickness of open basement interval, m | 21° | 54° | 104 | 312 |
| Observatory emplaced | ODP 168, 8/7/96 | ODP 168, 7/21/04 | IODP 301, 7/21/04 | IODP 301, 8/16/04 |

| Table 2. Chronology of IODP Expedition 301 Basement Operations, Holes 1026B, 1301A, and 1301B |
|-----------------------------|-----------------------------|-----------------------------|-----------------------------|
| Hole 1026B | Hole 1301A | Hole 1301B |
| Upper basement drilling, casing | NA | 3–8 and 17–18 July 2004 | 12–17 and 21 July 2004 |
| Upper basement coring | NA | NA | 21–31 July 2004 |
| Packer experiments | NA | 18–19 July 2004 | 9–10 Aug. 2004 |

See Fisher et al. [2005a].

*Includes drilling without coring of upper 100 m of basement, cementing of casing, and waiting 7–10 days to drill out cement prior to other operations in upper basement.

Only in Hole 1301B; extends 318 m subbasement (see Table 1).


CORK subseafloor observatory installed in wellhead.
thermal perturbation associated with many years of upflow of warm formation fluid [Davis and Becker, 2007].

Hole 1027C was also drilled during Leg 168, 2200 m to the east of Hole 1026C, where sediment thickness is 575 m above a buried basement low (Figure 1). Hole 1027C penetrated to 632 mbsf. The upper part of the hole was cased through sediments and uppermost basement, with 54 m of open hole. The open interval near the base of Hole 1027C comprises a diabase sill, intercalated sediments, and basalt breccia overlying 26 m of extrusive volcanic rocks [Shipboard Scientific Party, 1997]. A CORK was installed in Hole 1027C during Leg 168, including a data logger, pressure sensors, thermistors at multiple depths, and a fluid sampler. These instruments were retrieved in 1999 to allow recovery of a basement fluid sampler, and the pressure logging system was replaced. In contrast to Hole 1026B, Hole 1027C was underpressured with respect to ambient hydrostatic conditions (−26 kPa) [Davis and Becker, 2002].
Hole 1027C was fully sealed and recorded formation pressure before, during, and after Expedition 301.

Site 1301 is located 1 km south of Hole 1026B above the same buried basement ridge (Figure 1). Hole 1301A was drilled through 262 m of sediment and 108 m into upper basement. Hole 1301B was drilled 35 m to the NNE of Hole 1301A, through 265 m of sediment and 318 m into basement (Table 1; Figure 2). Both of the Site 1301 boreholes contained four nested casing strings: 0.50-m (diameter) casing in the uppermost sediments, 0.41-m casing extending just across the sediment-basement interface, 0.27-m casing extending into basement, and a 0.11-m inner CORK casing that houses instrument strings and plugs (Table 1; Figure 2). The two largest casing strings were sealed by collapse of unconsolidated sediments and (for the 0.41-m string) cement across the sediment-basement interface.

The annulus between 0.41-m and 0.27-m casing strings at Site 1301 was supposed to contain a rubber, mechanical casing seal near the seafloor, but this component was not available for use during Expedition 301, and an attempt was made to seal the 0.27-m casing strings at depth with cement. Rubbly basement prevented this cement from sealing between casing and the borehole wall, and operations were additionally complicated in Hole 1301B by the separation of the unwelded 0.27-m casing string into two sections, leaving a gap just above the sediment-basement interface (Figure 2). The CORK installed in Hole 1301A included a casing packer (as part of the 0.11-m inner casing) that was set inside 0.41-m casing. In contrast, the CORK installed in Hole 1301B included two casing packers set in open hole, intended to hydraulically isolate sections of the upper crust [Fisher et al., 2005b].

Pressure data recovered from the Site 1301 CORKs in 2005 confirm that these observatories were incompletely sealed. The lack of a casing seal allows water to enter the annular space between the 0.41-m and 0.27-m casings through (at the most restrictive part of the path) sixteen, 0.030-m-diameter passages. Pressure data were also recovered in 2005 from the CORK in Hole 1027C. This system was fully sealed and provided a long-term record of pressure perturbations associated with operations during Expedition 301 and the subsequent 13 months. These pressure data are the basis for the analyses presented in the following sections.

3. Experimental Procedures, Constraints, and Results

3.1. Pressure Observations, Filtering, and Response to Nearby Basement Operations

The long-term record of seafloor and subseafloor pressure data from Site 1027 illustrates several important characteristics (Figure 3). First, unfiltered pressure data are strongly influenced by tides and other oceanographic variations, the former having an amplitude of ~10–15 kPa at the seafloor, and ~5 kPa within the upper crust (Figure 3a). The effects of seafloor loading must be removed from the formation signal before the influence of Expedition 301 operations can be observed and interpreted. Second, pressure conditions within the isolated, upper basement interval in Hole 1027C tend to be offset (negatively) with respect to pressure conditions at the seafloor. This pressure difference results from the natural underpressure associated with local ridge flank hydrothermal circulation [Davis and Becker, 2002, 2004].

The Hole 1027C basement record was filtered and corrected using methods similar to those applied in earlier studies [Davis et al., 2000; Davis and Becker, 2004]. First, the instantaneous loading efficiency (\(\gamma\)) was evaluated by cross-plotting simultaneous responses at the seafloor and at depth during times when there was little or no long-term change in pressure, yielding a value of \(\gamma = 32.3\%\). This loading efficiency was used to correct the basement record, removing responses to short-term seafloor-pressure variations. Next, a low-pass filter was applied to the seafloor record, and 67.7% of the resulting long period signals were removed from the formation record (assuming fully drained conditions). Finally, the data were trend corrected to remove long-term instrument drift (~0.286 kPa/a). The resulting formation record contains some residual short-term variability, but clearly shows the influence of basement operations at Site 1301 during Expedition 301, with pressures rising abruptly several times during the expedition, then more slowly to ~1.5 kPa above baseline over the subsequent 13 months.
Figure 4. Detail of Hole 1027C pressure record and shipboard pumping before, during, and after IODP Expedition 301. Vertical bands indicate operations in Holes 1301A, 1301B, and 1026B. Abbreviations are as follows: D, drilling and casing operations; P, packer experiments; C, basement coring; O, observatory emplacement operations. (a) Relative pressure versus time. Arrows denote times when CORK observatories were released in the reentry cones in Holes 1301A, 1301B, and 1026B. The most abrupt changes in formation pressure measured in Hole 1027C are associated with basement drilling, coring, and packer operations in Hole 1301B. (b) Pumping rates on the drillship during basement operations, as recorded with rig instrumentation system. These are rates at which water was pumped down the drill pipe. During packer testing in Hole 1301B was this fluid forced to flow into basement. At all other times, some of this fluid may have returned to the seafloor rather than entering the formation. In comparing pumping rates to pressure records from Hole 1027C, there is little or no response to drilling or other operations in Hole 1301A. The largest abrupt pressure change in Hole 1027C follows packer experiments in Hole 1301B; the maximum pumping rate once the packer was set in the open hole (and fluid was forced into the formation) was 4–10 L/s.

[17] Basement fluid pressure variations in Hole 1027C during and after specific Expedition 301 basement operations show a clear response to short-term and longer-term perturbations at Site 1301 (Figure 4; Table 2). Surface seawater is pumped into basement boreholes during many operations, particularly when drilling and coring, but the amounts of drilling fluid that flow into basement or back up the hole to the seafloor (outside the drill pipe) are unknown. Basement pressures in Hole 1027C varied little (no more than ±0.1 kPa) during initial basement drilling and casing operations in Hole 1301A, but there was a clear response to basement drilling of the same interval in Hole 1301B, only 35 m away from Hole 1301A (Figure 4). Operations in both basement holes included drilling without coring through the upper 100–110 m of basement; pumping of surface seawater at 10–50 L/s; installation of casing; and an unsuccessful attempt to cement the casing strings to the borehole walls. There was virtually no pressure response in Hole 1027C to attempted packer experiments in Hole 1301A, which occurred immediately prior to CORK installation. This is consistent with the packer having been set near the base of unsealed casing, with little or no pressurization of the formation during testing. Similarly, there was little pressure response in Hole 1027C to setting the CORK in Hole 1301A (Figure 4a).

[18] In contrast, basement operations in the deeper Hole 1301B caused several increases in fluid pressure in Hole 1027C, although pumping rates were similar to those in Hole 1301A. Pressure continued to rise by 0.1–0.2 kPa/week during a break in Hole 1301B basement operations, prior to setting the CORK, and packer experiments in Hole 1301B resulted in the largest immediate pressure response in Hole 1027C, more than 0.2 kPa (Figure 4a). Pumping during packer tests comprised seven periods of fluid injection, each lasting one hour, with a maximum pumping rate of 10 L/s [Becker and Fisher, 2008]. Because the packer was inflated in open hole during these tests, all of this fluid was forced into the formation, creating pressure perturbations in Hole 1301B of up to 80 kPa (above beyond those from other causes), and resulting in relatively large, short-term pressure responses in Hole 1027C. The subsequent, long-term rate of pressure rise in Hole 1027C following packer testing and emplacement of the Hole 1301B CORK is less than that during the month prior to packer testing (Figures 3 and 4), probably because of resistance to flow associated with the incomplete CORK seal. There was no resolvable pressure response in Hole 1027C associated with emplacement of the subseafloor observatory in Hole 1026B, most likely because there was no fluid pumped into the formation during that operation (Figure 4).

[19] In summary, basement operations in Hole 1301B caused the greatest pressure response in Hole 1027C, whereas shallow basement operations in Hole 1301A and Hole 1026B had little or no influence. The packer experiments caused the greatest immediate perturbation, despite modest pumping rates, because pumped fluids were forced to enter basement, rather than being allowed to flow back up the borehole to the overlying ocean. These observations have two important implications. First, the upper 108 m of shallow basement surrounding Hole 1301A may be less well connected hydrologically to the upper 26 m of basaltic oceanic crust at the base of Hole 1027C than the deeper basement in Hole 1301B. This interpretation is important because it allows identification of the time at which the flow of cold bottom water into basement initiated the long-term pressure perturbation in Hole 1027C. Second, the rate of pressure rise following emplacement of the subseafloor observatory in Hole 1301B is less than the rate of pressure rise following packer experiments, and is also less than that associated with drilling, coring and other upper basement operations (Figure 4). This last observation provides a quantitative constraint on the long-term flow rate into basement at Site 1301, as discussed in the next section.

3.2. Rate of Fluid Flow Into Hole 1301B

[20] We need to know the long-term flow rate into Hole 1301B, Q, in order to interpret the pressure response in Hole 1027C. Q can be shown to fall within a relatively narrow range on the basis of (1) correlations between Expedition 301 basement operations at Site 1301 and pressure...
responses in Hole 1027C, (2) comparison to pressure responses during packer testing (when pumping rates are known), and (3) understanding of properties and processes that drove flow into the seafloor long after the end of drilling operations. The first two topics were addressed in the previous section, whereas the third topic is addressed in this section.

[21] Bottom seawater is driven into Hole 1301B by the difference between ambient formation pressure in the formation and cold hydrostatic fluid pressure in an open borehole. Ambient formation pressure in basement varies as a function of the local geothermal gradient, the temperature dependence of fluid density, and the nature of regional and local hydrothermal circulation. Basement drilling and other operations at Site 1301 included pumping of surface seawater, which lowered the temperature of the borehole relative to ambient conditions. Surface seawater is typically warmer than bottom water initially, but the drill pipe acts as an efficient heat exchanger as water is pumped to the seafloor, so the fluid temperature is close to that of bottom water when it exits the pipe and enters the hole (~1.8°C around Site 1301). The emplacement of a cold column of water in a borehole results in imposition of a hydrostatic pressure that is greater than that in basement [e.g., Becker et al., 1985; Morin et al., 1992], even if basement is naturally overpressured (as expected at Site 1301), leading to the long-term flow of cold bottom water into the formation.

[22] The primary requirements for this flow to continue within a hole drilled into basement that is initially overpressured are that (1) the magnitude of the pressure increase at the base of the borehole is greater than the natural formation overpressure, (2) formation permeability is sufficiently high, and (3) there is enough exposed basement at depth to allow rapid fluid flow down the borehole (cooling the borehole hydrostat). If one or more of these conditions is not met, then the flow of water into the formation may not be sustained. Not all basement holes become self-sustaining “hydrothermal siphons” in this way; flow down Hole 1026B spontaneously reversed 10–14 days after the end of drilling operations during Leg 168, when the natural formation overpressure was sufficient to overcome the cool borehole hydrostat [Fisher et al., 1997]. However, Hole 1026B had a short basement interval that was partly obstructed behind a metal liner, so it was relatively easy for the drilling-induced flow down that hole to be reversed by the natural overpressure.

[23] The differential pressure that draws bottom seawater down Hole 1301B and into the surrounding basement is calculated from the integrated difference between the densities of cold borehole water and ambient formation fluid (ρc and ρa, respectively): ΔP_\text{diff} = \int_0^L (\rho_c - \rho_a) g dz. Because some of this differential pressure must counteract a natural formation overpressure, calculations on the basis of this equation yield an upper limit as to the available driving force. Downward flowing bottom water cools the surrounding formation over time, and this can result in a positive feedback: sustained downward flow helps to lower borehole temperatures, resulting in a cooler borehole and a stronger driving force. But there is a limit to the extent of this feedback; once the borehole becomes isothermal at the temperature of bottom water, higher rates of flow down the hole do not generate a greater driving force.

[24] These considerations and the long-term temperature and pressure influence of flow down Hole 1301B are quantified analytically (Figures 5 and 6). Borehole temperatures are calculated using a transient heat-exchanger model developed for flowing gas wells, and the resulting rates of fluid flow into the formation are calculated using a transient radial-diffusion model [Becker et al., 1983; Fisher et al., 1997; Lesem et al., 1957; Matthews and Russell, 1967]. Borehole temperature calculations, and associated differential pressures, indicate that a downhole flow rate of 2 L/s is sufficient to generate nearly isothermal conditions in the upper 200 m of basement in Hole 1301B (Figure 5a), creating a differential pressure of 20–70 kPa depending on the depth-extent of the flow (Figures 5b and 6a). Given the locations of CORK packers set against the formation (~175 and 220 mbsf; Figure 2), flow probably extends to ~200 m into basement. There could be flow into the deeper crust once fluid leaves the borehole, but this fluid would lose much of its driving force as it penetrates to greater depths, on the basis of the near-hole formation properties calculated from packer tests [Becker and Fisher, 2008]. At the lower limit, a differential pressure of only 20 kPa will barely counteract the natural formation overpressure. At the upper limit, it is difficult to generate a driving force much greater than 70 kPa (Figure 6a). Thus calculations that follow are based on possible driving forces of 30–70 kPa.

[25] If there were a constant 30–70 kPa of driving pressure in Hole 1301B, the rate of fluid flow into the surrounding formation would decrease with time, counteracting (at least in part) the positive feedback associated with transient borehole cooling. The slowing of flow with time occurs as the excess fluid pressure in the borehole penetrates into the formation, reducing the lateral pressure gradient. Flow slows most in the first few months, but approaches an equilibrium value that is 40–50% of the initial value after one year (Figure 6b). One complexity in this last calculation is that the rate of flow into the formation depends on formation permeability, a quantity that is treated as an unknown in later aquifer analyses. However, we know something about formation properties adjacent to Hole 1301B from packer testing [Becker and Fisher, 2008], and assumptions used to estimate the long-term flow rate can be checked for consistency with subsequent cross-hole analyses.

[26] If the rate of flow is too low (≤0.05–0.1 L/s), borehole temperatures will differ only slightly from ambient formation conditions (Figure 5a), and the resulting differential pressure will be insufficient to sustain flow (Figures 5b and 6a). Borehole flows of this magnitude would occur if formation permeability were ≤10^{-13} m^2. At the other extreme, downhole flow for a year at a rate that is too high (≥10 L/s) would have resulted in a pressure response that is much larger than that observed (Figure 4), on the basis of comparison with the pressure response during and after packer testing. In any case, there is little justification for assuming a flow rate ≥10 L/s because the associated thermal anomaly generates little or no additional driving force beyond that generated by fluid flowing at ~2 L/s (Figure 6a). The upper end of the available long-term driving pressure (70 kPa) would result in a downhole flow rate that is too great if formation permeability were 10^{-11} m^2, so this permeability must be an upper limit.
On the basis of all of these constraints, and in consideration of the magnitude and timing of the cross-hole pressure response, the flow rate is most likely in the range of $Q = 2–5 \text{ L/s.}$

3.3. Calculated Hydrogeologic Properties Based on Cross-Hole Response

The simplest commonly used analytical model of transient fluid pressure response to flow to or from a nearby well, on the basis of conservation of mass, has the form (modified from Theis [1935]):

$$\Delta P = \frac{Q_{bg}}{4\pi T} W(u)$$

where

$$u = \frac{r^2 S}{4Tt}, \quad W(u) = \int_{u}^{\infty} \frac{e^{-u}du}{u}$$

$r =$ radial distance between the wells, $t =$ time, $T =$ formation transmissivity, and $S =$ formation storativity. Transmissivity is a parameter comprising formation and fluid properties within a tabular, horizontal aquifer ($T = k \rho bg / \mu$, where $k$ is aquifer permeability, $b$ is aquifer thickness, $\rho$ is density, and $\mu$ is viscosity). Storativity comprises the instantaneous formation fluid response to an incremental pressure perturbation ($S = \rho bg (\beta_m + n \beta_f)$), where $\beta_m$ is aquifer (matrix) compressibility, $n$ is aquifer porosity, and $\beta_f$ is fluid compressibility. Transmissivity and permeability are heterogeneous tensor quantities (particularly within fractured rocks), having values that vary with the temporal and spatial scales of testing. Formation storativity is a bulk (volumetric) scalar quantity, but it too can vary considerably with location, the spatial scale of testing, and the frequency of pressure perturbations.

The Theis equation is based on numerous assumptions and idealizations, including isotropic and homogeneous conditions; a constant rate of flow to/from the perturbation well; horizontal, radial flow in the formation to/from this well; and laminar flow conditions at the borehole wall and within the aquifer (i.e., Darcy’s law applies). The applicability of these assumptions in the present study are discussed later. The standard approach to applying the Theis equation to evaluate aquifer properties is to use known values of $Q$, $r$, and $t$, and initial values of $T$ and $S$ to calculate pressure responses. Values of $T$ and $S$ are shifted in an effort to minimize the misfit between observations and modeled pressures. We accomplished this

(Figure 6b). On the basis of all of these constraints, and in consideration of the magnitude and timing of the cross-hole pressure response, the flow rate is most likely in the range of $Q = 2–5 \text{ L/s.}$
task using nonlinear, least squares regression, assuming flow into the formation beginning with the start of Hole 1301B basement drilling, 7/12/04, 0130 hrs (Figures 4 and 7a).

29 The pressure response in Hole 1027C for 13 months following drilling operations at Site 1301 is matched well with the Theis equation (Figure 7a; Table 3). We convert values of transmissivity, $T$, to bulk formation permeability, $k$, using the definition of the former:

$$k = \frac{T \mu}{\rho gb}$$  \hspace{1cm} (2)

where $\rho = 1018$ kg/m$^3$ and $\mu = 4.4 \times 10^{-4}$ Pa-s (both appropriate for seawater at 65°C), and aquifer thickness, $b = 300$ m (Table 3; Figure 7b). The assumed value of $b$ is consistent with the geophysical logs (Figure 2), results of packer experiments [Becker and Fisher, 2008], flow rate calculations presented earlier, and observations from boreholes at other ridge-flank sites that suggest that the upper few hundred meters of crust are often the most permeable [e.g., Anderson et al., 1985b; Bartetzko, 2005; Fisher, 2004; Larson et al., 1993; Moos, 1990]. However, the most permeable part of the formation surrounding Hole 1301B could be considerably thinner in aggregate, comprising breccia, rubble, and pillows and flows within which fractures are most concentrated. Given a particular transmissivity value, the bulk permeability scales linearly with the combined thickness of the basement zone(s) through which most of the flow occurs. In fact, packer experiments suggest that the upper crustal interval with the greatest permeability may not be immediately adjacent to the sediment-basement interface (Figure 8).

30 We convert values of storativity, $S$, to aquifer matrix compressibility, $\beta_m$, using the primary definition for $S$:

$$\beta_m = \frac{S}{\rho gb} - n\beta_f$$  \hspace{1cm} (3)

where fluid compressibility, $\beta_f = 5 \times 10^{-10}$ Pa$^{-1}$ (Table 3; Figures 7c and 7d). Aquifer (matrix) compressibility comprises both grain and frame compressibility, but the former may be negligible with respect to the latter in fractured igneous rock. Porosity of the extrusive crust determined from borehole logs and other methods typically

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**Figure 6.** Downhole flow rates and differential pressures in Hole 1301B on the basis of observational and calculation constraints. (a) Differential pressures at a range of borehole depths and flow rates using data from Figure 5. The horizontal band at the base of the plot indicates the likely magnitude of the natural formation overpressure. This overpressure (~20 kPa) helps to explain why Hole 1026B recovered after drilling prior to emplacement of a long-term observatory [Fisher et al., 1997]: the excess fluid pressure created by the pumping of cold seawater into the hole was insufficient to overcome the natural formation overpressure. However, if borehole fluids penetrate to 100–200 m into basement, an excess borehole pressure approaching 70 kPa can be generated. A range of possible excess borehole pressures of 30–70 kPa is used for subsequent analyses. (b) Downhole flow rates as a function of time, on the basis of initial excess borehole pressures of 30–70 kPa (solid and dashed lines), and near-borehole formation permeability of $10^{-13}$, $10^{-12}$, and $10^{-11}$ m$^2$ (open circles, filled squares, and open triangles, respectively). Annotations to the right of the plot indicate additional observational constraints on the long-term flow rate into Hole 1301B, as discussed in the text. The range of long-term flow rates that is most consistent with an array of observations is 2–5 L/s.
ranges from 0.01 to 0.2 [Anderson et al., 1985a; Bartetzko et al., 2001; Bartetzko, 2005; Bartetzko and Fisher, 2008; Becker et al., 1982; Moos and Marion, 1994], with the highest values found within relatively thin zones surrounded by more massive rock. When averaged over an upper crustal interval of 300 m, it is likely that the effective porosity is at the lower end of this range; greater porosity requires either a higher long-term flow rate into the formation, or aquifer (matrix) compressibility that is much less than that of seawater (Figure 7d), both of which seem unlikely. A value of \( n = 0.02 \) has been adopted for this analysis.

Given a long-term flow rate into basement at Site 1301 of \( Q = 2–5 \) L/s, the apparent basement transmissivity

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**Figure 7.** Comparison between observations in Hole 1027C and results of analytical calculations for response to flow into Hole 1301B, as well as implications for formation hydrogeologic and mechanical properties. (a) Comparison between filtered pressure observations (a segment of data from Figure 3b) and analytical calculations using equation (1). Values associated with an inferred long-term flow rate of 2–5 L/s are shown; this range of flow rates is indicated in subsequent panels with a vertical, dotted band. The quality of the fit is insensitive to selected flow rate; values of transmissivity and storativity can be adjusted to allow a fit to essentially any flow rate. (b) Formation transmissivity and permeability as a function of long-term flow rate on the basis of the assumption that properties apply to the upper 300 m of basement. (c) Formation storativity as a function of long-term flow rate. (d) Formation compressibility on the basis of inferred values of storativity and the functional relation between formation compressibility and storativity (the latter comprising formation and fluid compressibilities). There is a dependence of formation compressibility in the assumed formation porosity, but this dependence is relatively unimportant at the inferred range of flow rates. The arrow indicates the compressibility of seawater, which is a rational lower limit to the compressibility of the upper basement aquifer. Other studies suggest that the compressibility of the upper basement aquifer should be closer to \( 10^{-9} \) Pa⁻¹ [e.g., Davis and Becker, 2007], a value consistent with the range inferred in the present study.
### 3. Discussion

#### 3.1. Implications of the Observed Cross-Hole Response for Crustal Anisotropy

[33] The range of bulk permeability values inferred from the cross-hole response between Sites 1301 and 1027 (0.7 to $2 \times 10^{-12}$ m$^2$) is at the lower end of values determined through single-hole packer testing and evaluation of tem-

| Table 3. Calculated Basement Properties Based on Cross-Hole Response in Hole 1027C to Long-Term Flow Down Hole 1301B$^a$ |
|---|---|---|---|---|---|
| Q, L/s | $T^h$ m$^2$s | $S^h$ | $k^h$ m$^2$ | $\beta_{m,d}$ Pa$^{-1}$ | $\beta_{m}/\beta_f$ |
| 0.2 | $4.8 \times 10^{-4}$ | $1.1 \times 10^{-4}$ | $6.8 \times 10^{-14}$ | $2.5 \times 10^{-11}$ | 0.05 |
| 0.5 | $1.2 \times 10^{-3}$ | $2.7 \times 10^{-4}$ | $1.7 \times 10^{-13}$ | $7.7 \times 10^{-11}$ | 0.2 |
| 1 | $2.4 \times 10^{-3}$ | $5.3 \times 10^{-4}$ | $3.4 \times 10^{-13}$ | $1.6 \times 10^{-10}$ | 0.3 |
| 2 | $4.8 \times 10^{-3}$ | $1.1 \times 10^{-3}$ | $6.8 \times 10^{-13}$ | $3.4 \times 10^{-10}$ | 0.7 |
| 5 | $1.2 \times 10^{-2}$ | $2.7 \times 10^{-3}$ | $1.7 \times 10^{-12}$ | $8.6 \times 10^{-10}$ | 1.7 |
| 10 | $2.4 \times 10^{-2}$ | $5.3 \times 10^{-3}$ | $3.4 \times 10^{-12}$ | $1.7 \times 10^{-9}$ | 3.5 |
| 20 | $4.8 \times 10^{-2}$ | $1.1 \times 10^{-2}$ | $6.8 \times 10^{-12}$ | $3.5 \times 10^{-9}$ | 6.9 |

$^a$Range of preferred interpretations given in boldface.

Values of transmissivity ($T$) and storativity ($S$) consistent with fit of Theis’s [1935] equation to observations in Hole 1027C, given the radial distance between perturbation and observation boreholes of 2.4 km, and the long-term flow rates ($Q$) into Hole 1027B indicated in the first column.

Values of bulk permeability calculated using equation (2) and parameter values discussed in the text.

Aquifer (matrix) compressibility calculated as $\beta_m = S/(Qgh) - n/\beta_f$ where porosity ($n$) is 0.02, and fluid compressibility ($\beta_f$) is $5 \times 10^{-10}$ Pa$^{-1}$. $\beta_{m,d}/\beta_f$ indicates the ratio of aquifer (matrix) compressibility to seawater compressibility.

is 0.5 to 1.2 $\times 10^{-2}$ m$^2$/s, suggesting a bulk permeability within the upper 300 m of crust of 0.7 to $2 \times 10^{-12}$ m$^2$ (Figure 8). Permeability would be commensurately greater if it were focused within one or more thin zones in the upper crust, as inferred from packer experiments in Hole 1301B [Becker and Fisher, 2008]. On the basis of the cross-hole response between Holes 1301B and 1027C, the storativity of this region of seafloor is 1 to $3 \times 10^{-3}$ (Table 3; Figure 7b). This storativity implies aquifer compressibility of $3 \times 9 \times 10^{-10}$ Pa$^{-1}$, a value close to or somewhat larger than that of seawater (Table 3; Figure 7d), and consistent with values considered typical of the uppermost oceanic basement rocks on young ridge flanks [Davis and Becker, 2007]. Qualitatively, a relatively “stiff” crustal aquifer is required so that relatively small pressure perturbations can travel long distances without being severely attenuated. On the other hand, it seems unlikely that fractured and rubbly upper oceanic crust can be much stiffer than seawater.

[32] Although the distance between Holes 1301B and 1027C is 2.4 km, the region of seafloor tested during the 13 months following Expedition 301 extends to a greater radial distance. This distance ($r_0$) can be estimated using a “distance-drawdown” formula derived from an approximation to the Theis solution [Cooper and Jacob, 1946]:

$$r_0 = \sqrt{\frac{2.25T}{S}}$$

(4)

On the basis of the ranges of $T$ and $S$ values listed above, and assuming that flow into the formation around Hole 1301B is distributed radially, $r_0 = 10$–30 km. This is a much larger crustal scale than that tested during single-hole packer experiments. If flow into the formation were not distributed radially, e.g., if basement were azimuthally anisotropic, or if most flow occurred through a small number of fractures, then the effective distance of influence could be considerably greater in the preferred flow directions. This range of spatial scales is comparable to those applied in numerical modeling studies, using pressure and temperature constraints, and to those associated with analyses of tidal and crustal-strain pressure responses in sealed boreholes in the oceanic crust.
temperature logs in similarly aged seafloor from this region (Figure 8). However, numerical modeling and calculations on the basis of tidal responses and drainage following tectonic strain events have suggested that upper basement permeabilities in the region around Sites 1026, 1027, and 1301 may be as great as $10^{-10}$ to $10^{-9}$ m$^2$, two to three orders of magnitude higher than inferred in the present study [e.g., Davis et al., 2000; Davis and Becker, 2002; Spinelli and Fisher, 2004]. Why is crustal-scale bulk permeability inferred through cross-hole analyses at the low end of estimates from single-hole packer and thermal experiments that test much smaller rock volumes, and much smaller than other methods that evaluate commensurate rock volumes? We propose that basement in this region may be azimuthally anisotropic.

[34] Anisotropy in the seismic properties of young oceanic basement has been recognized for decades, and has been interpreted to result from preferential orientation of cracks, faults, and fractures (the crustal “fabric”) [e.g., Almendros et al., 2000; Detrick et al., 1998; McDonald et al., 1994; Shearer and Orcutt, 1985; Sohn et al., 1997; Stephen, 1981]. The orientation of the dominant crustal fabric on ridge flanks is generally thought to be subparallel to abyssal hill topography and to the orientation of the spreading center, and should tend to favor fluid flow in the upper crust in the “along-strike” direction [e.g., Delaney et al., 1992; Fisher, 2004; Haymon et al., 1991; Macdonald, 1982, 1998; Macdonald et al., 1996; Wilcock and Fisher, 2004]. We do not suggest that the same cracks may be responsible for both seismic and hydrogeologic anisotropy, although both kinds of anisotropy may result from the tectonic fabric of the crust. This interpretation is consistent with thermal and chemical observations from this region, which suggest that there may be a preferential fluid flow direction within basement [Fisher et al., 2003; Hutnak et al., 2006; Wheat et al., 2000].

[35] Quantifying azimuthal anisotropy in permeability with a cross-hole hydrogeologic experiment requires monitoring two or more boreholes arrayed in different directions from one or more perturbation wells [e.g., Hantush and Thomas, 1966; Heilweil and Hsieh, 2006; Hsieh, 1985a, 1985b; Neuman, 1975; Neuman et al., 1984]. Monitoring with a three-dimensional borehole network will occur during the full experiment around Site 1301, but we can use the available data to calculate possible azimuthal responses. Site 1027 is located along an azimuthal direction of N70°E from Site 1301, roughly halfway between the direction in which transmissivity may be greatest (N20°E, subparallel to the trend of abyssal hills and the active spreading center to the west) and the direction in which is may be the lowest (N110°E, perpendicular to the trend of abyssal hills and the spreading center) (Figure 1). The magnitude of apparent transmissivity, $T_{app}$, within an azimuthally anisotropic aquifer can be calculated as [Papadopulos, 1965; Dawson and Istok, 1991]:

$$\frac{1}{T_{app}} = \frac{1}{T_{max}} \cos^2 \varphi + \frac{1}{T_{min}} \sin^2 \varphi$$  \hspace{1cm} (5)

where $\varphi$ is the angle between the observation direction and the direction of maximum transmissivity. Calculations on the basis of this formulation illustrate that the apparent transmissivity in an anisotropic aquifer determined from the Site 1301 to Site 1027 response should be much closer to the actual value in the low-transmissivity direction (Figure 9a).

[36] The pressure perturbation in Hole 1026B resulting from fluid flow into basement at Site 1301 is unknown because Hole 1026B was still recovering from long-term borehole discharge during the period of interest, but the expected pressure response in Hole 1026B can be calculated on the basis of properties and flow rates inferred in the present study. If the crust were isotropic, the pressure rise in Hole 1026B would have been about 2.0 kPa, somewhat higher than observed in Hole 1027C (because the former is closer to Site 1301). Although higher transmissivity in the orientation between Sites 1301 and 1026 would result in greater fluid flow and a more rapid pressure response in this direction, the measured pressure rise in Hole 1026B would be lower than in the case with no anisotropy, as indicated.

Figure 9. Calculations of effective formation transmissivity and anticipated pressure response to the cross-hole experiment analyzed in the present study, assuming different values for azimuthal hydrogeologic anisotropy. (a) Effective transmissivity ratio (apparent transmissivity/highest transmissivity) as a function of the angle of measurement. Vertical band indicates orientation of the Site 1301 to Site 1027 experiment, assuming that the direction of highest transmissivity is 20° (subparallel to the crustal fabric) and that the direction of lowest transmissivity is perpendicular to this, 110°. The orientation of Site 1027 from Site 1301 is 50°. (b) The calculated formation pressure response in Hole 1026B as a result of long-term flow into the crust at Site 1301 at rates inferred in the present study. If there were no crustal anisotropy, the pressure response in Hole 1026B would have been about 2.0 kPa, somewhat higher than observed in Hole 1027C (because the former is closer to Site 1301). Although higher transmissivity in the orientation between Sites 1301 and 1026 would result in greater fluid flow and a more rapid pressure response in this direction, the measured pressure rise in Hole 1026B would be lower than in the case with no anisotropy, as indicated.
sivity were anisotropic, there would be much more fluid flow within the crust in the high-transmissivity direction, but the pressure rise in this direction would be smaller, relative to the isotropic case. The difference between pressure response and fluid flow rate should be apparent in the complete cross-hole experiment, which will include pumping of tracers followed by long-term sampling and monitoring in distant boreholes [Fisher et al., 2005b].

[37] The occurrence of anisotropy in crustal structure and hydrogeologic properties in this region would require a reassessment of earlier estimates on the basis of models and responses to tidal and seismic events [e.g., Davis et al., 2000, 2001, 2004; Fisher et al., 2003; Hutnak et al., 2006; Spinelli and Fisher, 2004]. For example, earlier numerical studies of this area have assumed radial or two-dimensional flow geometries. In some cases the properties inferred from these studies would apply to the preferential flow direction, but in others the models were crafted to simulate or interpret cross-strike flow. Tidal and seismic responses have been modeled using one- or two-dimensional (planar) geometries, with the diffusion of pressure perturbations in the cross-strike direction. The possibility that crustal hydrogeologic properties are azimuthally anisotropic remains a hypothesis that will be tested through future multidirectional, cross-hole experiments.

4.2. Limitations and Needs for Controlled Testing

[38] Although there is a good fit between an analytical model of aquifer response and pressure observations in Hole 1027C (Figure 7a), there are important limitations to these analyses. First, we have assumed that all flow occurs in Hole 1301B, based mainly on the limited pressure response to operations in Hole 1301A relative to those in Hole 1301B (Figure 4). It is possible that cooling of the crust around Hole 1301B led to water being drawn into Hole 1301A as well. However, Hole 1027C is far enough from Site 1301, and Holes 1301A and 1301B are sufficiently close, that the pressure response of crustal fluid would be insensitive to whether flow occurs in Hole 1301B, 1301A, or a combination of the two. All of the constraints on the total flow rate into the crust at Site 1301 apply even if there were leakage into (or out of) Hole 1301A.

[39] The model used to fit the Hole 1027C pressure data is highly idealized. Other models were used to fit the data, including those for a leaky aquifer (to allow flow to/from the overlying sediment layer or to/from nearby basement outcrops) and those for fractured, dual-porosity, dual-permeability systems [e.g., Hantush, 1960; Moench, 1984], but because pressure observations are fit so well by a model that neglects these and other complexities, more complex models yield properties essentially identical to those suggested by the simpler model. We could not apply more sophisticated models of crustal anisotropy [e.g., Hsieh, 1985b] because we lack multiple observation wells at different orientations and distances from the perturbation well. On the other hand, the assumptions upon which the analytical Theis solution is based are consistent with many of the assumptions used in packer experiments, analyses of thermal logs, numerical models of coupled flows, and analyses of tidal and seismic responses. Thus these interpretations of cross-hole response are particularly useful for comparison with earlier interpretations.

[40] Hydrogeologic observations in this area are likely to be influenced by crustal heterogeneity. The lack of pressure response in Hole 1027C to pumping in Hole 1301A suggests that there are significant variations in crustal properties across distances of tens of meters. Distinguishing between anisotropy, heterogeneity, and other nonidealities in experimental configuration will require longer-term data from multiple boreholes. The greatest limitations of the present study are the lack of time series measurements of the flow rate into basin at Site 1301 and the lack of monitoring points at multiple distances, depths, and directions from the perturbation well. Each of these limitations will be corrected as part of the full experiment, which is essential for applying more sophisticated interpretative methods. The leaking observatories will be sealed, and new observatories will be installed between Sites 1301 and 1026. One of these new holes will be allowed to discharge for 1–2 a, using the natural formation overpressure to drive flow, and an autonomous flowmeter will record a time series of flow rates as the pressure responses in five surrounding boreholes are monitored (borehole spacing of 40–2300 m).

5. Conclusions

[41] Long term, subseafloor observatories (CORKs) were installed in 3.5–3.6 Ma seafloor on the eastern flank of the Juan de Fuca ridge to evaluate formation hydrogeologic properties and processes in the upper oceanic crust. CORKs installed at Site 1301 in summer 2004 were incompletely sealed, allowing cold bottom water to flow into the oceanic crust, causing an observable pressure perturbation at a preexisting CORK in Hole 1027C, 2.4 km to the east.

[42] Consideration of pumping rates during drilling, packer experiments, and other basement operations prior to CORK installation; the timing and rate of pressure perturbations; and the thermal and pressure conditions in basement boreholes into which bottom water is drawn suggests that the flow rate into basin at Site 1301 during 13 months following drilling averaged 2–5 L/s. The pressure perturbation observed at Site 1027 is matched well with the solution to a radial conservation of mass equation, on the basis of idealized aquifer geometry and properties, using conventional analytical methods. Values of upper crustal transmissivity indicated by the fit of observations to the analytical solution are $T = 0.5$ to $1.2 \times 10^{-2} \text{m}^2/\text{s}$, suggesting bulk permeability in the upper 300 of basaltic crust of $k = 0.7$ to $2 \times 10^{-12} \text{m}^2$. The fit to the analytical model suggests storativity (a combination of fluid and aquifer compressibilities) of $S = 1$ to $3 \times 10^{-2}$, and a crustal aquifer (matrix) compressibility of $3 \times 10^{-10} \text{Pa}^{-1}$.

[43] The radius of influence of this inadvertent cross-hole experiment is much larger than that associated with most earlier, single-hole borehole experiments, ≥10–30 km, a scale similar to that examined with numerical models and analyses of formation responses to tides and tectonic events. The inferred basement permeability is consistent with, but at the low end, of permeabilities inferred from single-hole packer experiments at this and other young oceanic crustal sites, and two to three orders of magnitude less than values inferred from numerical models or analyses of tidal or tectonic event responses. The relatively low transmissivity and permeability values inferred from the cross-hole
response may be explained if the basement aquifer in this area is azimuthally anisotropic. The direction of preferred flow may be oriented at an azimuth of ~N20°E, subparallel to abyssal hill topography and the active spreading center to the west, as inferred from earlier thermal, chemical, and numerical studies. It is not possible to quantify azimuthal anisotropy in basement flow properties with a single observation borehole, but resolution should be possible when the complete basement observatory network is installed, and a controlled long-term flow experiment is initiated, beginning in 2010.

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