APPENDIX 10—Estimation of hydrologic properties in the southern Sacramento Mountains

ESTIMATION OF HYDROLOGIC PROPERTIES IN THE SOUTHERN SACRAMENTO MOUNTAINS

BACKGROUND

As the vast majority of the wells and springs in the southern Sacramento Mountains study area are within the Yeso Formation, the hydrologic properties of this unit are of significant interest. The rate of groundwater movement through an aquifer system from the recharge zone to discharge zone depends on the hydraulic gradient and hydrologic properties of the aquifer. The hydrologic properties and terms relevant to these analyses include:

- Hydraulic gradient (i, dimensionless)—the vertical change in the water table elevation over a lateral distance.
- Intrinsic permeability (k, cm^2)—the capacity of a porous rock to transmit a fluid. This is a property of the rock only and in this appendix the term is shortened to permeability.
- Hydraulic conductivity (K, m/s)—the ability of water to move through a porous medium in unit time under a unit hydraulic gradient. This is a property of the rock combined with properties of the fluid.
- Transmissivity (T, m^2/s)—hydraulic conductivity multiplied by the aquifer thickness. This value may be estimated from aquifer tests. The hydraulic conductivity can be derived from transmissivity by assuming an effective aquifer thickness.
- Effective porosity (n_e, %)—the proportion of the total aquifer volume that is occupied by interconnected pores and/or fractures.
- Specific yield (Sy, %)—the ratio of a volume of water that drains by gravity to the total volume of rock, also known as the drainable porosity. It is less than or equal to the effective porosity.
- Specific discharge (q, m/s)—volumetric flow rate per unit cross-sectional area.
- Linear or pore water velocity (v, m/s)—groundwater flow velocity.

The following mathematical relationships are important for the discussion in this section. Specific discharge is proportional to the hydraulic gradient according to Darcy's Law:

$$q = -Ki \qquad (1)$$

The pore water velocity is related to specific discharge by the following equation:

$$v = q/ne \qquad (2)$$

Compiled data on hydrologic properties of the Yeso Formation are presented in Table 10.1. The results presented by Wasiolek and Gross (1983) and Wasiolek (1991) are re-analyses of aquifer test data from the sources listed (details of the original references were not provided). Hydraulic conductivities were calculated from transmissivities by assuming that the screened interval of the well is equal to the aquifer thickness. The wells described by Wasoilek (1991) penetrated all of the various rock types present in the Yeso Formation, thus the hydrologic properties are likely averages over a variety of rock types. Summers (1976) data are derived from aquifer tests on several Town of Cloudcroft wells, and are also averages over many rock types. The specific yield value is a best estimate for the Yeso Formation derived from aquifer tests, and spring and stream flow measurements.

We estimated aquifer properties for the Yeso Formation using a hydrograph separation technique and heat flow modeling and compared the results to previous studies (Table 10.1). Well hydrograph analysis was used to estimate the transmissivity and specific yield for portions of the Yeso Formation. A numerical model of groundwater and heat flow was used to estimate an upper constraint on the hydraulic conductivity of the Yeso Formation along an assumed flow path through a fractured perched aquifer in the high mountains. Using this estimated hydraulic conductivity and a hydraulic gradient derived from the regional potentiometric surface (report Figure 18), we estimated the specific discharge (Equation 1). Tritium values from two springs along this assumed flow path were used to measure a linear flow velocity, which then provided an estimate for effective porosity (Equation 2).

As part of this study, Morse (2010) used carbon-14 (14C) age dates of groundwater along an apparent flow

Table 10.1–Compilation of hydrologic properties for the Yeso Formation in the southern Sacramento Mountains.

Transmissivity (T; m^2/s)	Hydraulic conductivity (K; m/s)	Specific yield (S_y)	Comments
Wasiolek 1991, reanalysis of data from Woodward-Clyde (1978)			
3.80e-06	2.50e-08		Mescalero well E1
1.00e-03	5.30e-06		Mescalero well E6
4.90e-04	2.10e-06		Mescalero well E3
6.00e-04	2.80e-06		Mescalero well E10
2.00e-05	1.30e-07		Mescalero piezometer E1
	1.0e-07		best estimate, unfractured siltstones
	1.0e-04		best estimate, fractured limestones
	5.0e-06		best estimate, formation average
Summers 1976, aquifer tests on Cloudcroft town wells			
4.32e-04	1.42e-06	0.0015	best Sy estimate for bulk Yeso
Wasiolek 1983, reanalysis of data from Summers 1978, Hood 1960			
5.33e-04			Summers, step drawdown test, bulk Yeso
1.73e-04			Summers, const discharge test, bulk Yeso
8.65e-04			Hood, specific capacity test
4.89e-04			Hood, specific capacity test
This study, separation of two recessions in well SM-0049 hydrograph			
		.00504 to .0178	unfractured limestone and/or siltstone/mudstone – “matrix”
8.2e-5 to 4.4e-4	4.9e-6 to 7.5e-5	.0015	fractured limestone + (some siltstone/mudstone/limestone matrix); S_y is bulk Yeso of Summers 1976 – “fractures + matrix”
		2.06e-4 to 4.25e-4	fractured limestone – “fractures”
This study, numerical modeling of fluid and heat transport			
	1e-5		Maximum value for fractured lime stone aquifer

path on the Pecos Slope to estimate a groundwater flow velocity and the hydraulic conductivity for the San Andres Formation.

YESO FORMATION HYDROLOGIC PROPERTIES

Hydrograph analysis

We applied the method of Shevenell (1996) to the continuous hydrograph from well SM-0049 (Figures 10.1, 10.2, and 10.3) to derive hydrologic properties for different rock types in the lithologically heterogeneous Yeso Formation. The method, derived for karstic and/or fractured carbonate aquifers, was applied to the SM-0049 hydrograph because it has several sharp water level rises and subsequent recessions that can be correlated with rainfall events. The recessions consist of three distinct segments with different slopes. Shevenell (1996) identified similar sharp rises and recessions composed of three distinct slopes in karst hydrographs and associated them with three portions of a karst aquifer. In her model, from steepest to shallowest, the slopes are associated with drainage of dissolution-enlarged conduits, fractures and carbonate matrix. In our application to the Yeso Formation, we associate the slopes with, from steepest to shallowest: 1) densely fractured limestone beds and/or karst collapse breccias (“fractures”); 2) less-fractured limestones and/or relatively porous sandstone intervals (“fractures and matrix”); and 3) limestone matrix and/or siltstone and mudstone intervals of relatively low permeability (“matrix”).

The method assumes that the different hydrograph slopes represent head changes with time due to changes in discharge and storage in the different aquifer divisions. This assumption was verified by Powers and Shevenell (2000) using the Bernoulli equation. To utilize the method, one calculates the ratios of recession slopes for the three aquifer portions, from which ratios of the specific yields associated with the three aquifer portions can be derived. If one of the specific yields is known, the other two can then be calculated. We applied Summers’s (1976) estimate of 0.0015 for the specific yield of the bulk Yeso Formation, which contains a mix of high and low permeability materials, to the aquifer portion associated with the intermediate hydrograph slope associated with fractures and matrix. Following Shevenell (1996), we then derived transmissivity values for this portion of the aquifer, and subsequently a range of hydraulic conductivity values. The method only yields one transmissivity value, which must be assumed constant over all aquifer portions. The aquifer thickness for well SM-0049 is unknown. To determine hydraulic conductivity, we used a range of values, from the height of the water level rise (~20 feet) to the distance from the well bottom to the mapped groundwater surface (~55 feet). Results agree favorably with those derived in other studies using traditional aquifer testing methods (Table 10.1). This lends confidence to the calculated specific yields, which we use subsequently in calculations of recharge.

Groundwater and heat flow modeling

In addition to the hydrograph analysis described above, we used a mathematical model of groundwater and heat flow to constrain hydraulic conductivity in the Yeso Formation. We used the field parameters

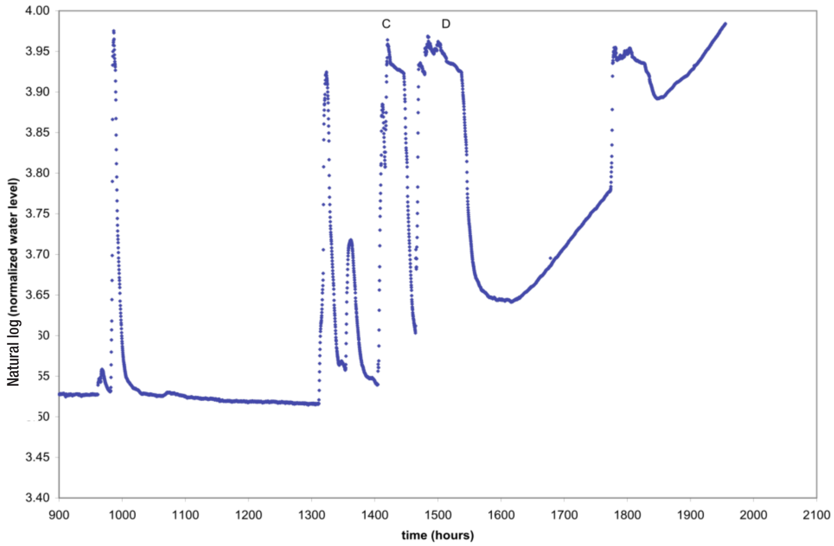


Figure 10.1–Portion of the hydrograph of well SM-0049, showing two water level rises "C" and "D" that were used to determine hydrologic properties and recharge. Water level rise "C" occurred from August 19 to 20, 2006; and water level rise "D" occurred from August 21 to 24, 2006.

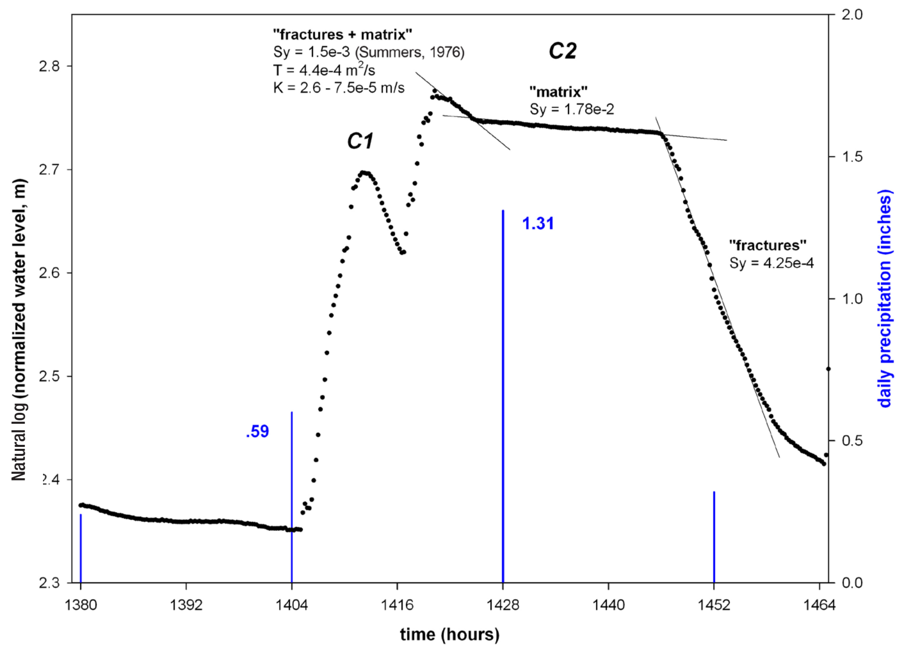


Figure 10.2–Expanded view of water level rise C from well SM-0049 hydrograph. The two individual rises are labeled C1 and C2, which occurred from August 19 to 20, 2006. Straight line segments are labeled with hydrologic properties calculated from them. Precipitation data are daily totals from NOAA and CoCoRHAS stations.

of specific conductance (SC) and temperature in springs to identify a possible flow path within a localized perched aquifer. Water temperatures in the springs of interest are slightly warmer than mean annual air temperature estimated for the elevation of the springs. Recharge to the aquifer associated with these springs likely took place at higher elevations, where the mean annual temperature is cooler than that of the spring locations. As indicated by stable isotope data (described in main report), winter precipitation (snow) contributes much of the groundwater recharge in the area, which again suggests that the recharge temperature of water collected from these springs was significantly cooler than the water temperature measured in the field. Recharge temperature estimates based on noble gas concentrations (described in main report) also indicate that recharge temperatures were cooler than the temperatures of the water samples collected from several wells in the study area. Therefore, groundwater must be flowing at a rate that is slow enough to allow the transfer of heat from subsurface rocks to the groundwater by conduction. Assuming a hydraulic gradient based on local topography, we used a numerical model of groundwater and heat flow to estimate the maximum hydraulic conductivity necessary for the shallow groundwater temperature to exceed that of average annual average air temperature at a specific location.

Identification of groundwater flow paths in the high mountains

Figure 10.4 shows the location of the several tributary drainages to the Rio Peñasco and Agua Chiquita Creek. Specific conductance (SC) measurements in springs sampled within the entire Rio Peñasco drainage do not show any obvious spatial trend (Figure 10.5). However, at a smaller scale, linear regressions indicate correlations between SC for springs and relative distance along the drainages (Figure 10.6) for several tributaries to the Rio Peñasco and the Agua Chiquita Creek. The SC measurements increase with increasing downstream distance. This correlation indicates the presence of a shallow aquifer with a groundwater flow velocity in the downstream direction where water/mineral interactions increase the specific conductance

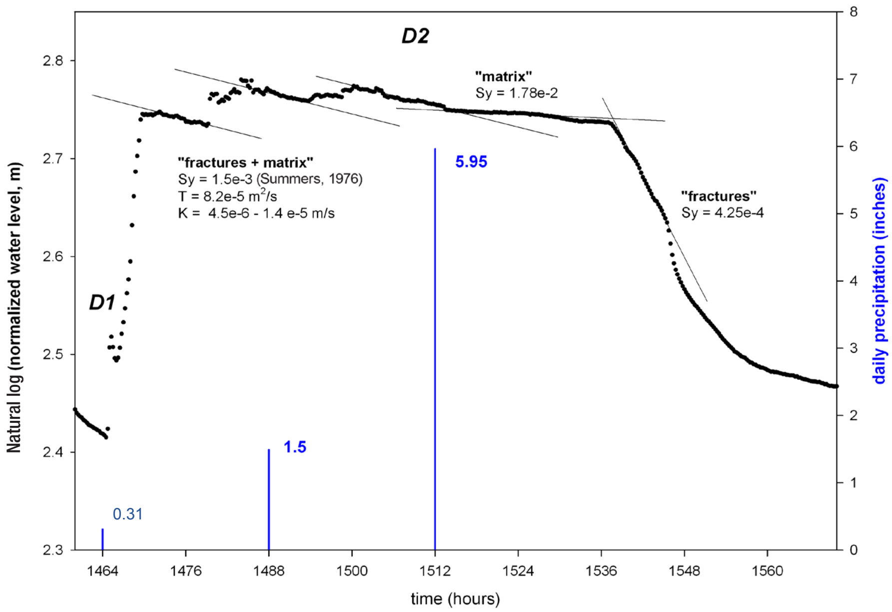


Figure 10.3–Expanded view of water level rise D from well SM-0049 hydrograph. The two individual rises are labeled D1 and D2, which occurred from August 21 to 24, 2006. Straight line segments are labeled with hydrologic properties calculated from them. Precipitation events are shown in blue.

as groundwater flows down gradient. These observations suggest the presence of localized perched aquifers in small drainages (1st to 3rd order). In higher order drainages such as Agua Chiquita Creek and Rio Peñasco, which are two of the few perennial streams in the Sacramento Mountains and are fed by multiple springs, groundwater from tributary drainages mix both in the subsurface and on the surface.

Characterization of the local hydrologic system in Hay Canyon

We chose to base the model domain on Hay Canyon, mainly due to the high correlation between SC and downstream distance (Figure 10.6). Hay Canyon is a northeast trending ephemeral tributary to Agua Chiquita Creek with five springs and one shallow well within it (Figure 10.7). For the purposes of this discussion, the springs have been labeled S1 through S5, which are arranged from upstream to downstream. Figure 10.7 shows the transect along which downstream distance is measured. This transect runs from the top of the drainage (A) to the where the canyon takes a turn towards the south (A'). For this modeling exercise, it is assumed that most groundwater recharge occurs near the top of this drainage, that groundwater flows in the northeasterly direction along the transect and that water discharging at each spring is representative (in terms of temperature and specific conductance) of groundwater at that position. The mechanisms causing groundwater to discharge at the spring locations are not considered.

Figure 10.8 shows temperature and SC as a function of relative downstream distance for each spring on 3/22/06 and 4/08/09. It can be seen that SC measurements taken in March 2006 increase in a linear fashion for S2 through S5, which supports the conceptual model of groundwater chemistry evolving (resulting in an increase in specific conductance) along its flow path. However, the SC value for S1 is higher than that of S2, suggesting that older, more evolved water appears up gradient of S2, indicating that the upper Hay Canyon drainage may not be the only water source for the springs in that canyon and that groundwater from another perched aquifer system is mixing with younger water in Hay Canyon. SC measurements taken in April 2009 indicate that this higher SC water is present in both S1 and S2. Water temperature data also suggests that older water is entering the system upstream of S2, as the water temperature is higher in S1 than in S2. Note that the hydrologic model that will be discussed does not account for all the complexities of the Hay Canyon hydrologic system, such as the addition of older water at the upstream portion of the canyon. The purpose of the model is not to simulate the actual hydrologic system in Hay Canyon, but to serve as a tool to help constrain hydrologic parameters in this system.

In general, water temperatures appear to increase with distance downstream. In order to compare spring water temperatures to average annual surface temperatures at the spring locations an adiabatic lapse rate was estimated using reported average annual temperatures from several weather stations in and around the study area. Figure 10.9 shows a linear regression for average annual temperature as a function of elevation with a relatively high R^2 value of 0.90. Figure 10.10 shows that for all but one spring in Hay Canyon, measured temperatures plot above the average annual surface temperature estimated for the spring locations. The water temperature in S2 plots below this line,

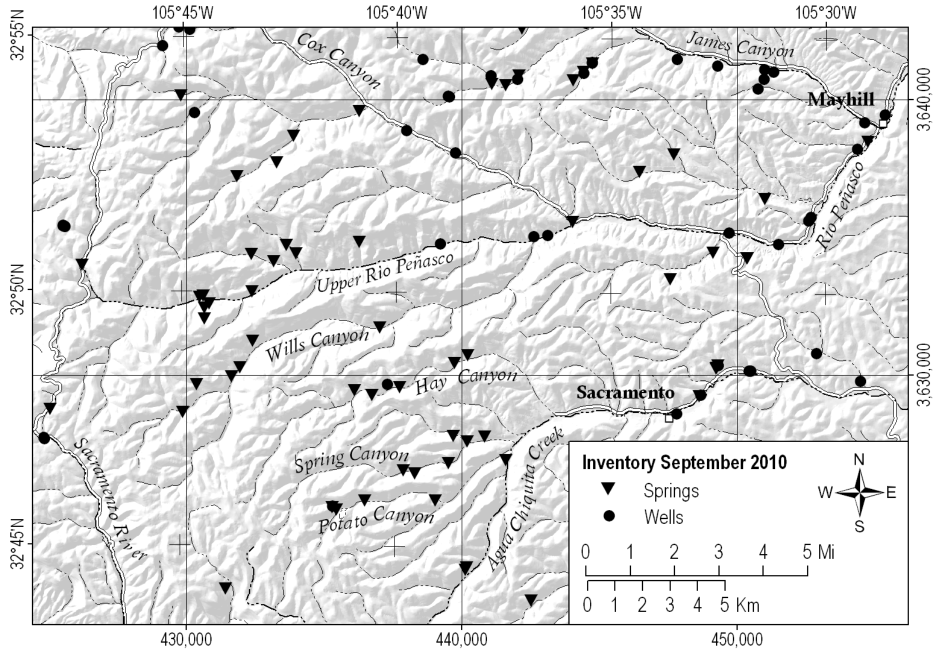


Figure 10.4–Location of different drainages and location of wells and springs.

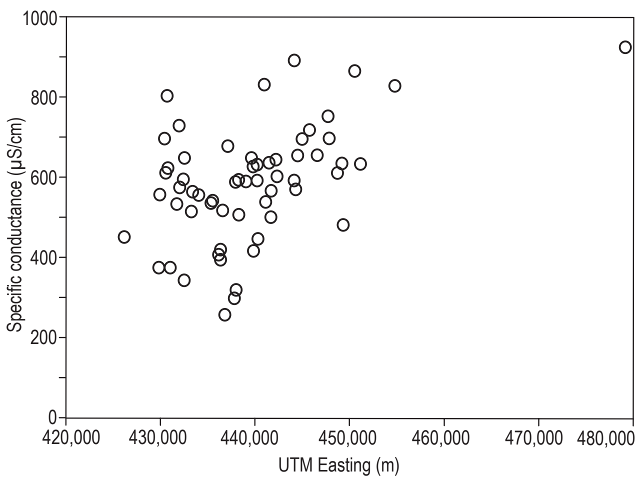


Figure 10.5–Specific conductance values for all springs in the Rio Peñasco drainage. Plot shows that the conductivity of the Rio Peñasco (including tributaries) is highly variable and do not correlate with UTM easting distances.

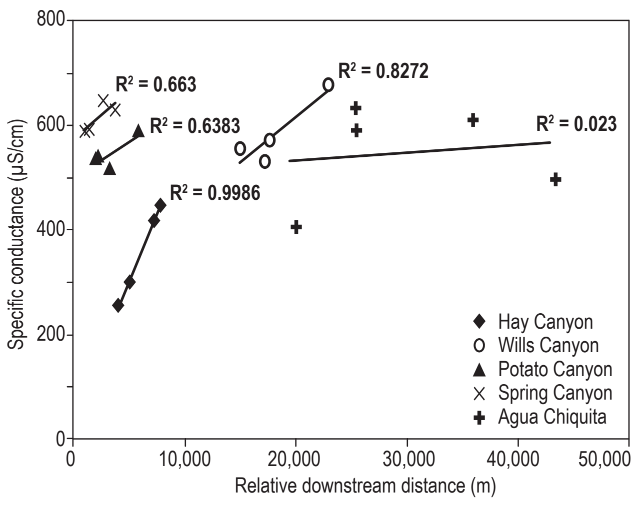


Figure 10.6–Individual small drainages show a correlation between specific conductance and downstream distance along the drainage.

suggesting very young groundwater that has not obtained much heat from the subsurface. This observed trend of shallow groundwater temperatures becoming warmer than the average annual temperature at down gradient locations is the primary criteria that was used to constrain hydrologic parameters in the modeling exercise described below.

Groundwater and heat flow modeling

We used Hydrotherm (Kipp et al., 2008) to model fluid and heat flow in two dimensions along the transect A-A' (Figure 10.7). The USGS code is a finite difference model to simulate groundwater flow and heat transport in the temperature range of 0 to 1200 °C. A topographic profile along the A-A' transect was used as the upper boundary of the model domain (Figure 10.11). As can be seen, we smoothed the profile to avoid modeling apparent very localized topographically controlled groundwater flow. The result is the general slope of the Hay Canyon bottom, which was assumed to be the hydraulic gradient in the shallow aquifer. The model domain consists of one shallow aquifer that sits on a unit of low hydraulic conductivity (Figure 10.12). The hydraulic conductivity of the impermeable basement rock was set to be at least four orders of magnitude lower than that of the aquifer. The dip of the base of the aquifer was set to be similar to average observed dips of strata (~4°). The domain bottom was placed approximately 3,000 feet below the surface boundary to avoid boundary effects in areas of interest in the shallow aquifer. We divided the domain into cells using an irregular grid with smaller cells in the shallow aquifer near the surface. The left and right boundaries were assigned the

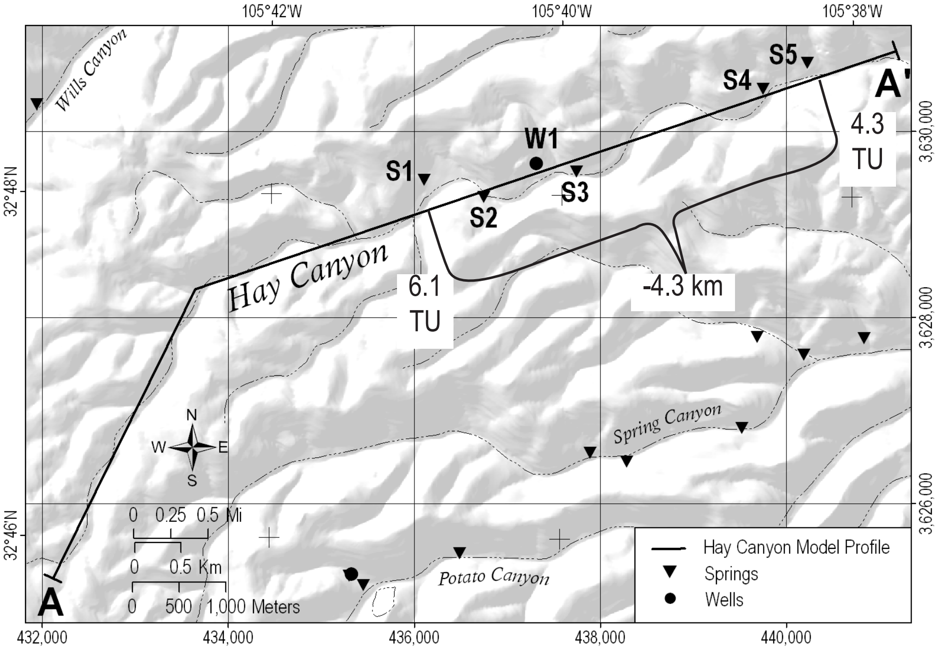
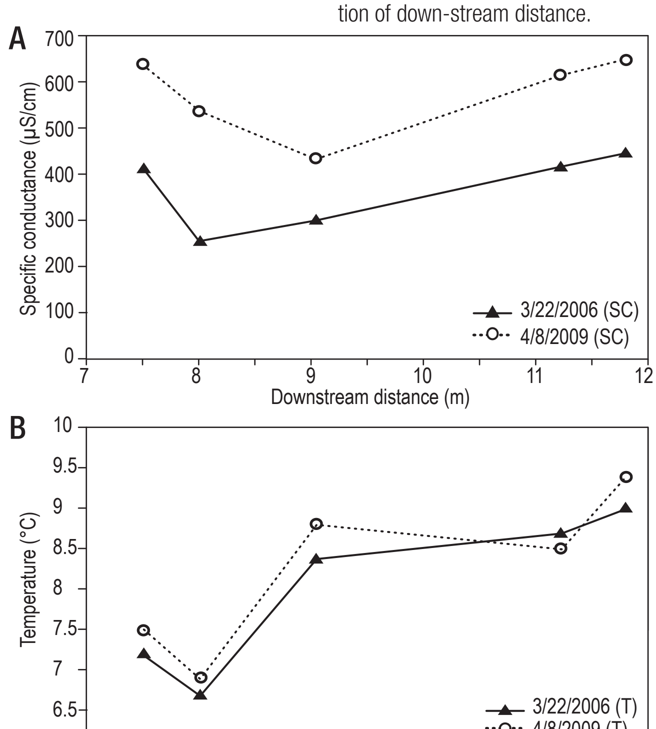


Figure 10.7–The model domain was based on transect A-A' along Hay Canyon, an ephemeral tributary to Agua Chiquita with 5 springs and one well. Tritium values for the springs S1 and S2 provide a means of calculating a groundwater flow velocity. Study point IDs for these sites are: S1=SM-1013, S2=1012, S3=SM-1011, S4=SM-1010, S5=SM-1009, and W1=SM-0018.

Figure 10.8–**A**–Specific conductance (SC), and **B**_temperature data for springs in Hay Canyon as a function of down-stream distance.



boundary conditions of no flow and constant head, respectively. The top boundary was divided into segments that were assigned as constant temperature and pressure boundaries. The temperature and pressure assigned for each segment was the average annual temperature for the segment elevation and atmospheric pressure, respectively. The basal heat flow flux at the bottom of the model domain was set at 60 meter watts per meter squared (mW/m$^{2)}$. Temperature observation points were placed in the cells beneath the top boundary cells at different distances from the no-flow boundary.

Figure 10.13 shows the temperature profile in the model domain under initial conditions. Note that surface temperature decreases with increasing elevation. Initial sub-surface temperatures were based on a geothermal gradient of 20°C/km. Simulations were run with different permeabilities in the shallow aquifer (which was isotropic). Simulations were run for a model time of 1 million years to assure that steady state had been reached. We initially used an aquifer thickness of about 500 feet. Figure 10.14 shows the temperature profile of the model domain at steady state where the permeability of the shallow aquifer materials is $1\text{x}10^{-7}$ cm^2. With this high permeability, cold water recharging the system decreases temperatures in the subsurface where the shallow groundwater temperatures are below average annual surface temperatures. Figure 10.15 shows the effects of groundwater flow on subsurface temperatures in an aquifer with a permeability of $1\text{x}10^{-9}$ cm^2. With a lower permeability, heat from the subsurface is conductively transferred to the groundwater, which advectively carries this heat down gradient, resulting in shallow groundwater temperatures that are higher than the average annual surface temperatures. The modeled shallow groundwater temperatures along the transect for several simulations are similar to measured temperatures from the springs and the well in Hay Canyon (Figure 10.16). It can be seen that no one simulated temperature profile fits the data very well. Simulations with high permeabilities produce groundwater temperatures that are similar to those observed up

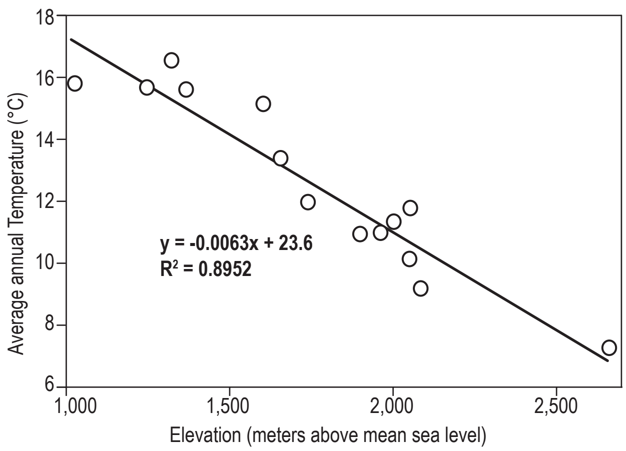


Figure 10.9–The adiabatic lapse rate was estimated with a linear regression of average annual temperature as a function of elevation for several weather stations in and around the study area.

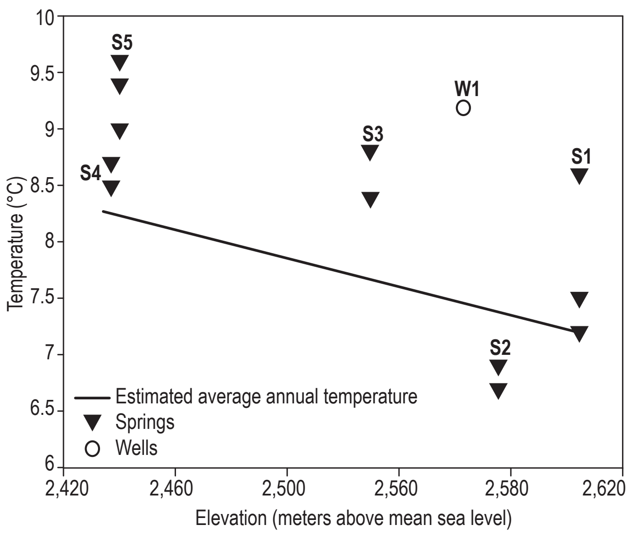


Figure 10.10–Measured temperatures for the well and most springs in Hay Canyon are higher than average annual surface temperatures for elevation of sample locations.

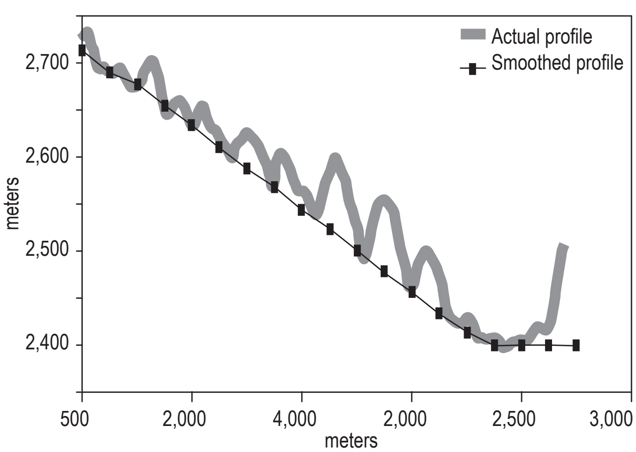


Figure 10.11–Groundwater and heat flow was modeled along a smoothed topographic profile of Hay Canyon, which defined the hydraulic gradient.

gradient, but that are much lower that those observed down gradient. Estimated groundwater temperatures from simulations with lower permeabilities (less than 10^{-7} cm²) generally fit the data better but tend to over estimate up gradient temperatures. This discrepancy between observed and modeled temperatures in the shallow groundwater is most likely due to complexities that cannot be accounted for in this 2-dimensional model. These complexities include the possible mixing of water from other shallow aquifers as discussed above, groundwater entering the system from the sides of the transect, and the effects of water discharging at spring locations on the subsurface temperature profile.

Figure 10.17 is a plot of the difference between the modeled shallow groundwater temperature and the average annual surface temperature as a function of aquifer permeability for all observation points. Observation points are identified by their downstream distance. With permeabilities above the approximate value of 1×10^{-8} cm², most shallow groundwater temperatures are estimated to be below the average annual surface temperature. Groundwater flow in an aquifer with permeabilities less than 1×10^{-8} cm², results in shallow groundwater temperatures that are warmer than the average annual surface temperature, suggesting that 1×10^{-8} cm² is the maximum permeability possible to account for the water temperatures observed in the springs and well in Hay Canyon.

We conducted a sensitivity analysis on aquifer thickness, within the ranges of 10 to 500 feet thickness, and modeling results were very similar.

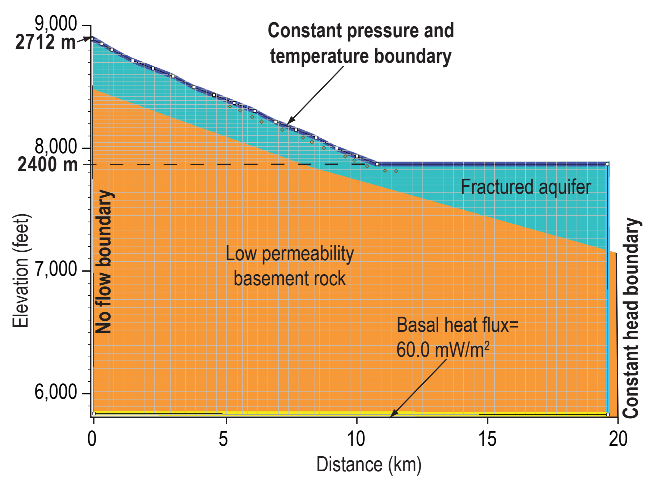


Figure 10.12–The model domain consists of one shallow aquifer (blue) that sits on a unit of low permeability. The hydraulic conductivity of the impermeable basement rock (orange) was set to be at least four orders of magnitude lower than that of the aquifer. The temperature for each segment on the top boundary was set as the average annual surface temperature for the corresponding altitude.

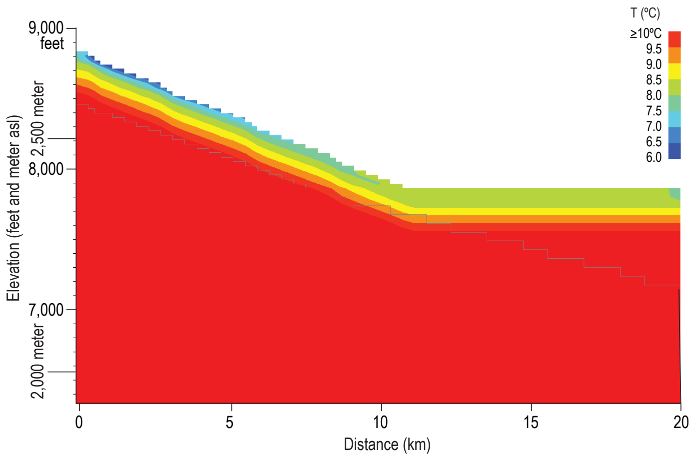


Figure 10.13–Temperature profile under initial conditions. Note that surface temperatures decrease with increasing elevation.

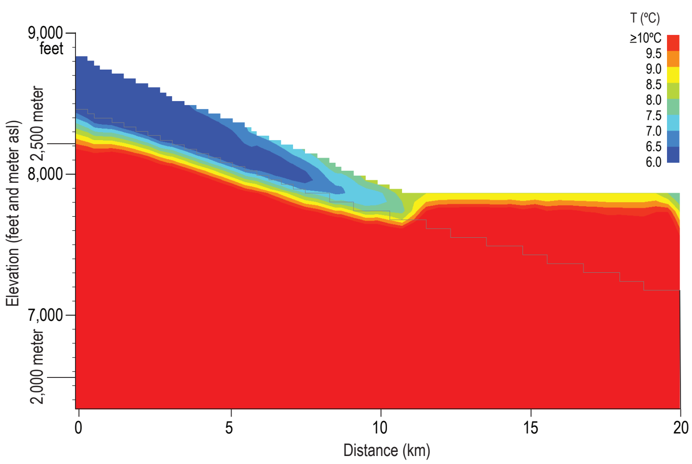


Figure 10.14–Modeled steady state temperature profile with a permeability of 1x10^{-7} cm^2. With this high permeability, high altitude water flows down-gradient quickly and cools shallow aquifer temperatures to where they are lower than average annual surface temperatures.

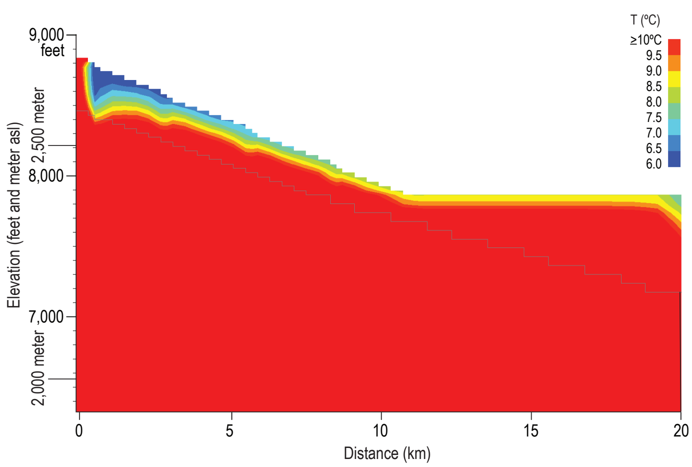


Figure 10.15–Modeled steady state temperature profile with a permeability of 1x10^{-9} cm^2. With lower permeabilities, water from high altitudes moves down-gradient slow enough to pick up heat conductively from subsurface rocks and move it advectively to lower elevations. Note that shallow groundwater temperatures are higher than average annual surface temperatures.

DISCUSSION AND CONCLUSIONS

The estimated maximum permeability of 1x10^{-8} cm^2, which correlates to a hydraulic conductivity of 1x10-5 m/s, is a typical value for fractured rock. This estimate of hydraulic conductivity is one to three orders of magnitude larger than estimates from Wasiolek (1991) and Summers (1976) shown in Table 10.1, which is consistent with this estimate being a maximum constraint on hydraulic conductivity. This estimate of hydraulic conductivity is just slightly lower than the higher limit estimated by hydrograph analysis discussed above. With an estimated hydraulic gradient of 0.04, which represents the slope of the bottom of Hay Canyon, and a hydraulic conductivity of 1x10^{-5} m/s, we estimated a maximum specific discharge of 0.035 m/day. Figure 10.7 shows tritium values for S1 and S5 of 6.1 and 4.3 TU respectively. Therefore, assuming groundwater flows from S1 to S5, the travel time between these two springs is approximately 6 years based on the radioactive decay of tritium, which has a half life of 12.3 years. S1 and S2 are approximately 4.3 km apart, resulting in an estimated groundwater flow velocity of 2 m/day, which when divided into the estimated maximum specific discharge, yields an approximate effective porosity of 0.02, which is a reasonable value for fractured limestone (Domenico and Schwartz, 1998). Using the flow velocity estimated by the tritium data, lower specific discharges would result in even smaller effective porosities, which can be expected for fractured aquifers. On the other hand, if the groundwater flow velocity is slower than that estimated based on the tritium data, the effective porosity would be larger.

The estimated maximum hydraulic conductivity of 1x10^{-5} m/s, based on groundwater and heat flow modeling is a reasonable value for a fractured limestone aquifer. As this estimate is a maximum constraint, other estimates of hydraulic conductivity are generally lower. This estimate of hydraulic conductivity, along with relatively high gradients and low effective porosities can result in high groundwater flow velocities in the high mountains aquifer system.

SUMMARY

We used two independent methods to estimate hydrologic parameters in the Yeso Formation. A hydrograph analysis for a well in the high mountains yielded an estimated hydraulic conductivity range of 4.9x10^{-6} – 7.5x10^{-5} m/s for a portion of the Yeso Formation with an intermediate permeability (fractures + matrix). Numerical modeling of groundwater and heat flow along an assumed flow path in a perched aquifer in Hay Canyon resulted in a maximum hydraulic conductivity estimate of 1x10^{-5} m/s, which is within the range of values estimated using the well hydrograph mentioned above. These hydraulic conductivity estimates are comparable to other estimates by Wasiolek (1991) and Summers (1976) (Table 10.1). Using a linear flow velocity calculated from tritium data in Hay Canyon and the specific discharge based on estimated maximum hydraulic conductivity, we estimated the effective porosity to be 0.02 (2%), which is a reasonable estimate for a fractured aquifer. With high hydraulic gradients, and low effective porosities associated with fractured flow paths, groundwater can flow very quickly in the high mountains.

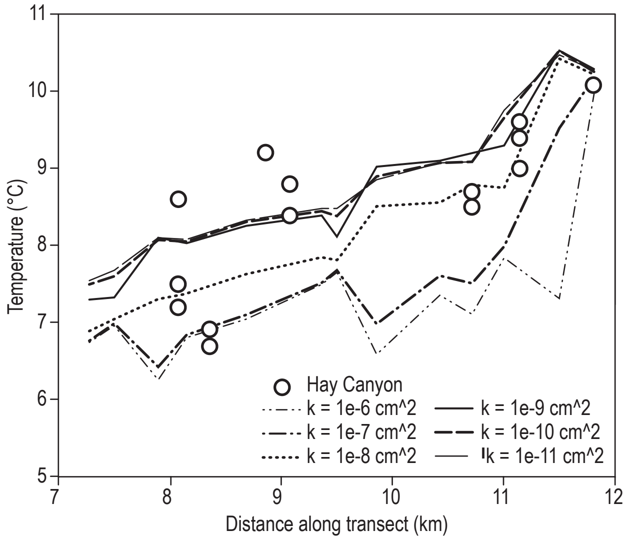


Figure 10.16–Modeled shallow groundwater temperatures are compared to temperatures measured in a well and five springs in Hay Canyon. Temperature profiles resulting from different permeability values (k) are represented with different lines.

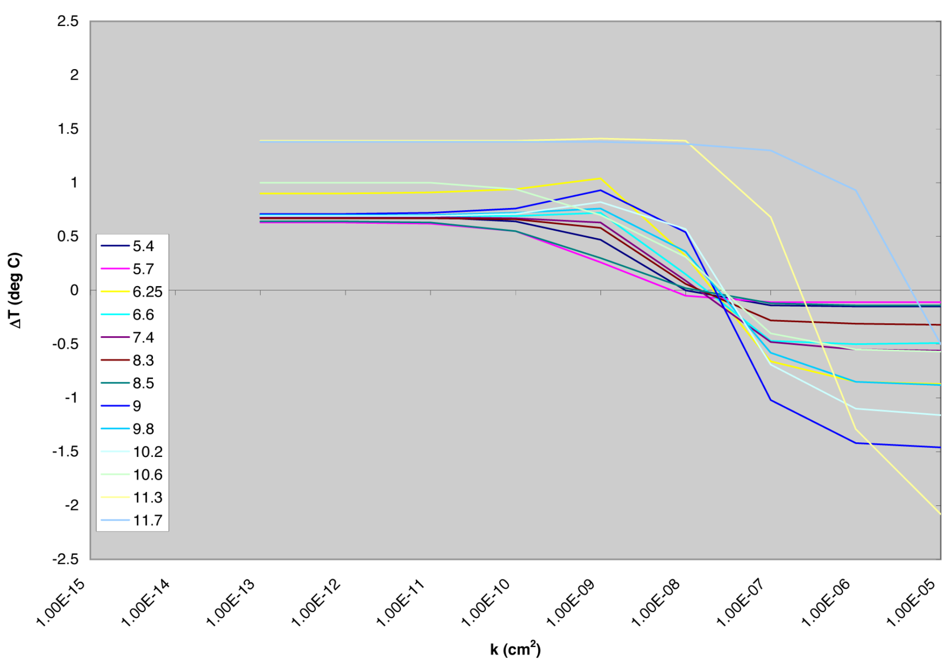


Figure 10.17–The difference between shallow groundwater temperatures and the average annual surface temperature as a function of permeability. Positive values indicate that groundwater temperatures are higher than surface temperatures.

REFERENCES

Domenico, P. A. and Schwartz, F. W., 1998, *Physical and Chemical Hydrogeology*, John Wiley & Sons, Inc., New York.

Kipp, K. L. Jr., Hsieh, P. A., Charlton, S. R., 2008, *Guide to the revised ground-water flow and heat transport simulator:* Hydrotherm – version 3, U.S.G.S. Techniques and Methods 6-A25, 160 p.

Powers, J. G., and Shevenell, L., 2000, T*ransmissivity estimates from well hydrographs in karst and fractured aquifers*. Groundwater, v. 38, p 361-369.

Shevenell, L., 1996, *Analysis of well hydrographs in a karst aquifer: estimates of specific yields and continuum transmissivities*. Journal of Hydrology, v. 174, p. 331-355.

Summers, W. K., 1976. *Groundwater resources of the upper James Canyon basin, Otero County, New Mexico*. W. K. Summers & Associates, Unpublished consultant report for the Village of Cloudcroft, 200 p.

Wasiolek, M., 1991, *The hydrogeology of the Permian Yeso Formation within the upper Rio Hondo Basin and the eastern Mescalero Apache Indian Reservation, Lincoln and Otero Counties, New Mexico,* in Barker, J. M., Kues, B. S., Austin, G. S., and Lucas, S. G. (eds.), *Geology of the Sierra Blanca, Sacramento, and Capitan Ranges, New Mexico:* New Mexico Geological Society, Guidebook 42, p. 343-351.

Wasiolek, M. and Gross, G. W., 1983, *Hydrogeology of the upper Rio Peñasco drainage basin between James and Cox Canyons, Otero County, New Mexico:* New Mexico Tech, Geophysical Research Center, Report H-13, 122 p.