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Thermal Characteristics of the FORGE site, Milford, Utah

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ABSTRACT

The FORGE Milford site is situated over the flank of a granitic intrusion that outcrops to the east in the Mineral Mountains and has a temperature of more than 200°C at 2.4 km depth directly beneath the site. Integration of the thermal data from over 100 thermal gradient wells and deep exploration wells suggests that about 100 km^2 of the intrusion has a temperature of at least 200°C at 4 km depth. The temperature profile from the 2.3-km-depth well 58-32 that was drilled as part of Phase 2B of the FORGE project confirms a conductive thermal regime within the granite, with the gradient at depth being 70°C/km. Measurements of cuttings and core from the well show a wide range of thermal conductivity depending on the composition of the granitic rock. Values vary from 2.1 W/m°C in quartz-poor dioritic rock to 3.9 W/m°C in more quartz-rich granite. These measurements were made at room temperature and need to be reduced by about 15% for the effects of temperature at the 2–3 km future reservoir beneath the site. Using representative values for the thermal conductivity and temperature gradient from the lower part of 58-32 yields a heat flow of 180 ± 20 mW/m^2. This value appears to be representative of the deep, thermally conductive, heat flow in impermeable granite surrounding the Roosevelt hydrothermal system. Laboratory measurements of the specific heat of selected granitic core from 58-32 are 0.7–1.0 kJ/kg°C, and the calculated thermal diffusivity ranges 1.0–1.7 mm^2/s. For correction to reservoir temperatures of 200°C, the heat capacity will be higher by about 20% and the thermal diffusivity will be lower by about 20%. These parameters will be required for simulation of heat sweep scenarios in fractures beneath the site. If the FORGE project is successful in developing the technology to create a fracture network in the impermeable rock beneath the FORGE site, the power potential of this region will be immense.

1. Introduction

A review of the thermal characteristics of the Roosevelt Hot Springs system (RHS) was carried out in Phase 1 of this project, with a summary presented at the 2015 Stanford Geothermal

Workshop (Allis et al., 2015). That review expanded the original assessment of the heat flow at RHS by Ward et al. (1978) by adding other thermal data from exploration wells drilled after 1978. The initial state and response of the Blundell borefield to production and injection was analyzed by Allis and Larsen (2012), and an update of the fluid chemistry changes was presented at the 2018 Stanford Geothermal Workshop (Simmons et al., 2018a). This paper integrates the thermal results from the deep well drilled for the FORGE project (Utah FORGE, 2018; **F**rontier **O**bservatory for **R**esearch into **G**eothermal **E**nergy), and the measurement of thermal rock properties from the drillhole.

2. Groundwater Temperatures

The near-surface thermal characteristics of the RHS are delineated by the temperatures at 200 m depth obtained from over 50 thermal gradient wells drilled mostly in the late 1970s. The highest temperatures (> 80°C) occur over the top of the hydrothermal system on the east side of the Opal Mound fault, and this area is tapped by the Blundell production wells (Figure 1). Temperatures at greater depth here are close to boiling-point for depth. The northwest extension of the high-temperature zone at 200 m depth is due to an outflow of hot groundwater. Originally, the main hot spring flowed here (RHS in Figure 1), and the area was developed into a sanitarium around the year 1900. By the mid-1960s the spring flow was ephemeral, and the facilities had fallen into disrepair. The 80°C isotherm at 200 m depth is approximately equivalent to a heat flow of 1000 mW/m^2. When integrated across the entire RHS system, the total heat output is 60–70 MW. Assuming a temperature of 270°C (enthalpy of 1180 kJ/kg) in the deep upflow zone of the hydrothermal system, the pre-development mass flow of the system was 60 kg/s.

At 200 m depth, for a typical conductive Great Basin heat flow of 90 mW/m^2 and an average surface temperature of 12°C, the temperature should be about 24°C and the gradient in the overlying unconsolidated sediment should be about 60°C/km (assuming a thermal conductivity of 1.5 W/m°C). In low porosity crystalline rocks (thermal conductivity of about 3 W/m°C), the equivalent temperature is about 18°C and the gradient is about 30°C/km). In the Mineral Mountains where the crystalline rocks occur, the elevation rises over a kilometer above the valley floor, so the mean annual surface temperature could be 3–5°C cooler. The temperature at 200 m depth near the crest of the Mineral Mountains, assuming thermal conduction from the surface, should be decreased by 3–5°C, implying a background temperature at 200 m depth of about 13–15°C. Temperatures at 200 m depth greater than about 30°C in the valley and greater than about 20°C in the Mineral Mountains, are therefore anomalously warm. Examples of thermal gradients are shown in Figure 2.

When the thermal gradient data for the central, southern and western sectors are plotted against depth, the near-surface profiles (less than about 80 m depth) are surprisingly similar. The gradient of 270°C/km is equivalent to a heat flow of 350–400 mW/m^2 (average thermal conductivity of 1.3–1.5 W/m°C). The thermal gradients decrease by almost a factor of four at greater depth and can't be explained by a corresponding increase in thermal conductivity. The two wells in the central Beaver River valley (MW22, 23) and the McCullough well near Acord-1 exhibit a uniform thermal gradient from the surface consistent with the zero-depth intercept being the mean annual temperature. There is no evidence of cross-flowing warm water in these wells. When the all the temperature profiles are plotted against elevation, the profiles are spread from warmer in the east (GPC-15) to cooler in the west (for example, TPC-3, near intersection of

Geothermal Plant Rd., and Antelope Pt. Rd.) The water levels also decrease in elevation from east to west, confirming an outflow of warm groundwater (50 m of head change over 10 km). The temperature of the outflow at the water table decreases from 42°C at GPC-15, to 25°C at TPC-3. This cooling of the groundwater appears to be due to both conductive heat flow through the vadose zone and mixing with infiltrating meteoric water (evidence of dilution towards the west; Simmons et al., 2018).

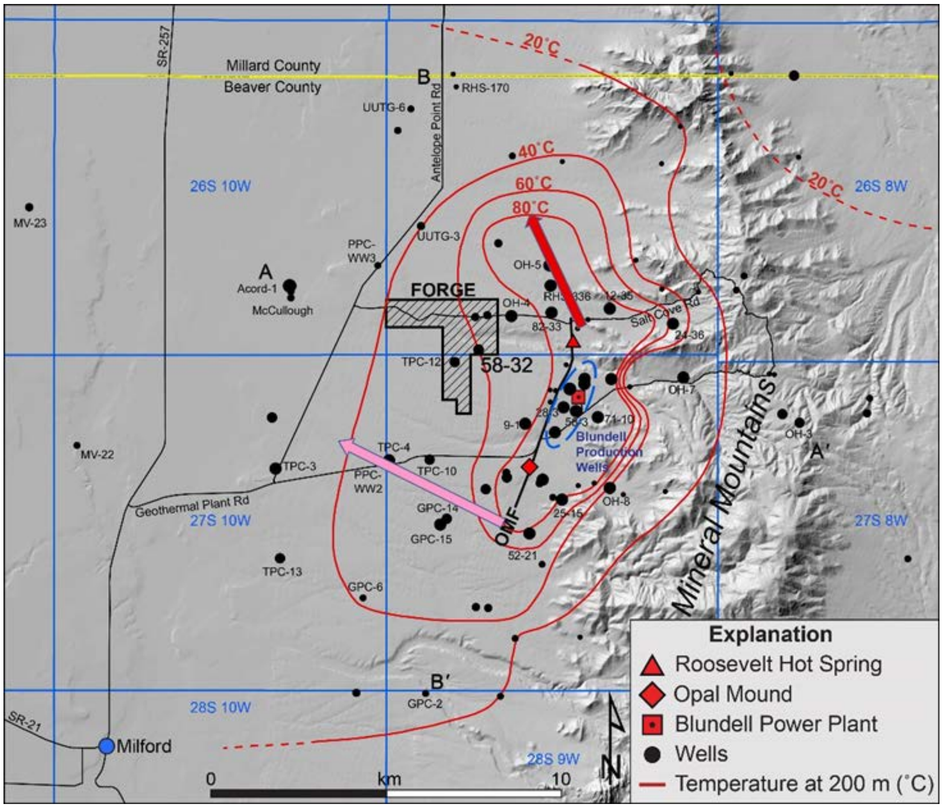

Figure 1: Thermal regime at 200 m depth. The degree of certainty of the thermal data is indicated by the size of the dot for the well location (all wells deeper than 50 m). The largest size is for wells greater than 200 m depth where the temperature was observed. The smallest size is for wells about 50 m deep where the temperature had to be extrapolated to a much greater depth and the temperature is considered the least certain. On the east side of the thermal anomaly, the contours represent the temperature at 200 m below the 1830 m above sea level (6000 ft asl) datum, which is the elevation of the alluvial fan near to where it laps against the Mineral Mountains. This allows the contours to be smoothed across the ridges and valleys, but requires that higher-elevation wells be extrapolated to greater depths (up to 405 m from the surface). Farther to the west, the contours are at 200 m depth from the surface, and near SR-257 in the middle of the valley this is at about 1325 m asl (4345 ft asl; ground surface about 5000 ft asl or 1525 m asl). OMF is the Opal Mound fault. A-A′ and B-B′ are the lines of cross section used in a later figure. Red and pink arrows indicate relatively young (red, ~ 1000 y) and long-lived (pink, ~ 10,000 y) outflows of geothermal groundwater.

In the northern sector of RHS, where the NW-directed plume flows from the northern end of the Opal Mound fault, there is evidence of hot water outflow between 1300 and 1500 m asl. Well OH-5 best captures the character of the outflow plume (Figure 3). The new FORGE well 58-32 has an isothermal section between 1300 and 1500 m asl and is suspected of being on the edge of this plume. The temperature of the plume near its outflow source probably exceeds 130°C, based on the temperature in 82-33.

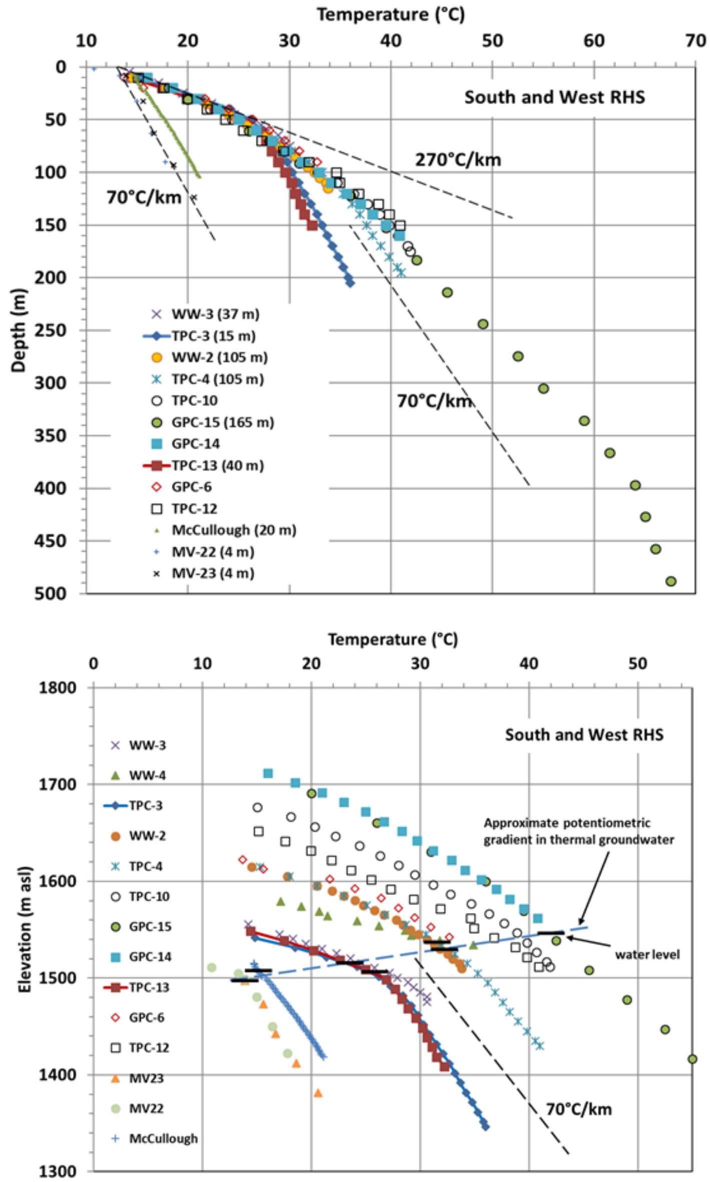

Figure 2. Upper: Temperature profiles from thermal gradient wells around the southern and western areas of the RHS system, plotted against depth below the ground surface. Although most wells over the RHS system have shallow temperature gradients exceeding 200°C/km, the deeper thermal gradients are 70°C/km (water level in brackets). Lower: When plotted against elevation (meters above sea level), there is a pattern of warm groundwater flowing from east to west, with the high thermal gradients at shallow depth in the vadose zone, and apparently conductive thermal profiles below the water table.

The pattern of a warm outflow in south central RHS having an underlying uniform temperature gradient, contrasts with the inversion present beneath the hot outflow in northern RHS. This is a phenomenon that was studied by Ziagos and Blackwell (1981, 1986). They showed that the thermal effect of a horizontal outflow of warm groundwater initially causes a temperature inversion beneath the outflow, but after a long time of continued warm outflow the underlying thermal regime heats up to reflect the regional equilibrium thermal gradient (Figure 4). The time frame for the inversion to disappear is tens of thousands of years. This may help explain the fossil siliceous sinter at Opal Mound near the southern end of the OMF and the active hot spring area at the northern end of the OMF. Although the sinter at Opal Mound exhibits a transition from Opal-A to diagenetic quartz, and in New Zealand this transition takes about 40,000 y, the two dates from the Opal Mound are between 1000 and 2000 y (Lynne et al., 2005). Perhaps the Opal Mound is significantly older than the two dates imply. In contrast, the hot groundwater plume northwest of the RHS maybe be relatively young (~ 1000 years). Additional thermal modeling is being undertaken to test these ideas.

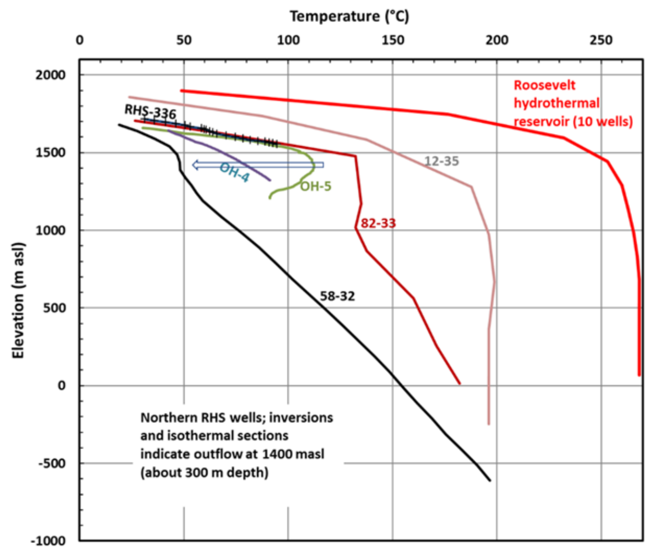

Figure 3. Temperature profiles in wells around the northwest of RHS system where the main outflow zone of hot water occurs. The outflow is characterized by temperature inversions (OH-5) and isothermal zones between 1500 and 1700 m asl (typically 200–400 m depth; blue arrow). 58-32 was drilled during Phase 2B of the FORGE project.

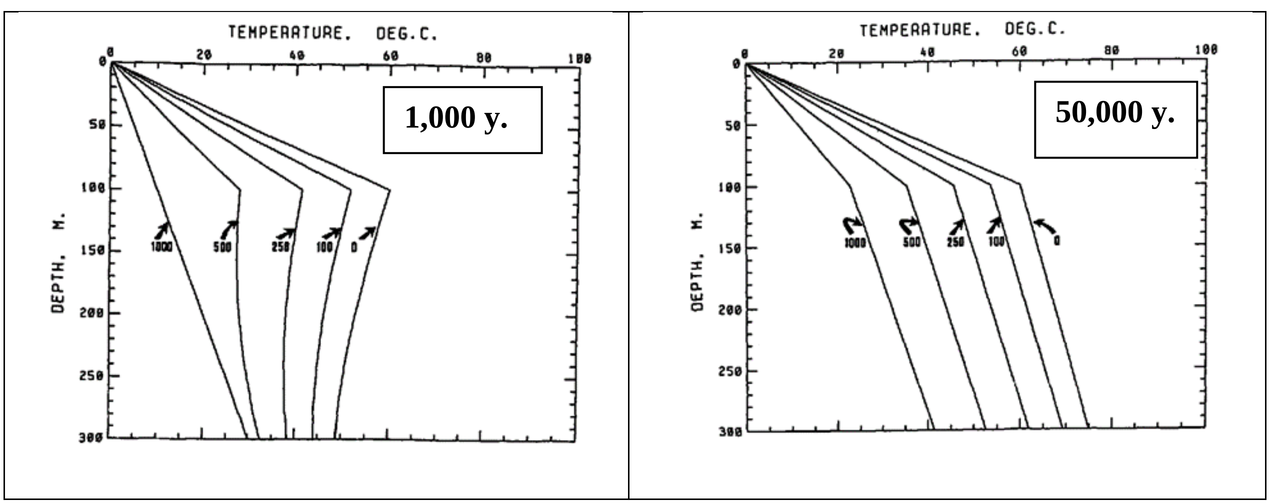

Figure 4. Theoretical temperature-depth profiles for the thermal effects due to lateral flow in an aquifer at 100 m depth at 60°C at the recharge point (0 m) and at distances up to 1000 m from the recharge point (water flow rate of 1 m/year assumed; 100°C/km ambient gradient). On a time scale of thousands of years, there is a temperature inversion beneath the aquifer; on a time scale of tens of thousands of years, the thermal regime at depth is the equilibrium gradient (Ziagos and Blackwell, 1981, 1986).

3. Deep Wells

Equilibrium profiles for the deeper exploration wells are shown in Figure 5 (locations on Figure 1). Wells tapping the hydrothermal reservoir are distinctive because of their high near-surface temperature gradient (typically close to a boiling point-for-depth profile) and near-isothermal zone below about 500 m depth in the upflow zone. Most wells outside the upflow zone are much cooler near-surface, but have temperature gradients below about 1 km depth which can be extrapolated to greater depth. This is indicative of poor permeability, lack of fluid flow, and thermal conduction as the dominant form of heat transfer. The new well drilled in Phase 2B, 58-32, is in this category, with a deep gradient of 70°C/km (discussed in next section). Most of these wells point to temperatures of more than 250°C at depths of 3–4 km. These deeper wells provide proxies for extrapolating nearby thermal gradient wells and enable isotherm maps to be compiled at different depths.

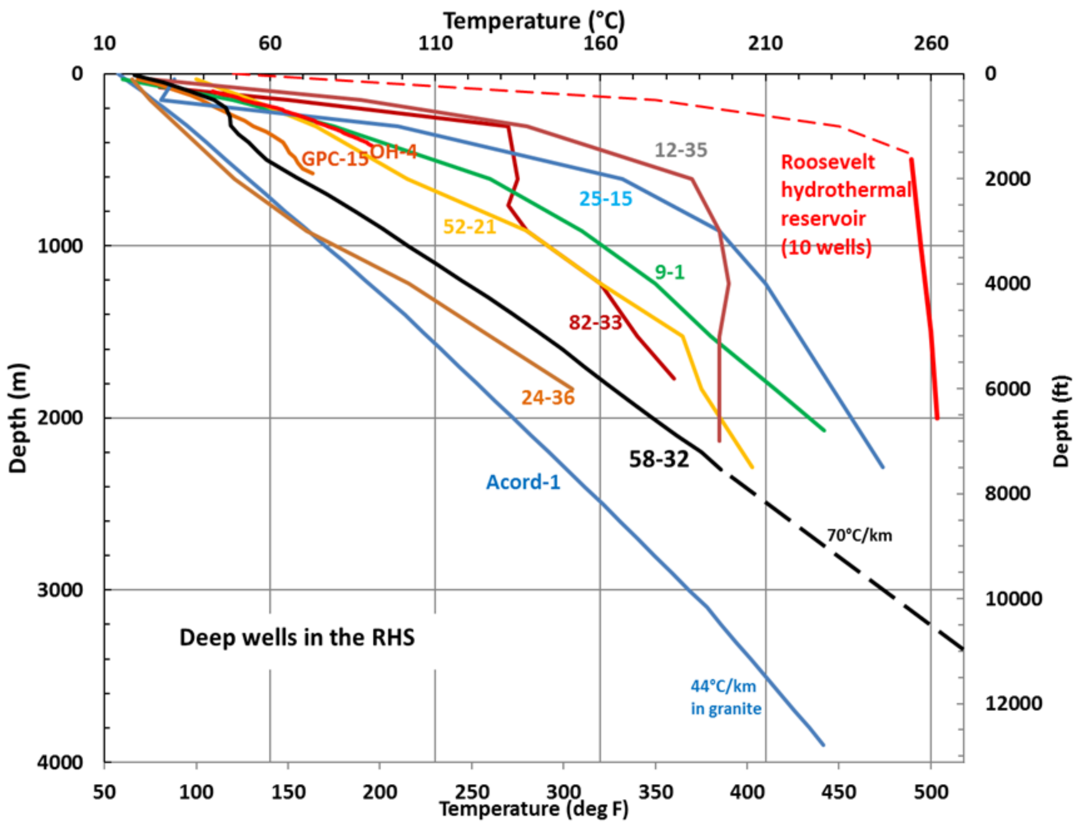

Figure 5. Thermal profile in 58-32 compared to other deep wells outside the hydrothermal reservoir that is tapped by the Blundell borefield. Within the production borefield, reservoir temperatures prior to development were near boiling point-for-depth, and typically in the range of 255–265°C at greater depth. Except for well 12-35, which is at the northern end of the reservoir and close to the main outflow zone, and Acord-1, which is near the western edge of the RHS system, the profiles seem converge on about 250–270°C at 3–4 km depth. Although the quartz-silica equilibrium temperatures closely match reservoir conditions and its evolution (cooling) with production, the Na-K equilibration temperatures of 270–310°C indicate hotter conditions representative of deeper parts of the hydrothermal system, below reservoir depths (Simmons et al., 2018).

4. FORGE well 58-32

FORGE well 58-32 was completed in October 2017, and a precision temperature profile was run in the well just over a month later (Figure 6). Details of the thermal measurements made during drilling and attempts to predic the equilibrium temperature during a 24-hour stoppage in drilling when the hole was at 2040 m depth are discussed by Allis et al. (2018). The well took almost two months to drill, and although the temperature at bottom of the hole may be within a degree of equilibrium, the observed thermal regime between about 300 and 1500 m depth is suspected to be several degrees warmer than equilibrium conditions. Temperature gradients derived from the 37-day profile in this depth range could be significantly impacted. The gradient averages 70°C/km between 1500 and the bottom of the hole at 2300 m, and should be a reliable indicator of temperatures at greater depth (subject to uniform thermal conductivity.) The higher gradient between 1500 and 500 m may be an artifact of non-equilibrium conditions and should be ignored. The near-isothermal section between 200 and 300 m coincides with the top of the groundwater table based on surrounding wells and is interpreted to indicate a lateral flow of water at 45°C.

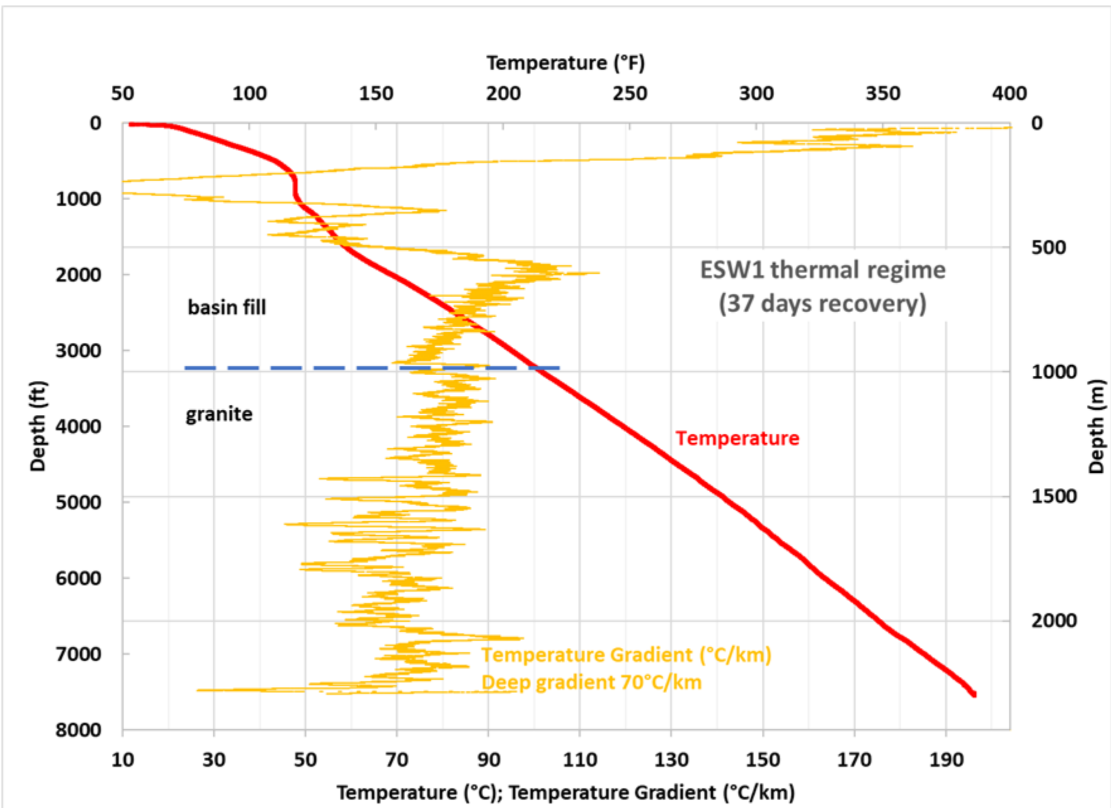

Figure 6. Thermal profile in 58-32 37 days after drilling and testing stopped. The temperature gradient near the bottom of the well of 70°C/km is probably close to the deep, equilibrium conductive gradient at this site. The increased gradient between 1500 and 500 m depth is suspected to be an artifact of non-thermal equilibrium. The near isothermal section between 200–300 m depth indicates a groundwater aquifer.

There is a 20 percent change in gradient centered at 975 m depth coinciding with the transition from basin fill to bedrock. This is caused by an equivalent decrease in thermal conductivity from 2.6 ± 0.2 W/m°C in the granitic fill to 2.2 ± 0.2 W/m°C in monzodiorite below (lithologies from

XRD data of Jones, 2018). Although the absolute gradients may change with equilibrium conditions, the relative change in gradient should be unaffected.

5. Thermal Properties in Core, Cuttings and Outcrop

5.1 Thermal Conductivity

Thermal conductivity of both core and cuttings were measured using a divided bar apparatus calibrated against fused quartz and water standards (Figure 7). Checks with calibration standards showed an accuracy of ± 5%. The cuttings were saturated with water after being evacuated to remove most of the air in the measurement cells. A matrix thermal conductivity for the cuttings at 33 m intervals was initially derived after correcting for the mass of water in the cells. The in situ thermal conductivity (at 25°C) was then calculated using the porosity from the formation density log. Both corrections used the geometric formula for mixtures:

$$K\,(\text{mix}) = K_1\text{\textasciicircum}(v_1) * K_2\text{\textasciicircum}(v_2)$$

where K is the thermal conductivity of the mixture and the two components, and v is the fractional volume of each component. There is a wide range in thermal conductivity (2–4 W/m°C) in the granitic rocks below 1000 m depth due to compositional variations. The uncertainty based on repeat measurements is ± 10%. The main component influencing thermal conductivity is quartz, which varies from varies from an average of 5% by volume in the rock with a conductivity of 2.1 – 2.6 W/m°C, to 15 – 35% with conductivities of 2.7 – 3.9 W/m°C.

5.2 Specific Heat

Specific heat was measured with a calorimeter calibrated with water and fused quartz standards. Rock samples were heated to 72°C and immersed in water at room temperature, and the resulting temperature rise was measured to better than 0.01°C. The measurement uncertainty was 0.08 kJ/kg°C, or 10%. Measurements ranged between 0.7 and 1.0 kJ/kg°C (Table 1).

	thermal conductivity	**specific heat**	**thermal diffusivity**	**heat productivity**	**density**
	W/m°C	kJ/kg°C	mm^2/s	$\mu W/m^3$	g/cm^3
25°C	2.7 ± 0.4	0.8 ± 0.1	1.2 ± 0.3	3	2.7 ± 0.1
~ 200°C *	2.3	1.1	0.8	3	2.7
*** inferred**					

Table 1. Summary of thermal properties for granite at the FORGE site, and the equivalent in situ values at about 200°C based on trends established in the literature.

5.3 Thermal Diffusivity

Thermal diffusivity has been calculated using the relationship:

$$\propto = \frac{K}{\rho * c}$$

where $\propto$ is the thermal diffusivity, K is thermal conductivity, ρ is density, and c is specific heat. Values range from 0.7 to 1.7 mm^2/s, with an uncertainty of 15% (measurements at 25°C).

5.4 Heat Generation

Radioactive decay of isotopes of potassium, uranium, and thorium releases heat. Rocks with high concentrations of these elements can generate sufficient heat to be a significant factor augmenting surface heat flow when present in thicknesses of kilometers. Spectral gamma measurements were made on the cuttings using a hand-held spectrometer on 4–6 kg of cuttings in a steel box to shield external gamma radiation. The values have been scaled to match the API value for total gamma obtained at the same depth from the downhole logging. The resulting average for "granitic" rock in 58-32 (granite, quartz monzonite, monzonite) and "dioritic" rock (quartz monzodiorite and diorite) are shown in Table 2.

Table 2. Heat generation characteristics of the granitic and dioritic rock encountered in well 58-32.

	K (%)	U (ppm)	Th (ppm)	density (kg/m^3)	Heat generation (μW/m^3)
granitic	4.0	6.7	12.4	2700	3.0
dioritic	2.4	4.1	7.9	2800	2.0

The heat generation of 2–3 μW/m^3 is typical for granitic rock, but its effect on the geotherm is small because of the relatively high conductive heat flow at this site.

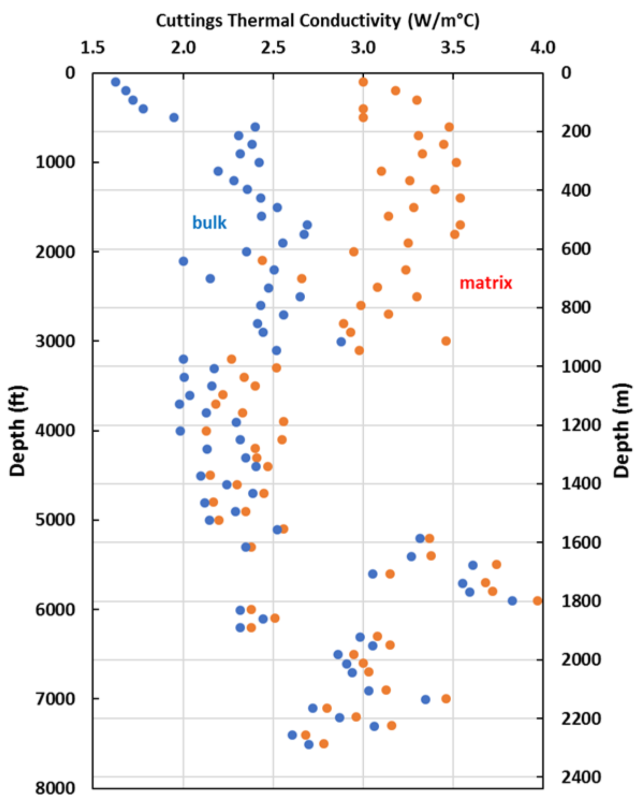

Figure 7. Thermal conductivity data for well 58-32 based on cuttings measurements using a divided bar. Matrix thermal conductivity values are corrected for the effects of rock porosity using the wireline geophysical logs (formation density log; Gwynn, 2018). The variation in thermal conductivity is largely due to the quartz content of the rock. The transition from granitic basin fill to granitic bedrock occurs at 1000 m.

5.5 Effect of temperature on thermal properties

Both thermal conductivity and thermal diffusivity decrease with increasing temperature. Based on the range of measurements shown in Figures 7 and 8, and measurements of eastern U.S granites at various temperatures (Robertson, 1988), the reservoir thermal conductivity will be about 15% lower than the laboratory measurements at room temperature (Table 1). Specific heat typically increases from about 0.7 to 1.0 kJ/kg°C between 25 and 200°C (Whittington et al., 2009). Recent measurements also suggest the typical crustal rock thermal diffusivity decreases from about 1.7 to 1.0 between 25°C and 200°C (Whittington et al., 2009). The in situ physical properties for granite at about 200°C are shown in Table 1.

6. Heat Flow

The heat flow map based on temperatures at less than 200 m depth (Gwynn et al., 2016; Figure 8) delineates the conductive heat loss from the top of the Roosevelt hydrothermal system and its outflow plume. Where the heat flow is less than about 100 mW/m^2 and the surface temperature intercept is 10–15°C, the near-surface temperature gradient may be a reliable indicator of temperature at greater depth. Over most of the higher heat flow areas of Figure 8 there is thermal groundwater, and shallow temperature gradients usually cannot be extrapolated to depth.

In well 58-32 the deepest portion of the well, which is thought to be close to thermal equilibrium when the log was run 37 days after drilling and testing ceased, the temperature gradient of 70°C/km and average thermal conductivity of 2.5 W/m°C (in situ temperature; 2.9 W/m°C at 25°C) implies a deep heat flow of 180 mW/m^2 with an estimated uncertainty of ± 20 mW/m^2. This contrasts with the heat flow over the uppermost 250 m in this well where the gradient is high but strongly decreasing with depth (Figure 6); using an average of 250°C/km and a thermal conductivity of 1.5 W/m°C implies a near-surface conductive heat flow of over 350 mW/m^2. The near-surface heat flow is double the deep heat flow because of a lateral flow of hot groundwater across the site.

The heat flow in Acord-1 has previously been estimated at 120 ± 20 mW/m^2 using both shallow and deep temperatures (Gwynn et al., 2016). Well 9-1, 3 km southeast of 58-32, has a deep gradient of 60°C/km and is likely to have a thermal conductivity similar to 58-32 because of the similarity in granitic rocks in this well. The deep heat flow is therefore 150 mW/m^2 with an uncertainty of at least ± 20 mW/m^2. The difference in deep heat flow between 58-32 and 9-1 is not significant given the magnitude of uncertainties.

The well with the highest temperature gradient below 1 km depth on Figure 5 is 24-36. This is located 3 km east of the Opal Mound fault, near the east end of a prominent valley in the Mineral Mountains (Figure 1). This exploration well was unproductive and was abandoned. The gradient of 85°C/km and an assumed in situ thermal conductivity for the granite of 2.5 W/m°C gives a deep heat flow 210 ± 20 mW/m^2.

The main source of uncertainty in the calculated heat-flow value is the thermal conductivity. The results from 58-32 show a variation of over 50% (2.1–3.9 W/m°C at 25°C) depending on whether the rock is predominantly dioritic or granitic. Detailed logging in 58-32 shows this compositional variation can occur on a scale of centimeters to 100 meters, making it difficult to assess average values. The relatively similar gradients of 50–90°C/km at depth in most of the unproductive exploration wells indicate a large volume of thermally conductive rock surrounding the Roosevelt hydrothermal system with a temperature of ~ 200°C at about 3 km depth. The heat flow in 58-32 allows extrapolation of the thermal regime to greater depth, and when the effect of decreasing thermal conductivity with increasing depth and temperature is included, a temperature of 400°C is predicted at 5 km depth and 600°C at 7 km depth.

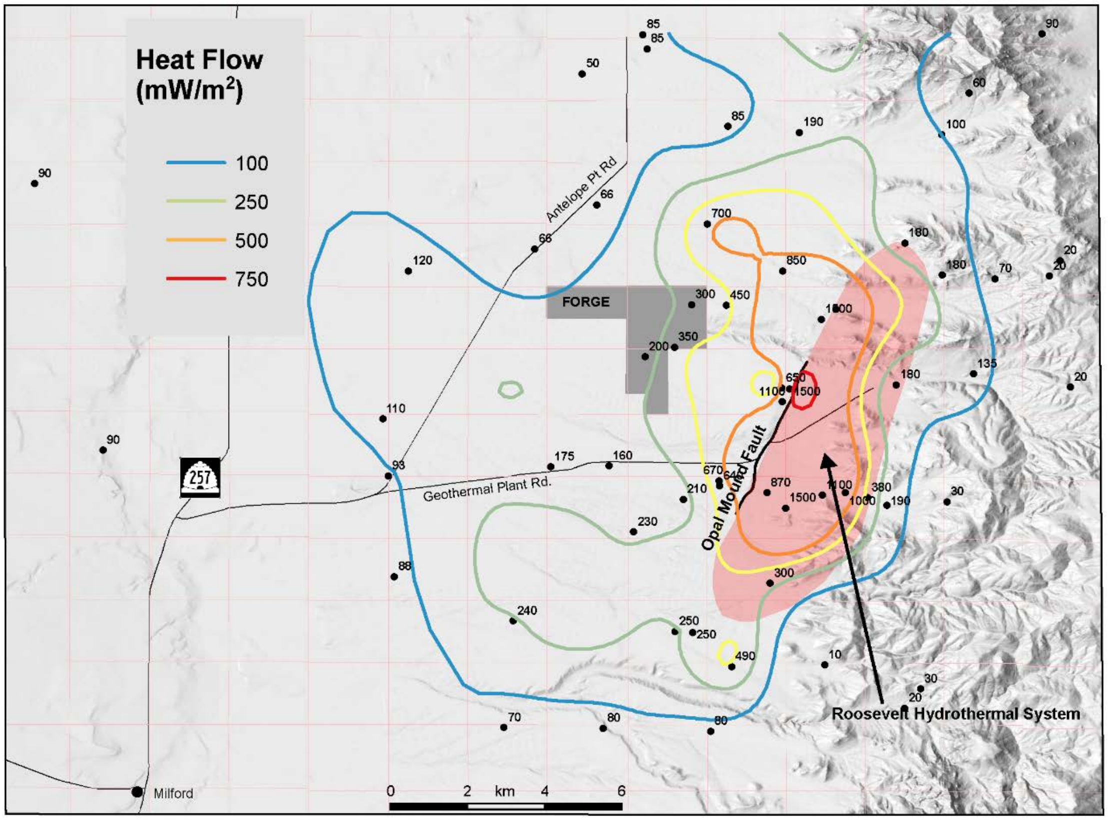

Figure 8. Contours of conductive heat flow derived from wells around the Roosevelt Hot Springs hydrothermal system (mW/m2; Gwynn et al., 2016). Most values are from the gradients in wells less than 200 m deep. Very high heat flows over the hydrothermal system reflect high temperature gradients overlying upflowing hot water, which at shallow depth is constrained by boiling-point-for-depth conditions. West of the hydrothermal system (that is, the Opal Mound fault) the thermal regime is conductive at depth and follows a pattern of decreasing heat flow towards the west.

7. Isotherm Maps

Isotherm maps that integrate the thermal gradient well data with the temperature profiles in deeper wells outside of the hydrothermal system, and take in account the contrast in thermal conductivity between basin fill and granitic rock, are shown in Figure 9. Uncertainties increase with increasing depth and with increasing distance from deep wells. The area or inferred impermeable rock at a temperature of more than 200°C in these maps (west of the Opal Mound fault) increases from about 25 km^2 at 2 km depth, to 70 km^2 at 3 km depth, and to about 100 km^2 at 4 km depth.

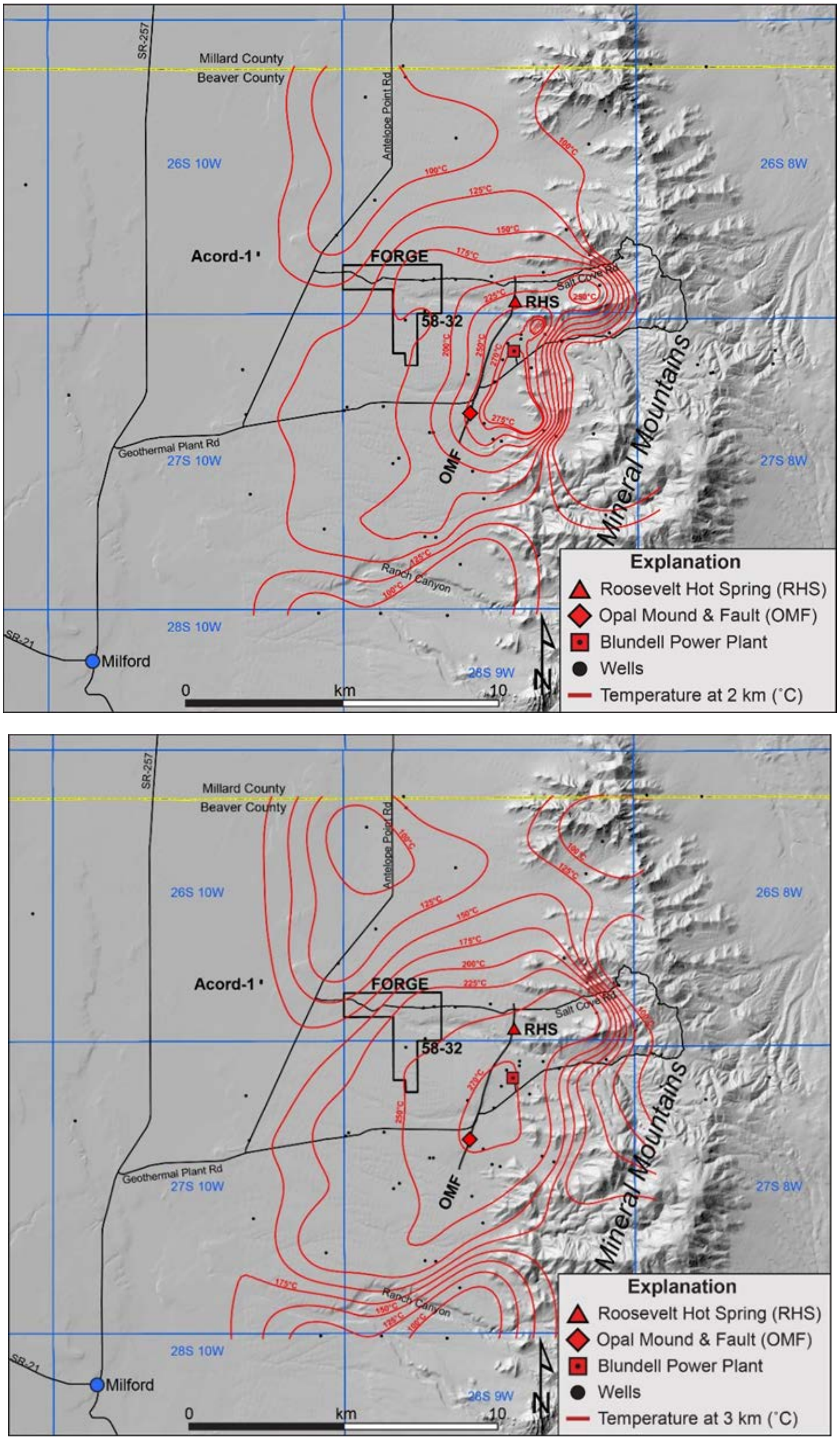

Figure 9. Isotherm maps at 2 and 3 km depth based on analysis of the temperatures in thermal gradient wells and deeper exploration wells.

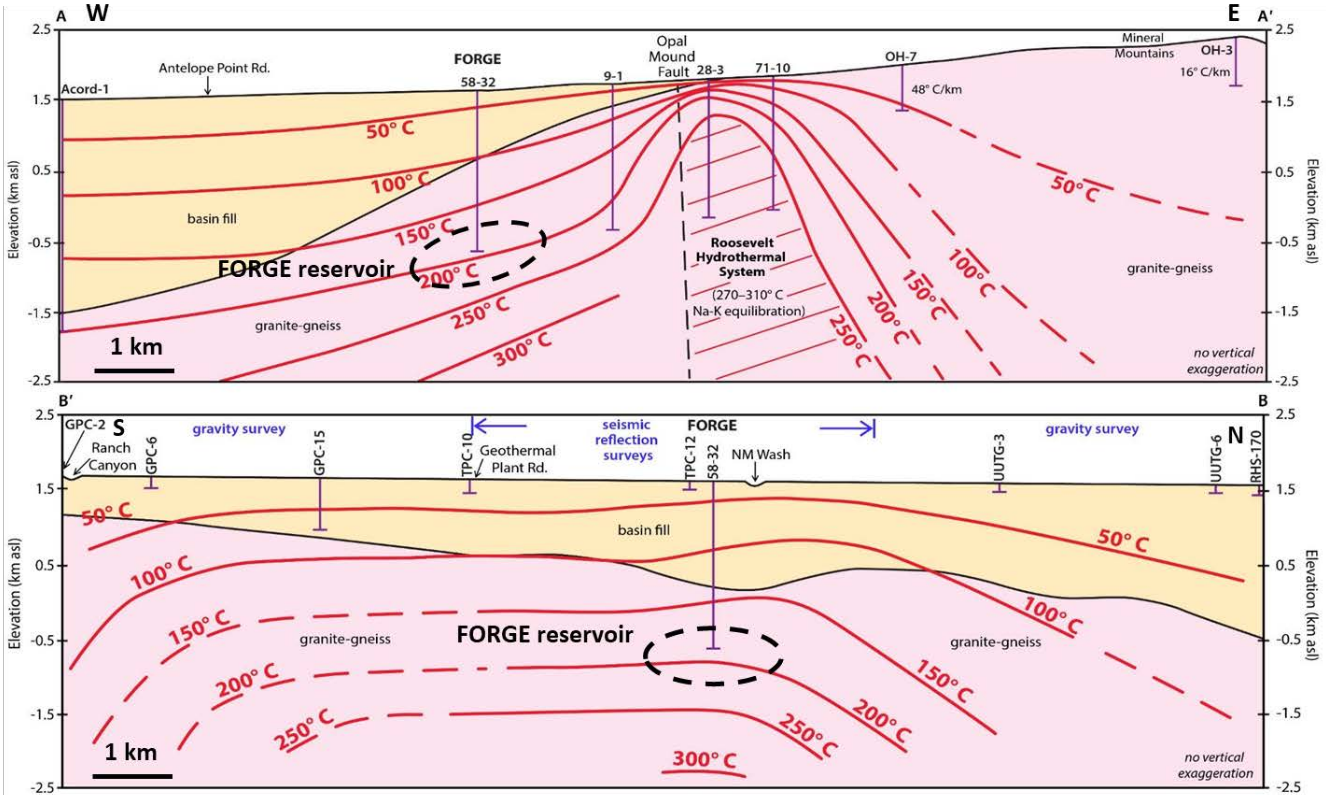

Figure 10. Two thermal cross sections based on integration of data from the many wells (locations in Figure 1). The contours are dashed where they are uncertain. The granite surface in B–B′ is based on preliminary interpretations from the 3-D seismic reflection survey near the FORGE site (Miller, 2018) and from gravity models at greater distance (Hardwick et al., 2018). Well 58-32 is 1 km east of line B–B′ and has been projected onto that line. The granite surface at that point is at 1.4 km depth, compared to 1.0 km depth in 58-32.

Conclusions

The drilling of FORGE well 58-32 confirms that a large area of hot, impermeable granitic rock lies to the west of the Opal Mound fault. Well 58-32 has a conductive thermal gradient from about 600 m depth to the bottom of the hole at 2.3 km depth. Based on inferred thermal equilibrium near the bottom of the hole at the time of thermal logging 37 days after the rig departed, the gradient averages 70°C/km and the heat flow is 180 ± mW/m^2. The future FORGE reservoir, which has been specified by DOE to range between 175 and 225°C, would be situated between 2.0 and 2.7 km depth. The area of inferred impermeable granite at a temperature at more than 200°C determined from isotherm maps increases from about 25 km^2 at 2 km depth, to 70 km^2 at 3 km depth, and to about 100 km^2 at 4 km depth. The thermal properties of core, cuttings and outcrop measured in this study confirm values typical for granitic rock, but with significant variations caused by varying quartz content. Observed variations from dioritic to granitic rock have quartz contents between 2% and 35% respectively, causing the thermal conductivity to vary from 2.1 to 3.9 W/m°C. This causes some of the thermal properties to also be sensitive to temperature, and measurements at room temperature may need up to 20% corrections for the expected reservoir conditions.

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References

Allis R. G. and Larsen, G. "Roosevelt Hot Springs Geothermal field, Utah – reservoir response after more than 25 years of power production." *Proceedings: 37th Workshop on Geothermal Reservoir Engineering*, Stanford University, Stanford, CA, 2012.

Allis, R.G., Gwynn, M., Hardwick, C., Kirby, S., Moore, J., and Chapman, D. "Re-evaluation of the pre-development thermal regime of Roosevelt Hot Springs geothermal system, Utah." *Proceedings, 40th Workshop on Geothermal Reservoir Engineering*, Stanford University, Stanford, CA., 2015.

Allis, R., Gwynn, M., Hardwick, C., and Moore, J. "The challenge of correcting bottom-hole temperatures – An example from FORGE 58-32, near Milford, Utah." *Proceedings, 43rd Workshop on Geothermal Reservoir Engineering*, Stanford University, Stanford, CA. 2018.

Gwynn, M., Allis, R., Hardwick, C., Hill, J., and Moore, J. 2016. A New Look at the Thermal Regime around Roosevelt Hot Springs, Utah. *Transactions*, Geothermal Resources Council, 40, 551-558.

Gwynn, M., Allis, R., Hardwick, C., Jones, C., and Hurlbut, W. "Compilation of rock properties from well 58-32, Milford, Utah, FORGE site." Chapter to final report of Milford, Utah FORGE site, 2018.

Hardwick, C., Hurlbut, W., and Gwynn, M. "Gravity survey of the Milford, Utah, FORGE site." Chapter to final report of Milford, Utah FORGE site, 2018.

Jones, C. "Jan. 2018 quarterly report summarizing XRD and petrographic activities for the FORGE project." Chapter to final report of Milford, Utah FORGE site, 30 pages, 2018.

Lynne, B.Y., Campbell, K.A., Moore, J.N., and Browne, P.R.L. "Diagenesis of 1900-year-old siliceous sinter (opal-A to quartz) at Opal Mound, Roosevelt Hot Springs, Utah." *Sedimentary Geology*, **179**, 249-278, 2005.

Miller, J., and Allis, R. "Seismic Reflection Profiling at the FORGE Utah EGS laboratory." Chapter to final report of Milford, Utah FORGE site, 2018.

Robertson, E. "Thermal properties of rocks." *U.S.G.S. Open File Report 88-441*, 106 pages, 1988.

Simmons S., Kirby, S., Allis, R., Moore, J., and Fischer, T. "Update on production chemistry at Roosevelt Hot Springs, Utah." *Proceedings, 43rd Workshop on Geothermal Reservoir Engineering*, Stanford University, Stanford, CA, 2018.

Utah FORGE. "Frontier Observatory for Research in Geothermal Energy – Milford Site, Utah Section V. APPENDIX A1 – Conceptual Geologic Model." Appendix to final report of Milford, Utah FORGE site, 2018.

Ward, S.H., W.R. Parry, W.P. Nash, W.R. Sill, K.L. Cook, R.B. Smith, D.S. Chapman, F.H. Brown, J.A. Whelan, and J.R. Bowman. "A Summary of the Geology, Geochemistry, and Geophysics of the Roosevelt Hot Springs Thermal Area, Utah." *Geophysics*, 43, 1515–1542, 1978.

Whittington, A., Hofmeister, A, and Nabelek, P. "Temperature-dependent thermal diffusivity of the Earth's crust and implications for magmatism." *Nature*, 458, 319-321, 2009.

Ziagos, J., and Blackwell, D. "A model for the effect of horizontal fluid flow in a thin aquifer on temperature-depth profiles." *Transactions, Geothermal Resources Council*, 5, 221-223, 1981.

Ziagos J., and Blackwell D. "A model for the transient temperature effects of horizontal fluid flow in geothermal systems." *Jl. Volcanology and Geothermal Research*, 27, 371-397, 1986.