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THERMAL SIGNATURE OF SUBSURFACE FLUID FLOW NEAR THE DIXIE VALLEY GEOTHERMAL FIELD, NEVADA

Colin F. Williams
U.S. Geological Survey
Menlo Park, CA 94025

John H. Sass
and
Frederick V. Grubb
U.S. Geological Survey
Flagstaff, AZ 86001

ABSTRACT

Most of the geothermal development in the United States during the past decade has occurred within the Great Basin. One of the largest developments to date has been the Dixie Valley Geothermal Field (DVGF), one of a number of geothermal fields located along the southern margin of the "Battle Mountain High," a region of high heat flow (> 100 mW/m^2) extending over much of the northern Great Basin. The DVGF lies east of the Stillwater Range and is centered on a southwest-northeast trending thermal anomaly generated by hot water moving up the Stillwater fault, a basin-bounding normal fault characterized by pervasive fracturing and hydrothermal alteration. As part of a multidisciplinary investigation of fracture permeability, in-situ stress, and fluid flow within the Dixie Valley hydrothermal system, we measured subsurface temperatures in four deep wells bordering the DVGF. Precision temperature logs from two flowing wells (45-14 and 66-21) provide detailed information on the location and in-flow rates of permeable fractures intersecting the well bores. Temperature logs from two shut-in wells (76-28 and 62-21) provide information on conductive heat flow near the DVGF. Analysis of the combined dataset yields estimates of the rate and resulting thermal effects of fluid flow along the Stillwater fault. In well 66-21, located 6 km southwest of the current producing limits of the DVGF, and in well 45-14, located 17 km southwest of the DVGF, water flowing up the Stillwater fault may increase the measured heat flow by 20 to 40% over the regional value. In wells 62-21 (4 km southeast of the DVGF) and 76-28 (2 km northeast of the DVGF), heat flow above the fault is close to the regional value, despite the presence of permeable fractures at depth. The limited spatial extent of anomalous heat and fluid flow associated with the DVGF is consistent with deep crustal thermal conditions indicated by the maximum hypocentral depths of nearby earthquakes and suggests that permeability enhancement along the Stillwater fault is highly localized to regions a few kilometers in extent.

INTRODUCTION

Over the past decade, geothermal power developments in the Basin and Range Province of the western United States have reached an installed capacity of approximately 510 MW (Benoit, 1994). Of the fourteen producing geothermal reservoirs located in the Basin and Range, ten are associated with a regional thermal feature known as the "Battle Mountain High" (Sass et al., 1971; Sass et al., 1981), which covers much of the northern Basin and Range (hereafter referred to as the Great Basin) and is characterized by conductive heat flow exceeding 100 mW/m^2. The "typical" geothermal reservoir in the Great Basin lies within basin-bounding normal faults produced as a consequence of Cenozoic extension throughout the region (e.g. Blackwell, 1983). These extensional processes are directly responsible for both the permeable fault zones and the high heat flow necessary for the circulation of thermal waters.

One of the largest developments in the Great Basin is the Dixie Valley geothermal field (DVGF), operated by Oxbow Geothermal Corporation. Dixie Valley is located in west-central Nevada between the Stillwater Range to the northwest and the Clan Alpine Mountains to the southeast (Figure 1). In addition to the high heat flow typical of the northern Great Basin, Dixie Valley lies within an area of active seismicity and late Cenozoic volcanic activity (Thompson and Burke, 1973; Wallace, 1984). The DVGF produces approximately 62 MW of electric power from a series of wells drilled along the Stillwater fault zone (SFZ), an active basin-bounding

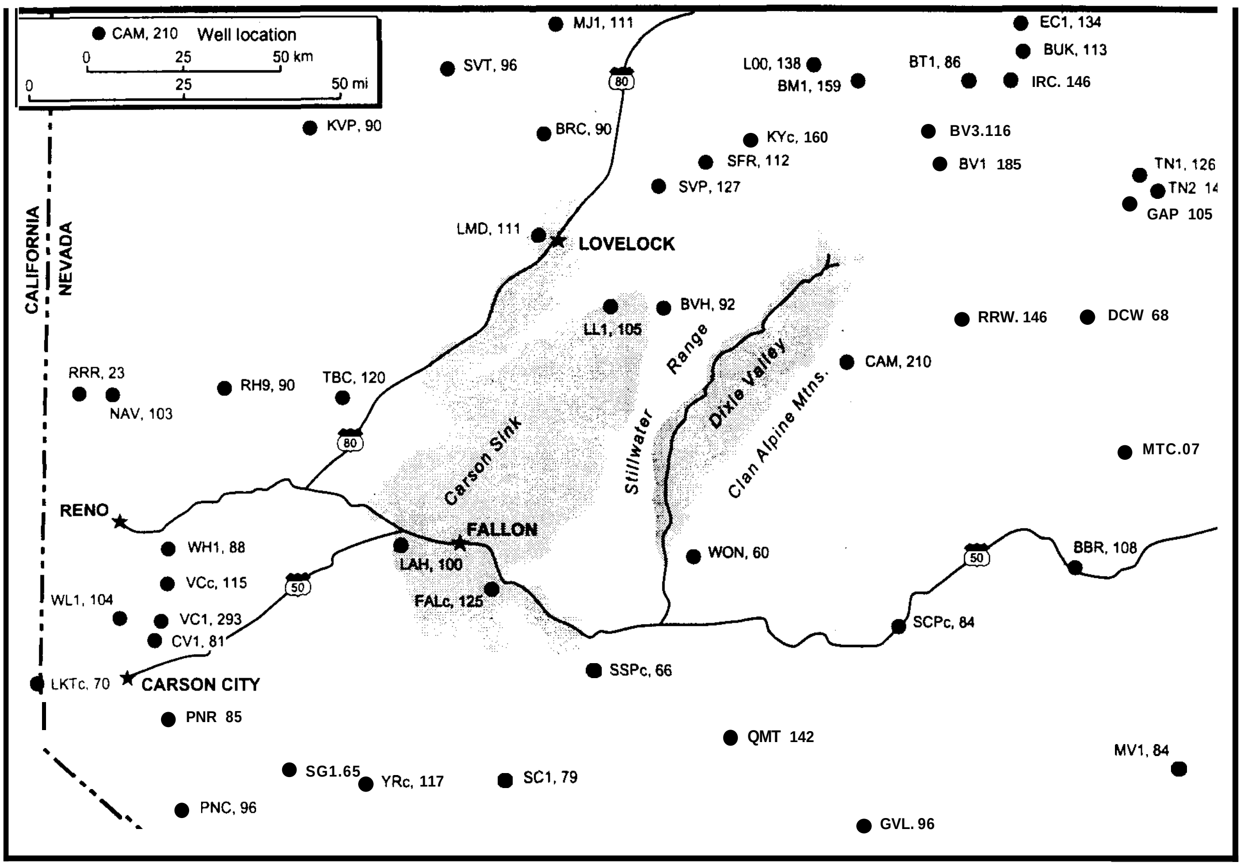

Fig. **1.** Map showing heat flow measurements (mW/m^2) in the vicinity of Dixie Valley, Nevada.

normal fault which has displaced the floor of the basin relative to the adjacent Stillwater Range by approximately 3 km over the past **10** Ma (Figure **2;** Okaya and Thompson, **1985).** The fault cuts and deforms a Jurassic multiphase intrusive body (the Humboldt Lopolith; Speed, **1976),** and the subsurface intersection of the SFZ with the lopolith serves as the primary producing zone for the DVGF (Benoit, in press, **1997).**

Hot water moving up the SFZ and other subsidiary faults and fractures introduces abnormally high temperatures in a narrow zone along the front of the Stillwater Range. Temperature gradients along this zone have been mapped in the range from **100** to greater than 200°C/km, with the corresponding heat **flows** exceeding 300 $\mathbf{mW/m^2}$ in places.

This paper reports on thermal **aspects** of an ongoing, multidisciplinary study of the factors controlling spatial and temporal variations in fracture permeability within and around the DVGF (Hickman and Zoback, this volume and Barton et al., this volume). We examine thermal data from wells bordering the DVGF and analyze evidence for the magnitude and extent of fluid flow up the SFZ at these locations. Comparison of these results with information from the DVGF proper should help provide quantitative constraints on the factors controlling the development of permeability on the SFZ.

DATA

The primary thermal data used in this study come from four observation wells **(45-14, 66-21, 62-21** and **76-28)** located southwest, southeast and northeast of the DVGF (Figure **2).** Wells **45-14** and **66-21** were drilled in **1979** and have **been** left open at the surface, allowing continuing flow up the wellbore from permeable fractures at depth. Well **62-21** was drilled in **1980** and also intersects significant permeability at depth, although the well has **been** kept shut-in at the surface. Well **76-28** was drilled **in** **1984** and differs from the other three in that it is not artesian. The fluid level in this well is encountered at a depth of 550 meters **(1800** feet). These and most

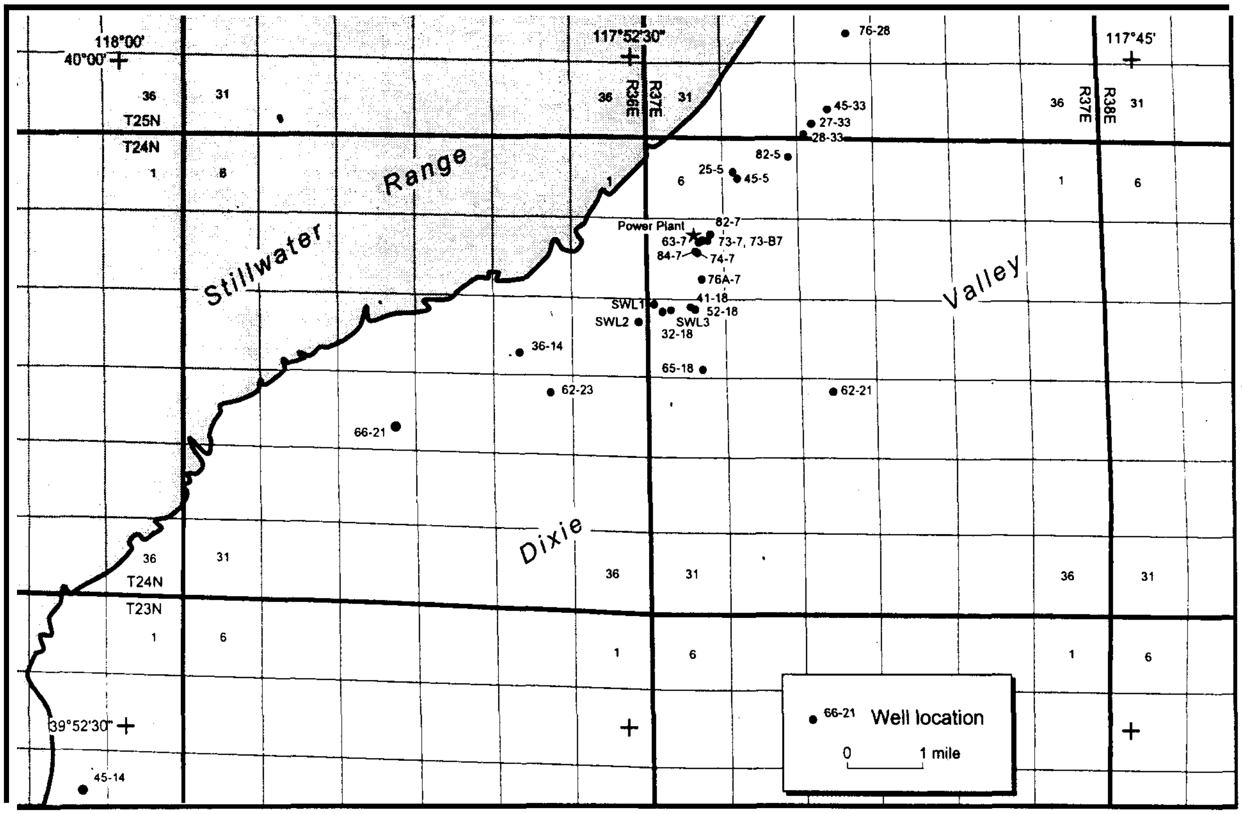

Fig. 2. Map showing locations of geothermal wells (producing, non-producing, and injecting) in Dixie Valley.

other wells in Dixie Valley pass through more than 1000 meters of basin-filling sediments and volcanics before entering the Mesozoic igneous and metamorphic rocks comprising the basement (e.g. Waibel, 1987). The four study wells are cased through the younger units and into the Mesozoic basement, with the length of the open hole sections in the basement ranging from 780 to 880 meters.

Temperature data acquired from all four wells in August, 1996, are shown in Figure 3. The primary factor controlling the differences among the four temperature profiles is the presence or absence of fluid flow within the wellbore. In well 76-28, temperatures follow a gradient varying between 35 and 70°C/km from the water level at 550 meters to the bottom of the log at 2350 meters. The measured gradient variations correspond well with changes in lithology from alternating sand- and clay-rich valley fill to volcanic tuffs to metasediments, with the highest gradients corresponding to the high porosity alluvium and the lowest to the low porosity metasediments. There is no thermal evidence for substantial fluid movement within well 76-28.

In well 62-21 temperatures follow the same general pattern as those within 76-28 with some notable exceptions. Temperature gradients reach or exceed 50°C/km in the sediments and volcanics of the upper 1800 meters and then decline to between 25 and 45°C/km in the mafic igneous and metamorphic rocks of the Humboldt lopolith and the underlying Triassic metasedimentary unit. Curvature in the temperature profile from 2200 to 3000 meters is consistent with entry of fluid into the wellbore (or behind the casing) at 2200 meters (near the top of the lopolith section) and consequent downflow with an exit at about 3000 meters (just below the base of the casing in the Triassic metasediments). Differences between the temperature gradient above 2200 meters and below 3000 meters reflect contrasting thermal conductivities.

The temperature profile in well 66-21 is dominated by the thermal signature of persistent flow up the wellbore to the surface, with abrupt offsets in the temperature profile opposite fluid entries from fractures within the lopolith. This upward flow results in the relatively high surface temperature (~42°C). The other prominent feature of the 66-21 temperature profile, a reversal with a temperature minimum at 2740 meters, appears to be a transient of unknown origin. This feature may reflect entry of gaseous fluid into the wellbore or transient convective instabilities. Subsequent temperature logs showed a

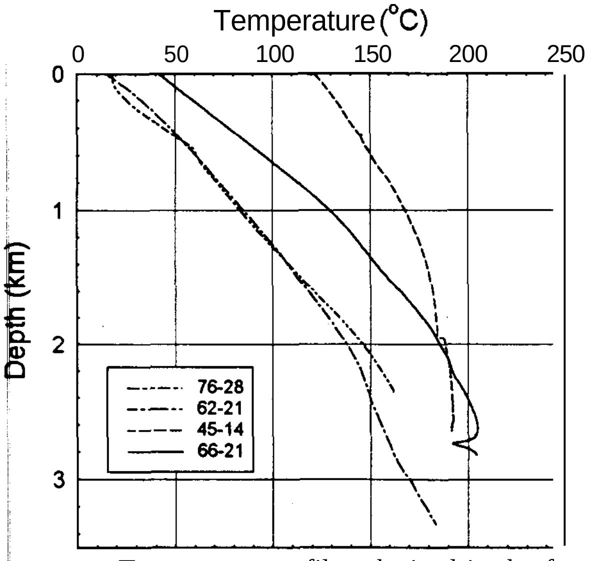

Fig. 3. Temperature profiles obtained in the four study wells in August, 1996.

diminishing anomaly, with nearly complete restoration of equilibrium temperatures in 24 days (C. Barton and R. Norman, written comm.).

More rapid upflow leads to higher temperatures in 45-14, with water entering from fractures within the phyllite basement (the lopolith is not present within this well) and exiting at the surface at a temperature of 122°C. As with well 66-21, the main fluid entries are easily located from offsets in the temperature profile.

THERMAL. EFFECTS OF WELLBORE WATER FLOW

In order to determine the thermal effects of naturally occurring fluid flow along the SFZ and related faults and fractures, the anomalous effects of fluid flow within the wellbore first have to be removed from the measured profiles. An analytical model for the long-term thermal effects of vertical fluid flow in a well was presented by Ramey (1962). The equation for temperature above the entry point takes the form of

$$T(z) = T(0) + \Gamma\cdot(z-z_f) - (\exp((z-z_f)/A)-1)\cdot\Gamma A \quad (1)$$

in which T is temperature, z is depth, z_f is the depth of fluid entry, Γ is the undisturbed geothermal gradient, and A is a measure of the rate of heat transfer. This time dependent factor is given by

$$A = v\rho_f C_f r^2 f(t)/2\lambda \quad (2)$$

in which v is the velocity of the fluid, ρ_f is the density of the fluid, C_f is the specific heat of the fluid, r is the radius of the borehole, f(t) is a time function describing the thermal response of the rock formation to the fluid flow, and λ is the thermal conductivity of the rock formation. This model is best applied when the ratio $\alpha t/r^2$ (where α is the thermal diffusivity) is greater than 1000 (Drury, 1984), which for a normal range of borehole radii is reached on the order of weeks to months. At these times, the function f(t) can be approximated by

$$f(t) = -\ln\left(\frac{r}{2\sqrt{\alpha t}}\right) - \frac{\Gamma}{2} \quad (3)$$

where Γ is Euler's constant (0.5772.. .).

Equations (1) through (3) were applied to the temperature profiles from wells 45-14 and 66-21. Figure 4 illustrates sample curve-fits of equation (1) to temperatures in well 45-14 resulting from fluid entries at depths of 1940 and 2510 meters. The temperature profile is well-matched with A= 1400 and 1500 meters, although variations of A by 10% in either direction are also close to the observed profile. With the assumption that well 45-14 has been flowing at the same rate for approximately 15 years, the estimated rate of flow is 1.1 l/s or 14 gpm.

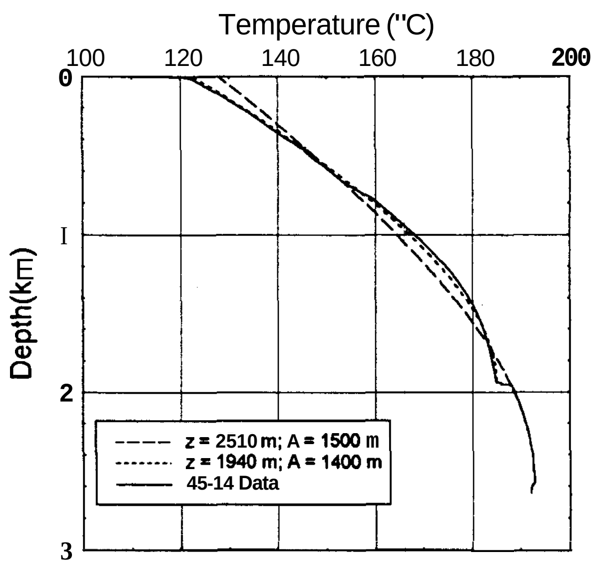

Fig. 4. Temperature profile from well 45-14 along with modeled temperatures for fluid entries at depths of 1940 and 2510 m.

Application of the same analysis to a fluid entry at a depth of **2260** meters in well **66-21** gives $A = 250$ meters (Figure ***5)*** and a flow rate of **0.17** l/s or **2.5** gpm. This is close to the measured flow from **66-21** of **0.12 l/s** or **1.8** gpm **(S.** Hickman, pers. comm.)

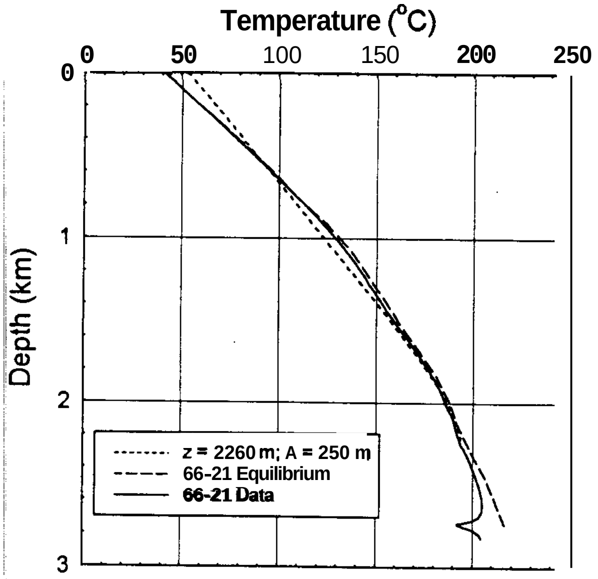

Fig. ***5.*** Temperature profile from well **66-21** along with modeled temperatures for fluid entry at depth of **2260** m. Equilibrium curve reflect temperatures after dissipation of the transient centered at **2800** m.

The flow rates estimated from the Ramey **(1962)** model are likely to deviate from the true value by 10 to **20%** due to the effects of vertical conduction in the rock adjacent to the well (Beck and Shen, **1987)** and to the vertical variation in the undisturbed geothermal gradient due to variations in formation thermal conductivity. This is particularly noticeable in the results from well **45-14.** The additional fluid entering at **1940** meters should increase the value of A above this point, but the model yields a slightly lower value. However, the primary value of this analysis lies in providing an estimate of the depth section disturbed by the wellbore flow. For distances more than 3A from the fluid entry point, the measured gradient varies by less than ***5%*** from the true gradient (Drury, **1984).** Consequently, in well **66-21** temperature gradients from the upper portion of the well and the temperature of the lowest fluid entry can be used in estimating undisturbed thermal conditions. In well **45-14** the high rate of flow **as** indicated by the large value of A leads to depressed gradients through the entire depth of the well. Only the temperature of the lowest fluid entry can be equated with the undisturbed formation temperature and used to estimate the overall geothermal gradient.

THERMAL EFFECTS OF FLUID FLOW ON **THE** STILLWATER FAULT

Removal of the effects of wellbore flow in wells **45-14** and **66-21** provides information **on** undisturbed deep thermal conditions in all four wells. Available thermal conductivity data are limited, but some simple assumptions regarding the similarity of lithologies encountered in each of the four wells provides useful information on relative differences in conductive heat flow.

In well **76-28,** application **of** alluvium thermal conductivity values from nearby exploratory heat-flow holes (M. Walters, written comm.) give an estimated heat flow of **110** mW/m^2, a typical background value for this part of Nevada. This in turn yields an approximate average thermal conductivity for the metasedimentary section of **2.5** $W/m{\cdot}K$, **assuming** constant heat flow with depth. Applying these values to the calculated temperature gradient in well **45-14** (which penetrates a similar section of alluvium, volcanics **and** metasediments) yields an estimated heat flow of **140** mW/m^2. Application of the same thermal conductivity data to well **62-21** yields a heat flow **90** mW/m^2, which equates to a lopolith thermal conductivity of **2.6** $W/m{\cdot}K$. Data from the equivalent section in **66-21** yields a value of **130** mW/m^2.

If the relative differences in apparent heat flow among the four wells are significant, then it is possible to estimate the thermal effect of flow up the SFZ at the locations where it is penetrated by wells **45-14** and **66-21. A** simple **2-D** model for the change in heat flow across an inclined fracture with water moving along the fracture was derived by Lewis and Beck **(1977).** If the flow along the inclined fracture has persisted for enough time to develop thermal equilibrium above and below the fracture, the difference in conductive heat flow across the fracture can be determined **as**

$$\Delta q = W C_f \Gamma \sin(\theta) \qquad (4)$$

where W is the mass rate of flow per unit length of the fracture, C is the heat capacity of the fluid, Γ is the undisturbed geothermal gradient, and θ is the dip of the fracture plane. For the SFZ, θ is approximately **52°,** and the difference in heat flow **between** wells with heat flow elevated by flow up the SFZ **(66-21, 45-14)** and wells with relatively undisturbed heat flow **(62-21, 76-28)** provides a Aq = **20-40** mW/m^2. With these values we find W =

1.4-2.8 x 10^{-4} kg/m.s. This works out to 4.4 to 8.8 m^3/yr for each meter of fault length southwest of the DVGF, compared to estimated flow rates of 23 to 46 m^3/yr for each meter of fault within the DVGF itself (Benoit, this vol.).

The validity of this model is limited by two primary factors. First is the assumption that increased heat flow in wells 45-14 and 66-21 is due to flow along the SFZ alone and does not reflect spatial variations in thermal properties, thermal refraction adjacent to the range front, or other unknown advective or conductive processes. Second is the 2-dimensional nature of the model. Near well 45-14 in particular, there is evidence from shallow temperature-gradient holes for 3-dimensional focusing of the thermal anomaly associated with the SFZ (Koenig et al., 1976). Given these limitations, the analysis may not provide precise quantitative information on the magnitude of flow up the SFZ but does indicate the spatial extent of flow outside of the DVGF.

In summary, subsurface temperatures and resulting heat flow values in wells 76-28 and 62-21 are consistent with regional averages. Subsurface temperatures and heat flow values in wells 45-14 and 66-21 are higher, perhaps by as much as 40%. These higher values of heat flow are consistent with local fluid flow up the SFZ southwest of the DVGF. The SFZ should lie at a depth of approximately 6 km below the site of well 62-21, yet the thermal data suggest the heat flow is consistent with the regional average. An implication of this observation is that thermally significant fluid flow up the SFZ is not a factor at depths greater than 3 to 4 km in the vicinity of well 62-21. The validity of these results can be tested by considering other available information on deep crustal thermal conditions: well-established relationships between the maximum depth of seismic faulting and crustal thermal conditions provide one such approach.

SEISMIC EVIDENCE OF DEEP CRUSTAL THERMAL CONDITIONS

As noted above, Dixie Valley lies in a region of active seismicity. Most notably, in the 1954 Rainbow Mountain-Fairview Peak-Dixie Valley sequence, five earthquakes of moment magnitude (M) ranging from 5.9 to 7.2 ruptured the faults in and south of Dixie Valley (Doser, 1986). The M=6.7 Dixie Valley earthquake was located approximately 30 km south of the DVGF and was responsible for up to 2 m of vertical offset along the SFZ to the southwest of the DVGF (Slemmons, 1957). Relocation of the hypocenter for the Dixie Valley earthquake places it at a depth of 12 ± 3 km (Doser, 1986), with similar results for the other, more distant, earthquakes in the sequence. Numerous studies of thermal constraints on brittle faulting and the nature of the seismogenic crust constrain the base of seismicity in the western United States to correspond with a temperature less than or equal to 400°C (e.g. Sibson, 1982; Williams, 1995). Consequently, the maximum focal depths of the 1954 earthquake sequence provides an additional thermal constraint on temperatures at depth and the corresponding value of heat flow.

Figure 6 shows geotherms determined for the range of heat flow values derived from the four study wells. Geotherms were determined through use of the appropriate equations for one-dimensional steady-state heat conduction with heat production and a temperature dependent thermal conductivity (Williams and Sass, 1995; Williams, 1996). The basic equation is given by

$$T = \frac{1}{b}\cdot[\exp(\frac{bq_s}{\lambda_0}z - \frac{bA_s}{2\lambda_0}z^2 + \ln(1-bT_s)) - 1] \quad (5)$$

where T_s is the surface temperature (15°C), q_s is the surface heat flow (varied), λ_0 is the thermal conductivity at 0°C (3.0 W/m·K), A_s is the surface radiogenic heat production (2.0 $\mu W/m^3$) and b is the temperature coefficient of thermal conductivity (0.0024 - 0.0052/λ_0).

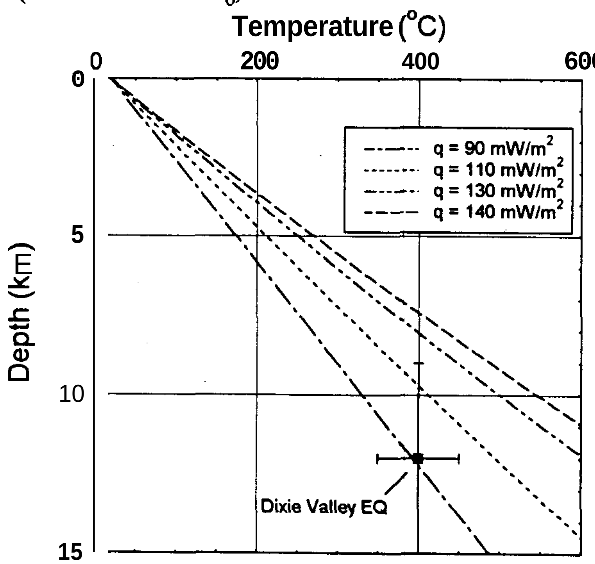

Fig. 6. Crustal temperature profiles determined for the range of estimated heat flow from the four study wells. Note the discrepancy between the temperature inferred for the hypocenter of 1954 Dixie Valley earthquake and temperatures from heat flow values above 110 mW/m^2.

The primary result of interest is the consistency of the hypocentral depth of the Dixie Valley earthquake with temperatures near 400°C for heat flow values of **90** mW/m^2 (well **62-21)** and **110** mW/m^2 (well **76-28).** Temperatures at this depth for heat flow values of **130** mW/m^2 (well **66-21)** and **140** mW/m^2 (well **45-14)** are far too high to represent in situ conditions at the hypocenter of the Dixie Valley earthquake. The consistency of the apparent heat flow in wells **76-28** and **62-21** with the regional average for the northern Great Basin and the focal depths of nearby earthquakes strongly suggests that these sites are not affected by fluid flow along the SFZ. Conversely, the apparent heat flow in wells **45-14** and **66-21** is inconsistent with either the regional average or the earthquake focal depths. Consequently, permeability along the SFZ must be highly localized and flow along the SFZ in the vicinity of the DVGF may be restricted to depths less than **6 km.**

CONCLUSIONS

Analysis of the temperature profiles from the four study wells provides valuable information on coupled heat and fluid flow within Dixie Valley. In particular, in the two wells **(66-21** and **45-14)** located southwest of the DVGF heat flow is elevated by 20 to **40%** above the regional background value. Fluid flow up the SFZ may be responsible for this. By contrast, heat flow southeast **(62-21)** and northeast **(76-28)** of the DVGF is indistinguishable from the regional value. The limited magnitude and spatial extent of anomalous heat and fluid flow is confirmed by comparison with thermal conditions likely to exist at maximum hypocentral depths of nearby earthquakes. Future research to test these preliminary results should involve thermal properties measurements, detailed mapping of subsurface temperature and permeability, and numerical modeling.

OWL

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REFERENCES

Beck, A.E., and P.Y. Shen, **1985,** Temperature distribution in flowing liquid wells, ***Geophysics,*** v. 50, no. **7,** p. **1113-1118.**

Benoit, W.R., **1997,** Injection of geothermal fluid in Nevada **as** typified by the Dixie Valley project, in ***Deep Injection Disposal of Hazardous and Industrial Wastes***, Academic Press, in press.

Benoit, W.R., **1994,** A review of geothed resources in the western Basin and Range, **1994,** ***Geothermal Resources Council Bull.,*** v. **24, n. 5, p. 153-156..**

Blackwell, D.D., Heat flow in the Northern Basin and Range, **1983,** in The ***Role*** of ***Heat in the Development*** of ***Energy and Mineral Resources in the Northern Basin and Range Province, Geothennal Resources Council Spec. Paper 13,*** p. **81-92.**

Doser, D.I., **1986,** Earthquake processes in the Rainbow Mountain-Fairview Peak-Dixie Valley, Nevada region **1954-1959,** ***J. Geophys. Res.,*** v. **91,** no. **12,** p. **12,572-12,586.**

Drury, M.J., **1984,** Perturbations to temperature gradients by water flow in crystalline rock formations, ***Tectonophysics,*** **102,** p. **19-32.**

Koenig, J.B., R.W. Greensfelder, and C.W. Klein, **1976,** Geothermal potential of the Quest leasehold Dixie Valley, Nevada, unpublished technical report, GeothermEx, Inc.

Lewis, T.J., and A.E. Beck, **1977,** Analysis of heat flow data - detailed observations in many holes in a small area, ***Tectonophysics,*** **41,** p. **41-59.**

Okaya, D.A. and Thompson, G.A., **1985,** Geometry of Cenozoic extensional faulting: Dixie Valley, Nevada, ***Tectonics,*** v. **4,** p. **107-125.**

Ramey, H.J., Jr., **1962,** Wellbore heat transmission, ***J. Petrol. Tech.,*** **14, 427-435.**

Sass, J. H., A. H. Lachenbruch, R. J. Munroe, G.W. Greene, T. H. Moses, **1971,** Heat flow in the Western United States, ***J. Geophys. Res.,*** **76,** p. **6376-6413.**

Sass, J.H., D.D. Blackwell, D.S. Chapman, J.K.

Costain, E.R. Decker, L.A. Lawver, and C.A. Swanberg, **1981,** Heat flow from the crust of the United States, in Y.S. Touloukian, W.R. Judd, and R.F. Roy, ***eds., Physical Properties of Rocks and Minerals,*** McGraw-Hill Book Company, p. **503-548.**

Sass, J.H., S.S. Priest, T.A. Ehlers, P. Morgan, D.S. Chapman, A.H. Lachenbruch, and C.F. Williams, **1996,** Thermal regime of the Great Basin, ***EOS,*** v. **77, no. 46,** F665.

Sibson, R.H., **1982,** Fault zone models, heat flow, and the depth distribution of earthquakes in the continental crust of the United States, ***Bull. Seismol. Soc. Am.,*** **72,** p. **151-163.**

Slemmons, D.B., **1957,** Geological effects of the Dixie Valley-Fairview Peak, Nevada earthquakes of December **16, 1954,** ***Bull. Seismol. Soc. Am.,*** **47,** p. **353-375.**

Speed, R.C., **1976,** Geologic map of the Humboldt lopolith and surrounding terrane, Nevada, ***Geol. Soc. Amer. Map and Chart Series,*** **n. 14.**

Thompson, G. A., and D. B. Burke, **1973,** Rate and direction of spreading in Dixie Valley, Basin and Range province, Nevada, ***Geol. Soc. Am. Bull.,*** **84, 627-632.**

Waibel, A.F., **1987, An** overview of the geology and secondary mineralogy of the high temperature geothermal system in Dixie Valley, Nevada, ***Trans. Geothermal Resources Council,*** v. **11,** p. **479-486.**

Wallace, R.E., **1984,** Patterns and timing of late Quaternary faulting in the Great Basin province and relation to some regional tectonic features, ***J. Geophys. Res.,*** v. **89,** p. **5763-5769.**

Williams, C.F., **1995,** Temperature and the seismic/aseismic transition: a new look, EOS, **76,** p. **410.**

Williams, C.F., and J.H. Sass, **1996,** The thermal conductivity of rock under hydrothermal conditions: measurements and applications, ***Proc. 21st Stanford Geothermal Workshop,*** p. **335-341.**

Williams, C.F., **1996,** Temperature and the seismic/aseismic transition: observations from the **1992** Landers earthquake, ***Geophys. Res. Lett.,*** v. **23, no. 16,** p. **2029-2032.**