Hydrology and Chemistry of Thermal Waters Near Wells, Nevada

by Paul W. Jewell, Thomas A. Rahn, and John R. Bowman[a]

Abstract

Anomalously warm water in shallow wells and surface springs is found in and around the town of Wells, Nevada. The geology of the Wells area is dominated by Paleozoic and Miocene sedimentary rocks which are cut by high angle normal faults. Thermal springs are associated with the normal faults, whereas thermal water from wells is the result of upward vertical fluid infiltration through the porous, Miocene lacustrine sedimentary rocks and Quaternary alluvium. Temperature logs from a nearby oil exploration well show locally high geothermal gradients (> 50° C/km). Analysis of static water heads indicates that vertical flow is greater than or equal to horizontal flow in low-lying areas near the town of Wells. Solution of the advection-diffusion equation using temperature data from the upper portion of the oil well suggests upward flow velocities of approximately 2-10 m/yr. Cold waters have a mixed cation-bicarbonate chemistry whereas thermal waters are dominantly (Na + K)-bicarbonate. Chemical geothermometers indicate fluid circulation to depths between 1-2 km. Deuterium and oxygen-18 analyses have considerable variability, with modest deviations from the global meteoric water line. The occurrence of deuterium as light as −145‰ suggests significant recharge to the hydrothermal system from very high elevations or water which has had a long subsurface residence time and is recording the signal of a cooler climate.

Introduction

Geothermal waters constitute a small but important portion of the world's energy resources. Thermal waters hot enough to generate electricity form the most economically viable resources, yet these systems are relatively rare. Moderate- to low-temperature resources are much more abundant, yet their use is constrained by geographic and economic factors which often renders them unusable. Furthermore, their manifestations are often less obvious than higher temperature resources.

The Basin and Range physiographic province of the western United States is characterized by relatively thin crust, high heat flow, and high-angle faulting. These features combine to produce a variety of geothermal waters, the majority of which are related to the circulation of meteoric water through high-angle faults (Garside and Schilling, 1979; Ward et al., 1979). A zone of anomalously high heat flow in the north-central portion of the Basin and Range province is known as the Battle Mountain heat flow high (Sass et al., 1971). A large number of moderate- to low-temperature geothermal resources are associated with this thermal feature.

One of the geothermal resources within the Battle Mountain heat flow high is located near the town of Wells, Nevada. Warm springs associated with Basin and Range faults north of Wells have been known for decades (Garside and Schilling, 1979). Water wells drilled for domestic and municipal water supplies within the town were also known to contain anomalously warm water (K. Taylor, former mayor of Wells, pers. comm.), although the potential energy applications of the warm water were not appreciated until the increased petroleum prices of the 1970s. General geological and geochemical surveys of this resource were conducted in the early 1980s. By the mid-1980s, low-temperature geothermal waters from wells within the town were being used for a district space heating project.

The extent and resource potential of the Wells geothermal system remained unappreciated for many years, largely because many of the thermal features were obscured by cold, shallow ground water. The present study was undertaken to characterize the nature, sources, and movements of the geothermal waters near Wells using geological information available in the public domain as well as standard geological and geochemical techniques. The goal of the study was a conceptual model of the Wells system which might serve as a guide for future, low-cost exploration and evaluation of low-temperature geothermal resources in the Basin and Range province.

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Methodology

Surface mapping of the geologic units and structure was conducted on 1:24,000 and 1:62,500 aerial photographs. Cuttings from an oil exploration well (the Dalton #1 well) drilled in 1973 immediately north of Wells were logged and borehole geophysical logs from the well obtained. All available water well records on file in the Nevada State Engineer's Office within approximately 7 km of Wells were examined and synthesized. Twenty-one springs, thermal wells, and selected nonthermal wells near the town of Wells were selected and sampled in November 1991. Well waters were sampled from pumps at the wellhead. All samples were filtered and a 100 ml split of the water sample acidified with HNO_3 for cation and silica analysis. Field measurements included temperture, pH, and conductivity. Alkalinity in the laboratory was measured using the titration technique of Fishman and Friedman (1989). Sulfate and chloride were measured with ion chromatography, and cations were measured with atomic absorption or inductively coupled plasma spectroscopy. Stable isotope analyses were performed using CO_2/H_2O equilibration for oxygen-18, and reduction of H_2O with zinc for deuterium.

Hydrogeologic Setting

Geologic Units

The geology around Wells, Nevada is typical of much of the eastern Basin and Range province (Coats, 1987). The oldest rocks are mapped in the metamorphic core complex of the East Humboldt Range (Figure 1). Archean gneisses and Proterozoic schists and quartzites are found within the core of the range along with metamorphosed Paleozoic sedimentary rocks (Snoke, 1992). A variety of unmetamorphosed Paleozoic rocks are found throughout the area (Figure 1). These include the massively bedded, fossiliferous Devonian Guillmette Formation and Pennsylvanian Ely Limestone, and the Permian Edna Formation, a quartz arenite. These rocks occur as erosional remnants in the Wood Hills southeast of Wells, and as portions of fault-bounded horsts in the East Humboldt Range, southwest of Wells (Figure 2).

A small stock of quartz-porphyry rhyolite occurs north of Wells, and scattered outcrops of this unit are found throughout the Wells area (Figure 2). The rhyolite is coarsely crystalline with abundant K-feldspar and quartz phenocrysts. Previous geologic mapping near Wells classifies this unit as the Jarbidge Rhyolite, which has been dated as 15.4 and 16.8 m.y. elsewhere in northeastern Nevada (Coats, 1987).

Lacustrine rocks of the Miocene Humboldt Formation are found throughout the Wells area and much of northeastern Nevada (Figures 1, 2). Exposures consist of poor to moderately lithified siltstones, sandstones, and conglomerates, with lesser amounts of calcareous siltstones and vitric tuffs. Outcrops have a distinctive grayish-white color. When it is adjacent to normal faults north of Wells, the Humboldt is well-lithified, indurated with iron oxides, and displays localized, but distinctive leisegang banding. The Humboldt Formation is reported to be 10 to 15 m.y. old (Stewart, 1980). A minimum thickness of approximately 1 km for the Humboldt Formation can be established from the oil exploration well immediately north of the town (Figure 3). This roughly corresponds with the maximum thickness of the Humboldt Formation described to the west (Smith and Ketner, 1976).

Two units of unconsolidated Quaternary alluvium have been mapped in the Wells area (Figure 2). The first consists of alluvial gravel deposits typical of those found in Basin and Range grabens. These alluvial deposits are relatively thin near the town of Wells, where exposures in gravel pits show only 1 to 5 m of alluvium overlying the Humboldt Formation. In the southern portion of the study area, the alluvial deposits appear to be much more extensive, although exact thicknesses are not known. The second mappable Quaternary unit consists of fluvial gravels and sands associated with the headwaters and streams of the modern-day Humboldt River (Figure 2).

Subsurface Lithology

The Dalton #1 oil exploration well (Figures 2, 3) provided an opportunity to obtain detailed subsurface geologic and temperature data. The lithologic log shows approximately 1 km of Humboldt Formation overlying an unknown thickness of Paleozoic siliciclastic rocks. The same siltstone, sandstone, and conglomerate lithologies observed in surface

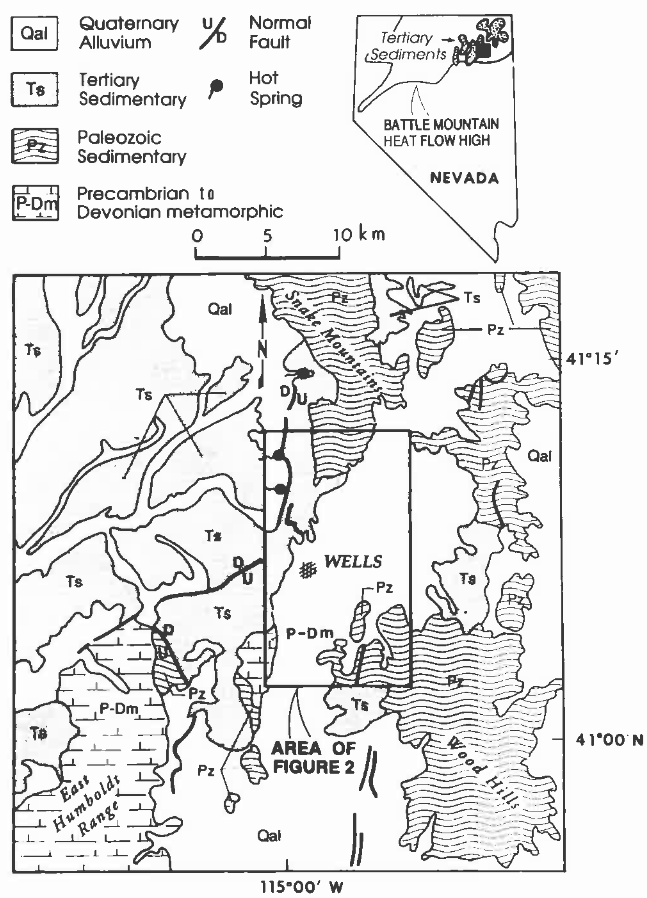

Fig. 1. Regional geology of northeastern Nevada near the town of Wells. Modified from Coats (1987). Also shown is the extent of the Battle Mountain heat flow high (Sass et al., 1971) and the Humboldt Formation (from Stewart, 1980).

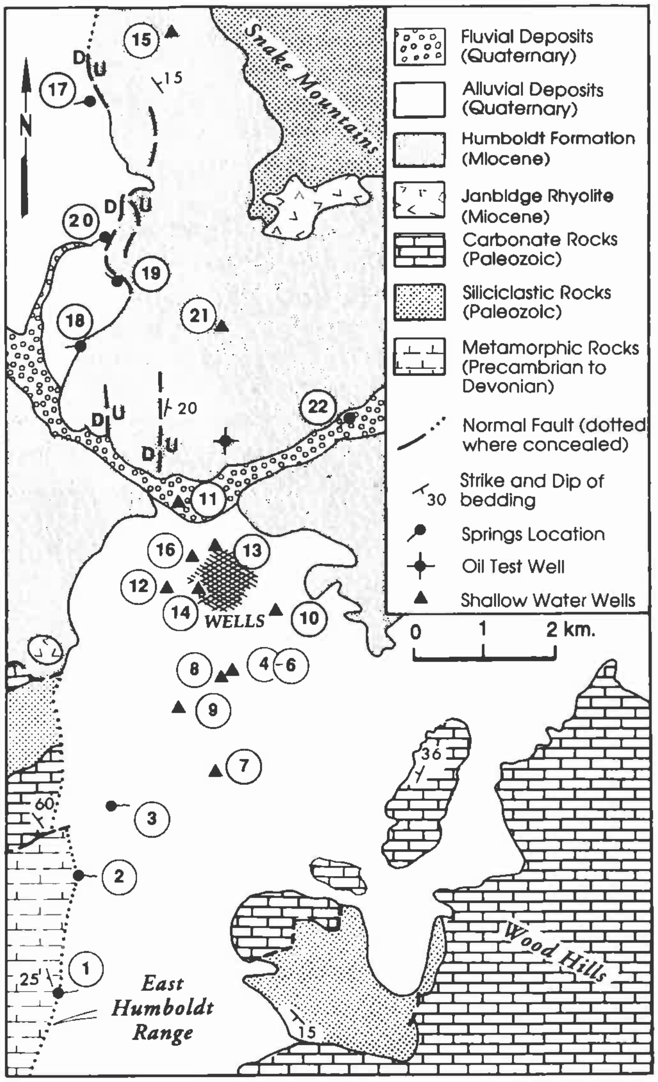

Fig. 2. Detailed geologic map and sample numbers of geothermal features discussed in the text near Wells, Nevada area (unpublished mapping by P. W. Jewell with modifications from Coats, 1987 and Snoke, 1992).

outcrop of the Humboldt Formation are seen in the lithologic log. Abundant clasts of Paleozoic sedimentary rocks in the upper portion of the well support the idea that the Humboldt Formation records the progressive unroofing of the Ruby Mountains-East Humboldt Range during the Miocene (Smith and Ketner, 1976). The porosity of the Humboldt Formation ranges from 15-30% and shows a general decrease with depth (Figure 3). Porosity of the Paleozoic rocks is variable but generally less than that of the overlying Humboldt Formation.

Geologic Structure

Basin and Range normal faulting dominates the structural geology of the Wells area, although Mesozoic compressional tectonics and subsequent low-angle faulting are known to have occurred in East Humboldt Range, Wood Hills, and intervening areas (Snoke, 1992). A set of well-defined normal faults occurs along the western edge of Snake Mountains and appear to control discharge of the associated springs (Figure 2). The Humboldt Formation dips to the east or northeast near the range-bounding fault. A pair of parallel normal faults in the Humboldt Formation immediately north of town are expressed as north-south drainages with minor silicification and alteration of the volcaniclastic sedimentary rocks. These normal faults become less pronounced to the east and south, and within the town of Wells, the Humboldt Formation is essentially flat-lying.

South of the town of Wells, the eastern boundary of the East Humboldt Range is marked by a concealed normal fault. This fault appears to control the flow of several low-temperature (maximum temperature of 22° C) springs. The western edge of the Wood Hills to the southeast of Wells is a pediment with no surface expression of faulting. Dips of Paleozoic sedimentary rocks south of Wells are dominantly to the west, in contrast to the eastward-dipping Humboldt Formation north of the town.

It is interesting to note that the surface expression of west-bounding normal faults in the Snake Mountains and the east-bounding normal fault in the East Humboldt Range form a linear trend. The location of the warm wells drilled in Wells (i.e., well numbers 6, 11, 14, and 16 in Figure

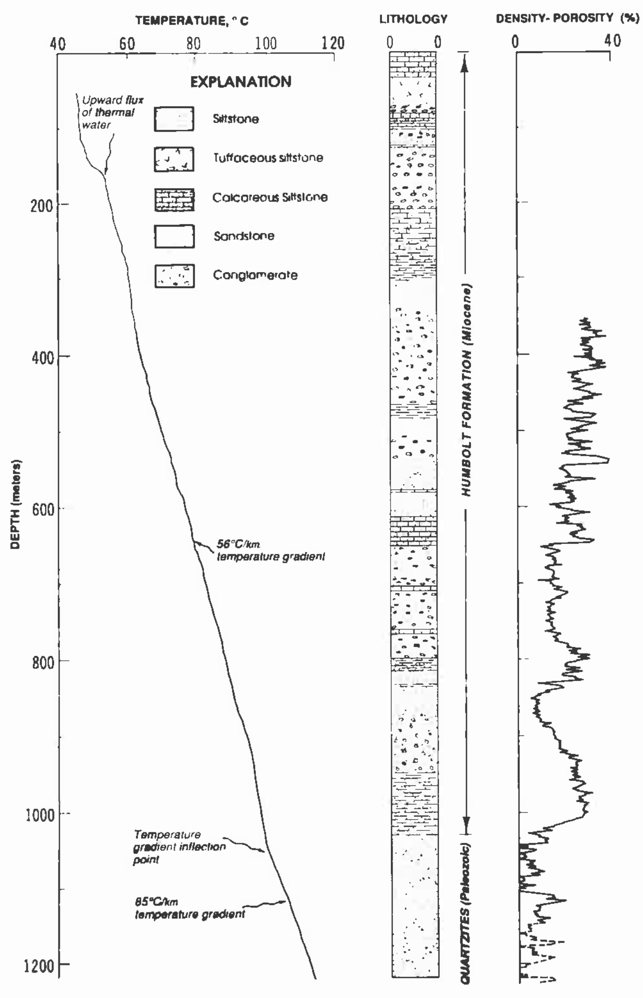

Fig. 3. Downhole temperature, lithologic, and density-porosity logs of the oil and gas test well located north of the town of Wells, Nevada.

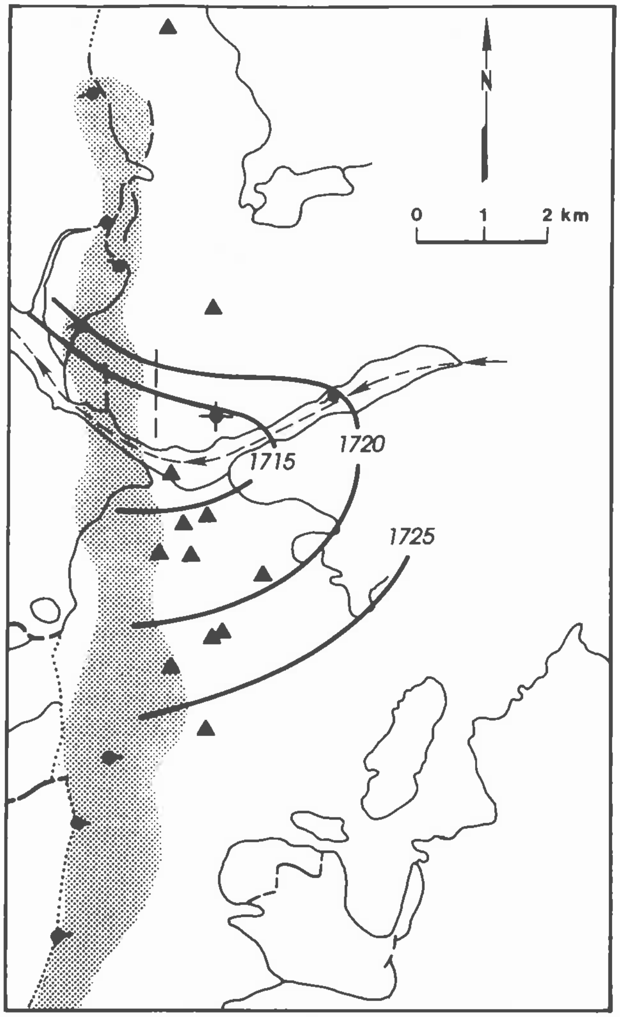

Fig. 4. Approximate potentiometric surface of shallow water wells drilled between 1975 and 1985 in the Wells area in meters above sea level. Outlines of the geologic units and sample locations shown in Figure 2 are reproduced for spatial reference. Stippled zone represents hypothesized structural connection between the normal faults on the west side of the Snake Mountains and the east side of the East Humboldt Range. Dashed line shows the Humboldt Wells drainage (headwaters of the Humboldt River).

2; Figure 4) lies slightly to the east of this linear trend. The nature of the subsurface structure between the Snake Mountains and East Humboldt Range is largely unknown. A major west-northwest trending, right lateral strike-slip fault of Mesozoic age, has been mapped though this area and named the Wells fault (Thorman and Ketner, 1979). The existence of the Wells fault is disputed by Stevens (1981) and Coats (1987). The offset of the north-south trending mountain ranges in this area implied some sort of subsurface structural discontinuity, however.

Shallow Ground Water

Cold (< 15°C), shallow ground water is pervasive in the Miocene Humboldt Formation and Quaternary gravels of the low-lying areas near Wells. Humboldt Wells, a group of springs which mark the headwaters of the Humboldt River, are found immediately north of town.

Approximate hydraulic gradients of the shallow unconfined aquifer in the Humboldt Formation near Wells were determined from public water well records. Static water level data from these types of records must be carefully examined for internal consistency before they are employed in hydraulic gradient calculations. For instance, examination of well records for this study showed a progressive lowering of the water table (approximately 1-3 m) at a given location between World War II and the present. On the other hand, water well records for a shorter period of time (several years) showed a consistent spatial relationship between static water level and topography, i.e., static water level was progressively deeper in topographically higher areas on either side of the Humboldt Wells drainage (Figure 4). For wells drilled between 1975 and 1985, a plot of static, shallow ground-water levels indicated hydraulic gradients of approximately .005 in the alluvial valley south of town, and values two to three times higher in the steeper topography associated with uplifted portions of the Humboldt Formation north of town (Figure 4).

Thermal Features

Thermal features of the Wells geothermal system consist of hot springs and thermal wells. Measured temperatures of wells and springs range from 8° to 60°C. The warmest waters emanate from the fault-controlled hot springs north of town. Water being used in the district space heating project is approximately 25-35° C. Springs and shallow water wells south of Wells have temperatures below 25°C.

A temperature log of the oil well shows several thermal regimes (Figure 3). As with most oil well temperature logs, the lower portion of the temperature profile has probably been depressed and the upper portion raised by the circulation of drilling fluids (Jessop, 1990). As such, the relative features of the temperature profile can be examined, while absolute temperatures must be observed with some caution.

Although there is no temperature log from 0-50 m (Figure 3), extrapolation of the upper portion of the log results in a surface temperature intersection of approximately 30° C. As suggested above, this is probably an artifact of the drilling fluid circulation. Confirmation comes from temperature-depth data of ground-water wells near the town of Wells which show significantly lower bottom hole temperatures than those recorded in the oil exploration well (Figure 5).

At approximately 150 m depth, the temperature log shows an abrupt shift. This shift probably represents mixing of warm thermal waters with cooler, near-surface waters. The remainder of the measured temperature gradient through the Humboldt Formation is approximately 56°C/km. A subtle thermal transition corresponds to the Humboldt-Paleozoic contact where the measured gradient becomes significantly higher (> 80°C/km) (Figure 3).

Regional heat flow is the product of the thermal conductivity of the rocks and the vertical temperature gradient. Since the regional heat flow is assumed to be constant in the area near Wells, differences in observed temperature gradients between the Humboldt Formation and the

Paleozoic sandstones should be the result of differences in thermal conductivity. The thermal conductivity of porous, water-saturated siliciclastic rocks (such as the Humboldt Formation) is less than the thermal conductivity of relatively impermeable, well-cemented sandstones (such as the Paleozoic rocks logged in Figure 3) (e.g., Jessop, 1990). Thus, the observed vertical temperature gradients in the porous Humboldt Formation should be *higher* than those observed in the Paleozoic rocks. The reverse is observed in the oil well near Wells (Figure 3). One possible explanation for the lower vertical temperature gradient in the Humboldt Formation is significant vertical flow (discussed in more detail below) in this relatively porous unit. Evidence for vertical flow in the upper portion of the Humboldt Formation can be found in public water well records which show artesian conditions in deep (> 200 m) wells and nonartesian conditions in shallow (< 50 m) wells.

Estimates of Fluid Velocity

As shown above, horizontal head gradients in the shallow, unconfined aquifer near the town of Wells range from .005 and .015 (Figure 4). A rough estimate of vertical head gradients near Wells was obtained from records of deep artesian (> 200 m) and shallow nonartesian (< 50 m) water wells near the town. While recognizing that reported heads of the artesian wells in public well records are only approximate, vertical head gradients of 0.11 to 0.19 were calculated from the data of adjacent artesian and nonartesian wells. This is approximately 7-35 times higher than the horizontal head gradients of the unconfined aquifer (Figure 4).

For a unit area, estimates of horizontal fluid flow can be obtained from Darcy's Law:

$$q = -K_l \frac{\partial h}{\partial l} \tag{1}$$

where q is horizontal specific discharge (m/s); K_l is horizontal hydraulic conductivity (m/s); and h is observed head (m). An expression for vertical flow is (e.g., Domenico and Schwartz, 1990):

$$q_z = -\frac{k_z}{\mu}\left[\rho_o g \frac{\partial h}{\partial z} - (\rho_f - \rho_o)\, g\right] \tag{2}$$

In this case, q_z and k_z are vertical specific discharge (m/s) and permeability (m^2), respectively; ρ_o is the reference density (kg/m^3); ρ_f is the fluid density; g is the gravitational constant (m/s^2); and μ is the dynamic fluid viscosity (kg/m/s). The two terms on the right-hand side of equation (2) represent vertical flow due to pressure gradients and buoyancy. Density differences due to the observed temperatures at less than 200 m depth are small (less than 1%; Sorey, 1978). Equation (2) can therefore be written as:

$$q_z = -\frac{k_z \rho_o g}{\mu}\frac{\partial h}{\partial z} = -K_z \frac{\partial h}{\partial z} \tag{3}$$

Although the Humboldt Formation exhibits a variety of siliciclastic lithologies (Figure 3), most outcrops observed in the field have a modestly anisotropic fabric. If the ratio of anisotropy, K_l/K_z, is of the same order as the ratio of horizontal to vertical head gradients (7-35), then vertical flow rates in the topographically low area near the town of Wells are the same order of magnitude as horizontal flow rates. If the siltstones are isotropic, then the vertical flow is significantly greater than horizontal flow.

An advection-diffusion analysis of the oil well temperature log between depths of 100-200 m (Figure 6) allows an estimate of vertical flow velocity to be calculated. The vertical, steady-state governing equation for temperature is:

$$\left[\frac{n\kappa_f + (1-n)\,\kappa_s}{n\rho_f c_f}\right]\frac{\partial^2 T}{\partial z^2} = v_z \frac{\partial T}{\partial z} \tag{4}$$

where n is porosity; κ_f and κ_s are the thermal conductivities (W/m/°C) of the fluid and solid respectively; c_f is the specific heat (J/kg/°C^{-1}) of the fluid; and ρ_f is fluid density (kg/m^3). Equation (4) can be simplified by combining the coefficient on the left-hand side into an effective thermal dispersivity, α (m^2/s). It is difficult to calculate an exact value for α because the thermal conductivity of the fluid, κ_f, is dependent on the dispersivity characteristics of the aquifer and the velocity of fluid. Compilations (e.g., Marsily, 1986, Table 10.3; Domenico and Schwartz, 1990, Table 9.3) suggest average effective thermal dispersivities between 2-10 $\times$ 10^{-7} m^2/s.

The shape of the steady-state thermal profile described by equation (4) is dependent on the relative magnitude of vertical velocity, v_z, and thermal dispersivity, α. An estimate of vertical velocity can be determined by calculating a thermal Peclet number:

$$\beta = v_z L/\alpha \tag{5}$$

Bredehoeft and Papadopulos (1965) present values of β which correspond to solutions of equation (4). A match of the expanded upper portion of the oil well temperature log (Figure 6) and the curves of Bredehoeft and Papadopulos (1965, their Figure 2), show β to be approximately −7.

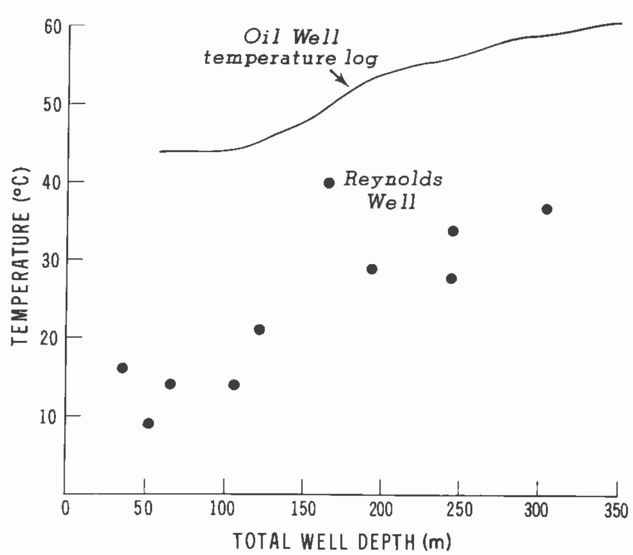

Fig. 5. Bottom hole temperature for shallow water wells (circles) reported from well driller logs and recorded temperature from the Dalton #1 oil exploration well. Shallow water temperatures are from aquifers at or near the bottom of the hole.

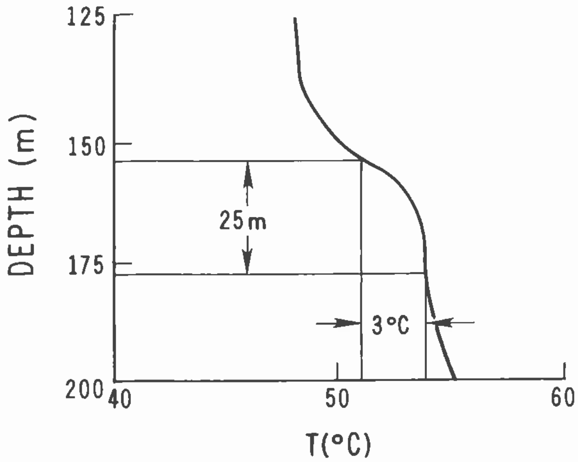

Fig. 6. Details of the upper portion of the temperature log from the oil exploration wells showing the depth and temperature range employed in the advection-diffusion analysis.

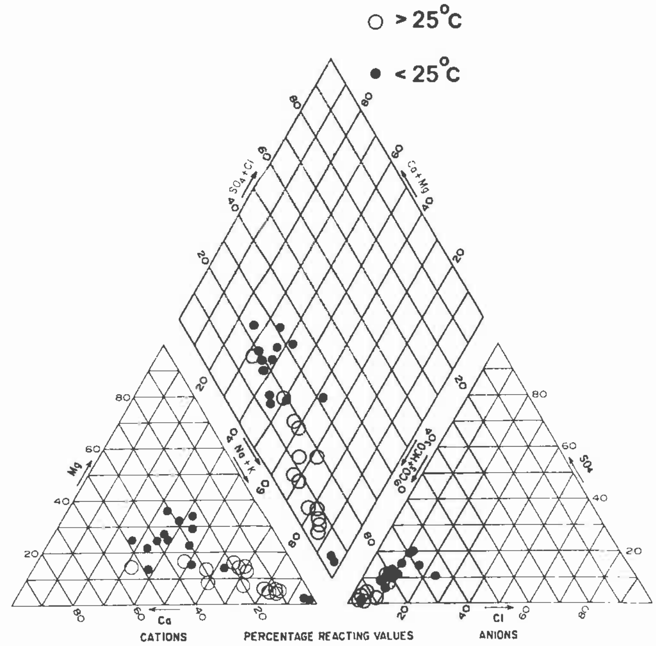

Fig. 7. Piper plot of cations and anions (eq) of Wells waters with measured temperatures below 25° C and measured temperatures above 25°.

Substituting the assumed thermal dispersivities and the vertical distance over which temperature changes (approximately 25 m) (Figure 6) gives a vertical velocity of 0.6-3 $\times$ 10^{-7} m/s (2-10 m/yr). The horizontal fluid velocity would therefore be equal to or somewhat less than this value near the town of Wells and approximately two to three times higher in the nearby mountains (Figure 4).

Geochemistry

Major Elements

Major element analyses of the 21 springs and well waters collected in 1991 were combined with analyses of some of the same springs reported elsewhere (Garside and Schilling, 1979; Jewell, 1982) and plotted in Figure 7. The analyses show distinctive patterns of the thermal and nonthermal waters. In general, waters with measured temperatures less than 25° C have approximately equal amounts of cations, while waters with greater temperatures are dominantly (Na + K) (Figure 7). However, two cold water samples which are dominantly (Na + K) emanate from a fault-controlled, nonthermal spring (#18 in Figure 2). Anions for both thermal and nonthermal waters are dominantly bicarbonate.

Geothermometry

Concentrations of silica and the relative proportions of cations can be used to gain a crude estimate of subsurface temperatures (Table 1). Application of these geothermometers is based on several assumptions: (1) chemical equi-

Table 1. Summary of Chemical Geothermometers for Waters with Surface Temperatures > 25° C in the Wells, Nevada Geothermal System

Sample	*Surface T (° C)*	*Quartz (cond.)*	*Quartz (ad.)*	*Chalcedony*	*Na-K-Ca*	*Na-K-Ca (Mg-corr.)*
6	37*	n.d.	n.d.	n.d.	102	—
11a	40*	130	127	103	119	97
11b	40*	n.d.	n.d.	n.d.	121	78
14	29*	115	114	86	106	81
14b	29*	n.d.	n.d.	n.d.	102	70
16	34*	n.d.	n.d.	n.d.	111	78
17a	61	140	135	114	89	—
17b	55	129	126	101	101	78
17c	46	n.d.	n.d.	n.d.	57	—
19a	60	143	137	116	88	—
19b	45	n.d.	n.d.	n.d.	92	82
20a	36	122	120	94	94	78
20b	n.d.	105	106	76	97	84
20c	35	n.d.	n.d.	n.d.	90	81

*Reported temperature from well drilling logs.
n.d.—not determined.
—Indicates that the calculated Mg-corrected Na-K-Ca geothermometers is below 70° C and therefore not valid (Fournier and Potter, 1979).

librium is established between the waters and the surrounding rock; (2) there is an adequate supply of chemical reactants; and (3) there is no chemical reequilibration during ascent of the hot water (Fournier et al., 1974).

Silica geothermometers rely on the premise that dissolution of various quartz phases are temperature dependent. The quartz conductive geothermometer assumes quartz equilibration with conductive heat loss; the quartz adiabatic geothermometer assumes no heat loss. The chalcedony geothermometer is based on equilibration with amorphous silica. This geothermometer is probably the most appropriate for the Wells geothermal system because the Humboldt Formation contains abundant volcaniclastic glass. Temperatures determined by the chalcedony geothermometers for waters with surface temperature > 25°C range from 76-116°C (Table 1).

Cation geothermometers assume equilibration of thermal fluids with aluminosilicate minerals (Fournier and Truesdell, 1973). A correction for the effects of magnesium is typically considered to be necessary (Fournier and Potter, 1979). Calculated Na-K-Ca geothermometers for waters > 25°C range from 57-121°C and Mg-corrected geothermometers range from 70-97°C. The latter values correspond relatively well to calculated chalcedony geothermometers.

The calculated geothermometers in conjunction with temperature data from the oil test well indicate that fluid circulation is relatively shallow (1-2 km). If so, then rock-water interaction would be restricted largely to the lacustrine Humboldt Formation. This appears to be confirmed by major element chemistry (Figure 7). Thermal waters have relatively higher amounts of Na and K, two elements which would be produced from dissolution of volcanic glasses which are common in the Humboldt Formation.

Stable Isotopes

Hydrogen and oxygen isotopic analyses of spring and well waters, and two streams fed by snowmelt runoff are illustrated in Figure 8. The cold spring and well waters have δD values ranging from −125 to −140‰ and $\delta^{18}O$ values from −15.9 to −17.1‰. These samples (crosses in Figure 8) plot slightly to the right of the global meteoric water line. The variability in δD values of the cold waters suggests local, elevation-controlled variations in δD values of recharge water, although we have not attempted direct sampling of precipitation as a function of elevation to confirm such a relationship. The differences in both δD and $\delta^{18}O$ values relative to the global meteoric water line may result from local conditions of precipitation or from minor evaporative effects or both. The isotopic variability of these cold waters and their modest displacement from the global meteoric water line is typical of cold waters in other Basin and Range ground-water systems (Young and Lewis, 1982; Jacobson et al., 1983).

The hydrogen and oxygen isotopic compositions of thermal waters range from −137 to −147 and −15.25 to −18.0‰, respectively. For any given δD value, the thermal waters are somewhat more enriched in ^{18}O compared to the cold waters. This is an expectable result of isotopic exchange with rock and/or sediments at elevated temperatures. Generally, the thermal waters are variably depleted in deuterium compared to most of the cold waters. Thus, the thermal and cold waters probably do not have a common recharge area. The lower δD values of the thermal water suggest that its recharge area(s) probably lies at somewhat higher elevation than those of the cold waters. Elevations in the East Humboldt Range are high enough (+3000 m) that sufficiently deuterium-depleted precipitation is possible. It is also possible that the thermal waters are derived from older glacial—and deuterium-depleted—waters (e.g. Rozanski, 1985). This alternative cannot be tested directly without dating the water with techniques such as ^{14}C.

Two samples of streams fed by snowmelt were collected in the late Spring of 1992 at an elevation of approximately 2500 m. Their δD values are −111 and −114‰, which are significantly enriched in deuterium compared to both local cold waters and the thermal waters in Wells Valley. These heavy δD values might reflect significant deuterium enrichment in the residual snowpack as a result of isotopic fractionation accompanying progressive melting of the snowpack. More likely, these δD values are representative of precipitation at this elevation. If so, then recharge of both the cold and thermal waters would have to be at still higher elevation. Analyses of high elevation spring water and precipitation in the Basin and Range record δD values as low as −145 to −150‰ (Jacobson et al., 1983), which would cover the range of measured δD values in both cold and thermal waters from the Wells area.

Discussion and Conclusions

A generalized picture of the Wells geothermal system emerges from the combined hydrologic, lithologic, and geochemical data. Recharge to the system probably occurs at a variety of elevations in the surrounding mountain ranges (Figures 1, 9). Fluid pathways appear to be diverse, as

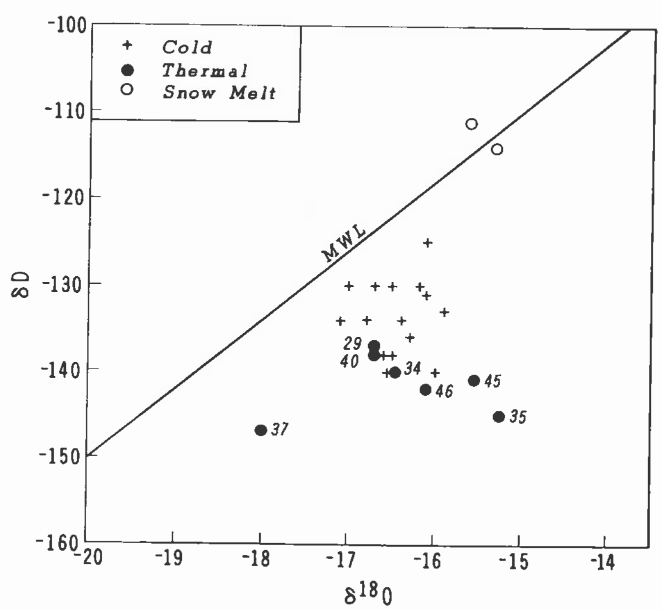

Fig. 8. Oxygen-18—deuterium plots for thermal and nonthermal waters. Numbers refer to measured temperatures.

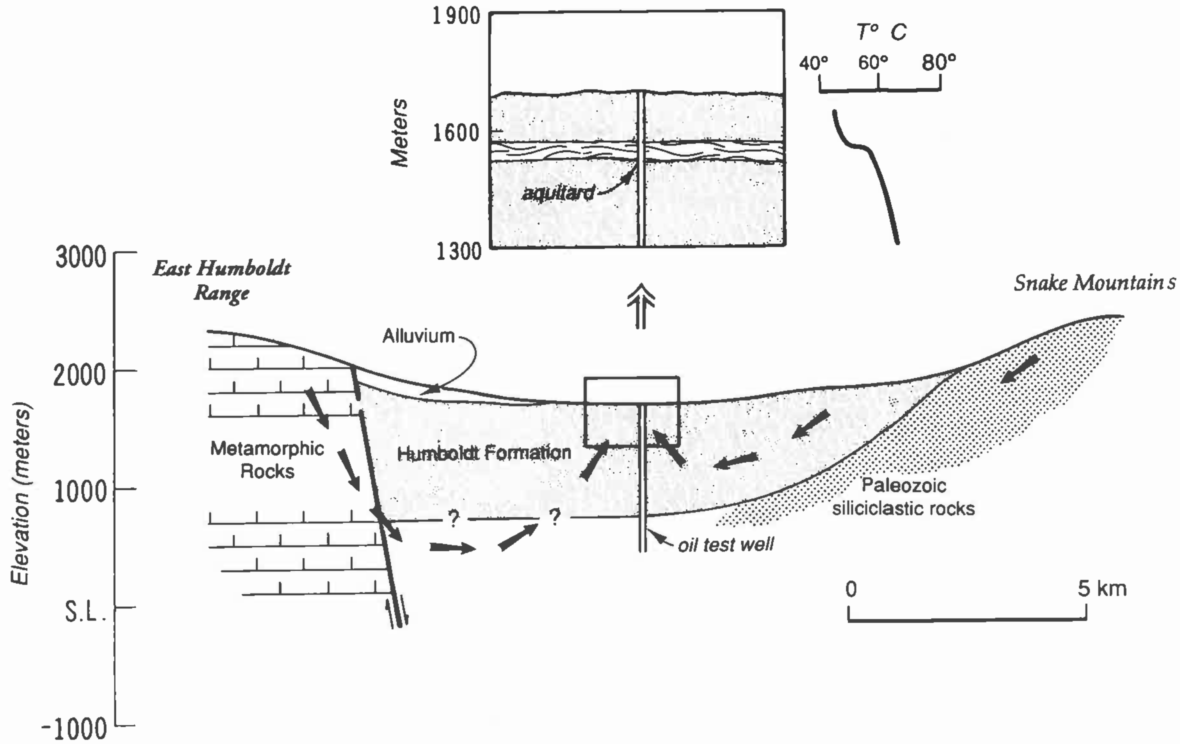

Fig. 9. Diagrammatic cross section of the Wells geothermal system. Recharge to the system comes from the surrounding East Humboldt Range and Snake Mountains. Water circulation is to depths of 1 to 2 km. Vertical flow is pronounced in the low-lying areas around the town of Wells. Upward flow is constrained by an aquitard in the upper portion of Humboldt Formation, thereby causing an abrupt change in the upper portion of the temperature profile (see inset).

evidenced by the variable light stable isotope values and chemical geothermometers. Maximum depth of circulation as predicted by chemical geothermometers is approximately 1-2 km. Thermal waters appear to have significant interaction with the volcaniclastic-rich, Miocene Humboldt Formation. Near the town of Wells, the thermal waters mix with shallow, cool ground water. Thermal waters at these relatively low elevations exhibit significant upward flow relative to horizontal flow. Upward migration of thermal water near Wells may be facilitated by a subsurface fault (possibly the east-west strike-slip Wells fault) which connects inferred normal faulting on the eastern side of the East Humboldt Range with observed normal faults on the western side of the Snake Mountains (Figure 2). No surface manifestation of this fault is observed in the Humboldt Formation or Quaternary alluvium near Wells, however (Figure 2). Portions of the Humboldt Formation near Wells appear to be acting as an aquitard to the upward migration of water in the topographically low area near Wells (Figure 9). Evidence for the aquitard includes the artesian nature of wells drilled between depths of 200 to 300 m and a sharp inflection in the oil exploration temperature log between 100 and 200 m (Figure 3).

Although exact recharge areas for the thermal waters cannot be pinpointed, recharge is presumably from the East Humboldt Range, Snake Mountains, and Wood Hills (Figure 1). The highest portions of these ranges are approximately 10 km away and represent the maximum horizontal travel distance for the fluids which emerge in the Wells area. Since the depth of water circulation is believed to be relatively shallow (1-2 km), the total travel distance for the water is probably < 15 km. Calculated horizontal velocities near the town of Wells are between 2-10 m/year. Maximum ground-water residence times at these velocities are 1500-7500 years, and the very light stable isotopes (Figure 8) would not represent a signal from the cooler climates which existed 12,000 years and greater before the present. Much smaller average fluid velocities (approximately 1 m/yr) would be necessary for the waters presently being discharged from Wells to have entered the regional hydrologic system during the last global glacial maximum.

The potential of the Wells geothermal system remained largely unappreciated until relatively deep ground-water wells (+ 100 m) were drilled into confined aquifers at relatively low elevations during the 1970s and 1980s. Systematic study of the Wells geothermal system suggests that similar systems might be commonly associated with other areas of the Battle Mountain heat flow high. The Humboldt Formation is present in much of the topographically low areas of northeastern Nevada (Figure 1). The anisotropic nature of this formation as well as cold, shallow ground water may be masking undiscovered geothermal resources within the Battle Mountain heat flow high.

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References

Bredehoeft, J. D. and I. S. Papadopulos. 1965. Rates of vertical groundwater movement estimated from the Earth's thermal profile. Water Resources Research. v. 1, pp. 325-328.

Coats, R. R. 1987. Geology of Elko County, Nevada. Nevada Bureau of Mines and Geology Bulletin 101. 112 pp.

Domenico, P. A. and F. W. Schwartz. 1990. Physical and Chemical Hydrogeology. Wiley. 823 pp.

Fishman, M. J. and L. C. Friedman, eds. 1989. Methods for determination of substances in water and fluvial sediments. Techniques of Water Resources Investigations of the U.S. Geological Survey. Book 5, Laboratory analyses, Chapter A1.

Fournier, R. O. and R. W. Potter. 1979. Magnesium correction to the Na-K-Ca chemical geothermometer. Geochimica et Cosmochimica Acta. v. 43, pp. 1543-1550.

Fournier, R. O. and A. H. Truesdell. 1973. An empirical Na-K-Ca geothermometer for natural waters. Geochimica et Cosmochimica Acta. v. 37, pp. 1259-1275.

Fournier, R. O., D. E. White, and A. H. Truesdell. 1974. Geochemical indicators of subsurface temperature. 1. Basic assumptions. U.S. Geological Survey Journal of Research. v. 2, pp. 259-262.

Garside, L. J. and J. H. Schilling. 1979. Thermal waters of Nevada. Nevada Bureau of Mines and Geology Bulletin 91. 163 pp.

Jacobson, R. L., N. L. Ingraham, and M. E. Campana. 1983. Isotope hydrology of a Basin and Range geothermal system. University of Nevada Desert Research Institute Publication 41087. 18 pp.

Jessop, A. M. 1990. Thermal Geophysics. Elsevier, Amsterdam. 306 pp.

Jewell, P. W. 1982. Geology and geothermal potential north of Wells, Nevada. Univ. of Utah Research Institute, Earth Science Laboratory Open File Report DOE/ID/12079-83. 38 pp.

Marsily, G. de. 1986. Quantitative Hydrogeology. Academic Press, Orlando. 440 pp.

Rozanski, K. 1985. Deuterium and oxygen-18 in European groundwaters—links to atmospheric circulation in the past. Chemical Geology. v. 52, pp. 349-363.

Sass, J. H., A. H. Lachenbruch, R. J. Munroe, G. W. Green, and T. H. Mises, Jr. 1971. Heat flow in the western United States. Journal of Geophysical Research. v. 76, p. 6373.

Sorey, M. L. 1978. Numerical modeling of liquid geothermal systems. U.S. Geological Survey Professional Paper 1044-D. 25 pp.

Smith, J. F., Jr. and K. B. Ketner. 1976. Stratigraphy of post-Paleozoic rocks and summary of resources in the Carlin-Pinon Range area, Nevada. U.S. Geological Survey Professional Paper 867-B. 48 pp.

Snoke, A. W. 1992. Clover Hill, Nevada: Structural link between the Wood Hills and East Humboldt Range. In: J. R. Wilson, ed., Field Guide to Geologic Excursions in Utah and Adjacent Areas of Nevada, Idaho, and Wyoming. Utah Geological Survey Miscellaneous Publication 92-3. pp. 107-122.

Stevens, C. H. 1981. Evaluation of the Wells fault, northeastern Nevada and northwestern Utah. Geology. v. 9, pp. 534-537.

Stewart, J. H. 1980. The geology of Nevada. Nevada Bureau of Mines and Geology Special Publication 4. 136 pp.

Thorman, C. H. and K. B. Ketner. 1979. West-northwest strike-slip faults and other structures in allochthonous rocks in central and eastern Nevada and western Utah. In: G. W. Newman and H. D. Goode, eds., 1979 Basin and Range Symposium, Rocky Mountain Assoc. of Geologists—Utah Geological Assoc. pp. 123-134.

Ward, S. H., H. P. Ross, and D. L. Nielson. 1979. A strategy for exploration for high temperature hydrothermal systems in the Basin and Range Province. Univ. of Utah Research Institute, Earth Science Laboratory Report 22. 42 pp.

Young, H. W. and R. E. Lewis. 1982. Hydrology and geochemistry of thermal ground water in southwestern Idaho and north-central Nevada. U.S. Geological Survey Professional Paper 1044-J. 20 pp.