AQUIFER PARAMETER ESTIMATION FROM AQUIFER TESTS AND SPECIFIC-CAPACITY DATA IN CEDAR VALLEY AND THE CEDAR PASS AREA, UTAH COUNTY, UTAH

by J. Lucy Jordan

SPECIAL STUDY 146
UTAH GEOLOGICAL SURVEY
a division of
UTAH DEPARTMENT OF NATURAL RESOURCES
2013

AQUIFER PARAMETER ESTIMATION FROM AQUIFER TESTS AND SPECIFIC-CAPACITY DATA IN CEDAR VALLEY AND THE CEDAR PASS AREA, UTAH COUNTY, UTAH

by J. Lucy Jordan

Cover photo: *A well and water-level measuring device used during an aquifer test in Cedar Valley, Utah County, Utah.*

ISBN: 978-1-55791-869-7

SPECIAL STUDY 146
UTAH GEOLOGICAL SURVEY
a division of
UTAH DEPARTMENT OF NATURAL RESOURCES
2013

STATE OF UTAH
Gary R. Herbert, Governor

DEPARTMENT OF NATURAL RESOURCES
Michael Styler, Executive Director

UTAH GEOLOGICAL SURVEY
Richard G. Allis, Director

PUBLICATIONS
contact
Natural Resources Map & Bookstore
1594 W. North Temple
Salt Lake City, UT 84114
telephone: 801-537-3320
toll-free: 1-888-UTAH MAP
website: mapstore.utah.gov
email: geostore@utah.gov

UTAH GEOLOGICAL SURVEY
contact
1594 W. North Temple, Suite 3110
Salt Lake City, UT 84114
telephone: 801-537-3300
website: geology.utah.gov

CONTENTS

FIGURES

TABLES

AQUIFER PARAMETER ESTIMATION FROM AQUIFER TESTS AND SPECIFIC-CAPACITY DATA IN CEDAR VALLEY AND THE CEDAR PASS AREA, UTAH COUNTY, UTAH

by J. Lucy Jordan

ABSTRACT

The purpose of this study was to investigate the nature of and determine the hydraulic properties of the basin-fill and bedrock aquifers in and near Cedar Valley in Utah County, north-central Utah. The hydraulic properties and aquifer characteristics determined in this study are two of the primary building blocks that were used to construct conceptual and digital groundwater-flow models. These models are part of a larger groundwater resource study that is key to groundwater resource development and protection in rapidly growing Utah County.

Primarily by conducting aquifer tests, I defined the hydraulic conductivity, transmissivity, aquifer thickness, specific storage, specific yield, storativity (storage coefficient), and, in some cases, the fracture conductivity, vertical to horizontal hydraulic conductivity ratio, horizontal anisotropy, and well-bore skin of the aquifers and the wells that penetrate them.

With the help of others, I conducted aquifer tests on the two most important aquifers in the study area—the principal basin-fill aquifer and the fractured-bedrock aquifer. The aquifer tests on bedrock wells are of particular interest because of the importance of the bedrock groundwater resource in the Cedar Pass area, where surface water and shallow groundwater are scarce, and because the tests provide much-needed scientific information on which to base water-right decisions near the boundary between two water-right administration areas. Additionally, the interface between basin-fill and bedrock aquifers near the aquifer-test sites at Cedar Pass is a key component in understanding the nature of groundwater discharge from the Cedar Valley groundwater basin. This report investigates previously suggested and newly identified geologic controls on groundwater flow near this interface.

Fieldwork for the aquifer tests involved measuring water levels in pumping and observation wells before, during, and after defined pumping periods. I made corrections to the water-level data for barometric pressure effects and antecedent water-level trends before I matched analytical type curves for different aquifer classes to the data using computer software. The aquifer model matched to each data set depended on the hydrogeologic setting involved in a particular test. I conducted three aquifer tests on the unconsolidated basin-fill aquifer in confined or leaky confined settings, and two tests on the fractured sedimentary bedrock aquifer at Cedar Pass in double-porosity fractured, confined, or unconfined settings. I supplemented aquifer-test data by using specific-capacity data to estimate transmissivity.

The thickness of the aquifers, which I determined using well logs, averaged 180 feet (55 m), but ranged from 5 to 1044 feet (1.5–318 m), depending on the well construction and location of the well. Hydraulic conductivity in Cedar Valley ranges over five orders of magnitude, from 2.6×10^{-3} to 5.3×10^{2} feet per day (7.9×10^{-4}–1.6×10^{2} m/d). Transmissivity ranges over six orders of magnitude, from 2.7×10^{-1} to 1.2×10^{5} feet squared per day (2.5×10^{-2}–1.1×10^{4} m^2/d). The hydraulic conductivity and transmissivity values in basin fill are more evenly distributed spatially and fall within a smaller range than those in bedrock because groundwater flow in bedrock is controlled primarily by fractures, which are unevenly distributed throughout the aquifer. Most transmissivity and hydraulic-conductivity values were within published ranges for the type of sediment or rock tested. The zones of highest hydraulic conductivity in the basin fill (20–50 ft/d [6–15 m/d]) are in moderately coarse alluvial fan sediments along the western and eastern margins the valley, and the zone of highest hydraulic conductivity in the bedrock aquifer (200–700 ft/d [61–210 m/d]) is on the eastern margin of the valley.

One of the three aquifer tests on basin-fill wells showed greater transmissivity in the direction parallel to the edge of the aquifer's confining unit, which coincides with fining of the aquifer and a corresponding decrease in hydraulic conductivity.

A five-month-long aquifer test on fractured bedrock showed the aquifer to have double porosity due to higher hydraulic conductivity in the fracture network and lower hydraulic conductivity in the matrix blocks. Anisotropy in this aquifer, determined using ideally cited observation wells, is two to three times greater in the direction parallel to structural fold axes.

The drawdown response in the observation wells shows that one or more impermeable or semi-permeable aquifer boundaries are located at the interface between the Cedar Valley unconsolidated basin fill and the Paleozoic Oquirrh Group carbonate- and clastic-rock aquifer. A wedge of Tertiary volcanic rocks and a normal fault are the likely barriers to groundwater flow. The aquifer may thin to the east, further limiting flow.

Another long-term (35 days) aquifer test in the Paleozoic Great Blue Limestone did not show the effects of a boundary, even though a thrust fault is near the well. This part of the bedrock aquifer has similar transmissivity to the Oquirrh Group aquifer.

INTRODUCTION

Hydraulic properties of an aquifer can be determined by conducting aquifer tests and measuring the specific capacities of wells. With help from well owners and my Utah Geological Survey (UGS) colleagues, I conducted five aquifer tests in Cedar Valley, analyzed data from seven pump tests conducted by geologic consultants and well-drilling contractors, and evaluated specific-capacity data from 70 wells located in Cedar Valley (figure 1). Additionally, I examined the specific capacity of 25 wells in neighboring northern Utah and Goshen Valleys to determine regional aquifer properties. In this report, I discuss the logistics, data analysis, and results from each test conducted by the UGS in individual sections; briefly describe my analysis of tests conducted by others; explain how I computed transmissivity and hydraulic conductivity from specific capacity; and summarize and discuss the properties of the aquifers in Cedar Valley as a whole. Transmissivity, hydraulic conductivity, and storativity estimates from these investigations are included in table 1.

BACKGROUND

This work is part of a larger groundwater resources investigation and groundwater modeling study in Cedar Valley, Utah County, north-central Utah, the results of which are documented in Jordan and Sabbah (2012). The study area includes Cedar Valley and parts of the adjacent mountain ranges and valleys. Cedar Valley is a fault-controlled valley surrounded by mountain ranges composed primarily of Paleozoic sedimentary rocks deformed into broad north- to northwest-striking folds (Bryant and Nichols, 1988; Allmendinger, 1992). Tertiary volcanic rocks are common in the surrounding mountains and are indicated on some well logs. The basin contains as much as 2200 feet (670 m) of fine-grained lake deposits interfingered with coarser grained and poorly sorted sediment along the valley margins (Hurlow, 2004).

Four important hydrogeologic units exist in the Cedar Valley groundwater system: (1) the principal basin-fill aquifer, (2) a near-surface clay unit, as much as 240 feet (73 m) thick, that confines the basin-fill aquifer over much of the valley, (3) Paleozoic carbonate and clastic units grouped in Jordan and Sabbah (2012) as the fractured-bedrock aquifer, and (4) a small perched basin-fill aquifer at Cedar Pass. Aquifer testing was conducted on the two most important aquifers—the principal basin-fill aquifer and the fractured-bedrock aquifer.

Precipitation on the Oquirrh Mountains, which border the valley on the northwest, is the primary recharge area for the basin-fill and bedrock aquifers (Feltis, 1967; Jordan and Sabbah, 2012). Groundwater-flow direction is generally from west to east, and there is interflow between the bedrock and basin-fill aquifers (Jordan and Sabbah, 2012).

GENERAL METHODS

I conducted five aquifer tests in Cedar Valley and Cedar Pass in 2005, 2006, and 2007. For this work, the most important aquifer-test site selection criteria were well location and degree of cooperation from the well owner. Pumping history, the existence and condition of observation wells, and a disposal site for pumped water were secondary concerns, and, therefore, not every test was an ideal controlled aquifer test. Valuable information was obtained from all the tests despite imperfect test conditions.

Water levels in the pumping and observation wells were monitored as frequently as possible for several weeks before and during the tests using electronic water-level sounders or steel tapes. The same equipment was used on each well throughout a test to avoid introducing error due to differences between sounders. Where possible, pressure transducers were also used before, during, and after the tests. Both absolute and vented pressure transducers were used; absolute pressure transducer water-level data were corrected for atmospheric pressure using atmospheric pressure data collected by the National Weather Service at 15-minute intervals at the Fairfield, Lehi, and/or Saratoga Springs weather stations (University of Utah Department of Atmospheric Sciences, 2008). Appendix A contains the water-level data collected for the tests.

My primary method of aquifer-test data analysis was to match theoretical curves to plotted groundwater-level drawdown and recovery data using a commercially available computer software program. The curve matching technique for transient conditions (non-steady state) was pioneered by Theis (1935), and has been continuously refined by others. The specific test methods used are discussed in the individual sections on each test.

I estimated hydraulic conductivity and transmissivity from specific-capacity data using a method by Bradbury and Rothschild (1985), which is based on an approximation of the Theis (1935) equation for transient radial flow to a well. The spe-

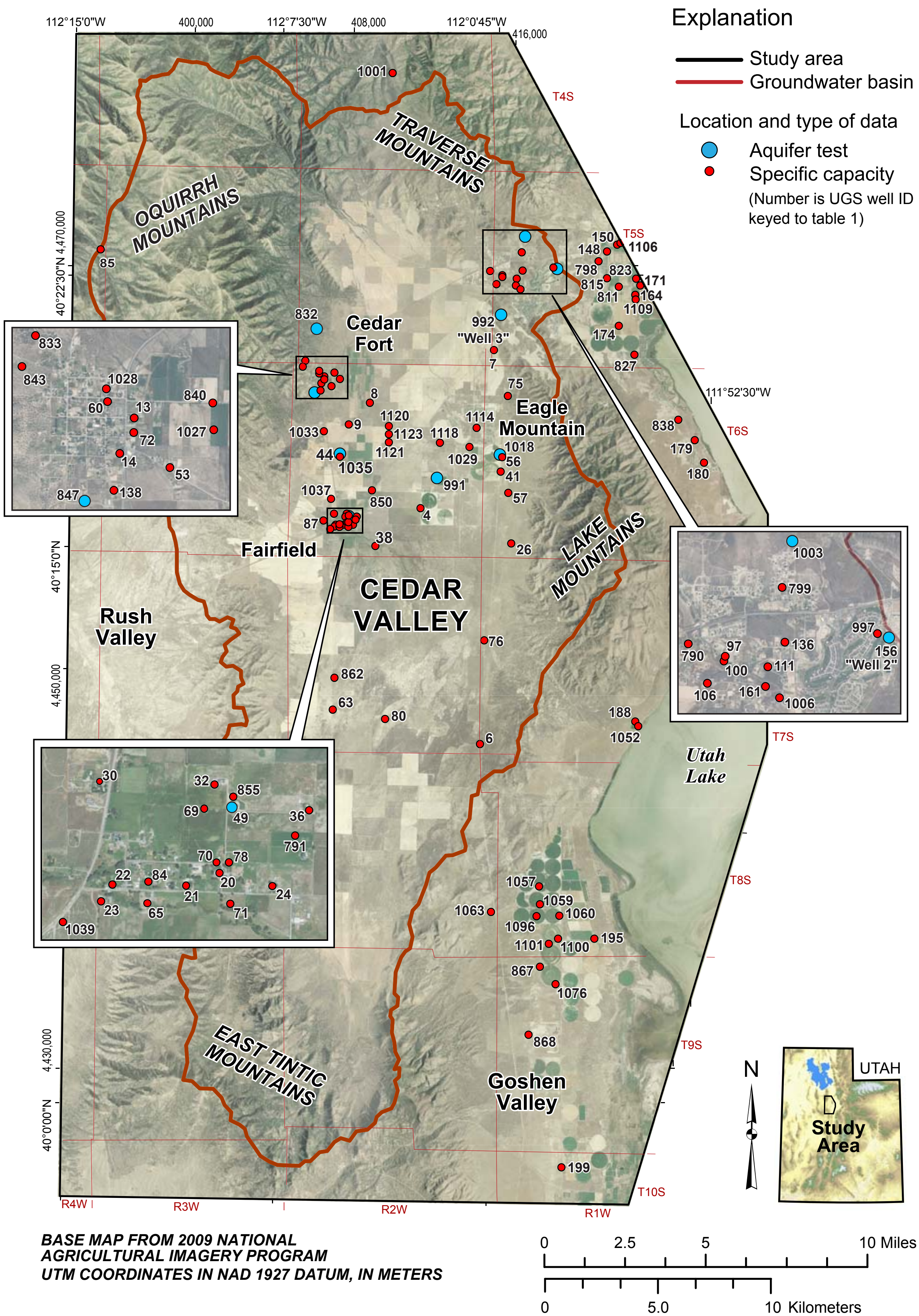


Figure 1. *Location of aquifer tests and specific-capacity data in the Cedar Valley study area, Utah County, Utah.*

Table 1. *Transmissivity (T) and hydraulic conductivity (K) calculated from aquifer tests and specific-capacity (SC) data.*

UGS ID	UTM easting (NAD27 m)	UTM northing (NAD27 m)	Screen Top (depth, ft)	Screen Bottom (depth, ft)	Aquifer[1]	S [2]	T sq ft/d	K ft/day	Source
44	407288	4460602	170	587	BF	0.02	25,000	60	Aquifer Test
49	407730	4457488	262	281	BF	0.01	70	2.0	Aquifer Test
156 "Well 2"	418187	4469812	510	940	bdrx	0.007	13,000	24	Aquifer Test
832	406121	4466823	818	1280	bdrx	0.025	400	0.9	Aquifer Test
847	406023	4463653	380	465	bdrx	0.01	441	7.7	Aquifer Test
991	412149	4459389	120	400	BF	0.01	1100	3.3	Aquifer Test
992 "Well 3"	415370	4467533	700	1210	bdrx	0.02	11,000	25	Aquifer Test
1003	416576	4471417	671	830	bdrx	0.01	20	0.3	Aquifer Test
1018	415304	4460557	355	475	bdrx	0.01	32,000	200	Aquifer Test
4	411326	4457902	210	500	BF	0.01	1740	6.0	Est. from SC
6	414312	4446142	400	505	BF	0.01	0	0.0	Est. from SC
7	415022	4465777	429	572	bdrx	0.01	2029	14	Est. from SC
8	408799	4463135	80	100	BF	0.01	367	18	Est. from SC
9	407734	4462061	58	202	BF	0.03	273	1.9	Est. from SC[3]
13	406511	4464446	101	250	BF	0.01	69	0.5	Est. from SC
14	406364	4464110	223	296	BF	0.01	581	8.0	Est. from SC
20	407661	4457130	101	265	BF	0.01	143	0.9	Est. from SC
21	407470	4457063	141	284	BF	0.01	120	0.8	Est. from SC
22	407067	4457069	0	30	clay	0.01	2083	69	Est. from SC
23	407006	4456979	191	203	BF	0.01	212	18	Est. from SC
24	407941	4457061	180	196	BF	0.01	94	5.9	Est. from SC
26	415877	4456130	200	235	bdrx	0.01	774	22	Est. from SC
30	406997	4457628	111	194	BF	0.01	309	8.6	Est. from SC
32	407625	4457612	112	211	BF	0.01	188	1.9	Est. from SC
36	408142	4457471	154	220	BF	0.01	106	1.6	Est. from SC
38	409065	4456011	235	275	BF	0.01	326	8.2	Est. from SC
41	415353	4459713	497	583	BF	0.01	2993	35	Est. from SC
53	406856	4463973	180	200	BF	0.01	594	30	Est. from SC
56	415306	4460537	250	471	bdrx	0.01	117,300	531	Est. from SC
57	415735	4458665	255	335	BF	0.01	780	9.7	Est. from SC
60	406248	4464604	231	235	BF	0.01	290	21	Est. from SC
63	406920	4447869	140	318	BF	0.01	28	0.2	Est. from SC
65	407258	4456968	101	258	BF	0.01	93	0.6	Est. from SC
69	407569	4457481	100	250	BF	0.01	184	1.2	Est. from SC
70	407636	4457189	100	193	BF	0.01	75	0.8	Est. from SC
71	407712	4456965	100	240	BF	0.01	403	2.9	Est. from SC
72	406504	4464307	128	190	BF	0.01	21	0.3	Est. from SC
75	415706	4463482	500	540	bdrx	0.01	15	0.4	Est. from SC
76	414520	4451315	190	465	BF	0.01	540	2.0	Est. from SC
78	407702	4457190	100	240	BF	0.01	81	0.6	Est. from SC
80	409559	4447388	400	405	BF	0.01	26	5.1	Est. from SC
84	407263	4457086	101	265	BF	0.01	271	1.7	Est. from SC
85	395283	4470782	380	410	bdrx	0.01	108	3.6	Est. from SC
87	406468	4457291	101	240	BF	0.01	49	0.4	Est. from SC
97	415442	4469485	183	223	bdrx	0.01	11	0.3	Est. from SC
100	415429	4469418	165	205	bdrx	0.01	1438	36	Est. from SC
106	415140	4469049	620	680	bdrx	0.01	14	0.2	Est. from SC
111	416160	4469317	480	540	bdrx	0.01	23	0.4	Est. from SC
136	416441	4469734	210	290	perched	0.01	250	3.1	Est. from SC
138	406312	4463754	258	285	BF	0.01	121	4.5	Est. from SC
161	416121	4468996	440	502	bdrx	0.01	4	0.1	Est. from SC
790	414824	4469703	350	480	bdrx	0.01	31	0.2	Est. from SC
791	408065	4457333	100	288	BF	0.01	126	0.7	Est. from SC

[1] Aquifer codes: BF = basin fill; bdrx = bedrock; perched = perched aquifer in the Cedar Pass area; clay = completed in the valley clay unit. More than one aquifer code is given if the well is screened or open to more than one aquifer.

[2] S is the unitless storativity estimate derived from aquifer test analysis where available or estimated as 0.01 for all other wells.

[3] Wells were invovled in a multi-well aquifer test, but transmissivity values were unreasonably high due to distance from the pumping well. Specific capacity is a better estimate in these cases.

Table 1 continued. *Transmissivity (T) and hydraulic conductivity (K) calculated from aquifer tests and specific-capacity (SC) data.*

UGS ID	UTM easting (NAD27 m)	UTM northing (NAD27 m)	Screen Top (depth, ft)	Screen Bottom (depth, ft)	Aquifer[1]	S [2]	T sq ft/d	K ft/day	Source
798	420271	4470188	400	1065	bdrx	0.01	115,300	185	Est. from SC
799	416405	4470642	124	1000	bdrx	0.01	184	0.2	Est. from SC
833	405556	4465238	500	560	bdrx	0.01	20	0.3	Est. from SC
840	407019	4464649	100	321	BF	0.01	227	1.0	Est. from SC
843	405427	4464949	302	362	bdrx	0.01	139	2.3	Est. from SC
850	408901	4458787	288	290	BF	0.01	276	23	Est. from SC[3]
855	407727	4457545	102	255	BF	0.01	95	0.6	Est. from SC
862	407015	4449451	130	382	clay-BF	0.01	79	0.3	Est. from SC
997	417996	4469874	650	1040	bdrx	0.007	7162	42	Est. from SC
1001	409928	4479561	680	890	bdrx	0.01	151	0.7	Est. from SC
1006	416351	4468797	482	542	bdrx	0.01	12	0.2	Est. from SC
1027	407281	4464333	141	375	BF	0.01	624	9.3	Est. from SC
1028	406241	4464729	215	216	bdrx	0.01	517	47	Est. from SC
1029	413783	4460942	120	339	BF	0.01	1523	7.0	Est. from SC
1033	406472	4461733	185	448	BF	0.01	13,000	49	Est. from SC
1035	407285	4460461	150	555	BF	0.02	8939	34	Est. from SC
1037	406843	4458367	65	640	clay-BF	0.006	108	0.5	Est. from SC[3]
1039	406796	4456866	205	595	BF	0.01	1069	5	Est. from SC
1057	417267	4439047	227	345	bdrx	0.01	57,500	488	Est. from SC
1059	417329	4438171	383	700	bdrx	0.01	221	0.7	Est. from SC
1063	414873	4437792	300	390	bdrx	0.01	25	0.3	Est. from SC
1114	414150	4461900	150	1254	clay-BF	0.01	101	0.1	Est. from SC
1118	412301	4461163	120	600	BF	0.01	180	0.4	Est. from SC
1120	409744	4461976	185	835	BF	0.01	978	1.5	Est. from SC
1121	409734	4461185	190	405	BF	0.01	913	5.7	Est. from SC
1123	409739	4461587	278	955	BF	0.01	690	1.0	Est. from SC
Wells outside Cedar Valley but within the Cedar Valley study area used to refine bedrock aquifer hydraulic parameters.									
148	420693	4470685	180	275	BF	0.01	1322	14	Est. from SC
150	421200	4471027	130	240	BF	0.01	331	3.0	Est. from SC
164	422101	4468514	100	118	BF	0.01	457	25	Est. from SC
171	422359	4468988	100	180	BF	0.01	116	1.4	Est. from SC
174	421269	4466977	220	300	BF	0.01	463	5.8	Est. from SC
179	425092	4461276	240	274	BF	0.01	62	1.8	Est. from SC
180	425563	4460151	100	110	BF	0.01	56	5.6	Est. from SC
188	422102	4447267	100	190	BF	0.01	41	0.5	Est. from SC
195	420057	4436444	250	350	BF	0.01	205	2.1	Est. from SC
199	418395	4425075	406	850	BF	0.01	6726	19	Est. from SC
811	421273	4468922	100	160	BF	0.01	186	3.1	Est. from SC
815	420691	4469345	126	215	BF	0.01	78	0.9	Est. from SC
823	422145	4469329	128	203	BF	0.01	482	6.4	Est. from SC
827	422050	4465546	151	163	BF	0.01	196	16	Est. from SC
838	424264	4462306	130	364	BF-bdrx	0.01	2849	12	Est. from SC
867	417321	4435055	325	798	unknown	0.01	441	0.9	Est. from SC
868	416740	4431667	253	293	BF	0.01	4	0.1	Est. from SC
1052	422245	4447040	116	125	BF	0.01	478	53	Est. from SC
1060	418285	4437590	100	595	BF-bdrx	0.01	8991	21	Est. from SC
1076	418101	4434180	280	740	BF	0.01	5012	16	Est. from SC
1096	417156	4437572	185	236	BF	0.01	3331	65	Est. from SC
1100	418230	4436462	230	700	BF-bdrx	0.01	934	3.8	Est. from SC
1101	417754	4436200	180	675	perched-bdrx	0.01	746	1.5	Est. from SC
1106	421381	4471097	100	216	BF	0.01	671	15	Est. from SC
1109	422116	4468292	100	130	BF	0.01	116	3.9	Est. from SC

[1] Aquifer codes: BF = basin fill; bdrx = bedrock; perched = perched aquifer in the Cedar Pass area; clay = completed in the valley clay unit. More than one aquifer code is given if the well is screened or open to more than one aquifer.

[2] S is the unitless storativity estimate derived from aquifer test analysis where available or estimated as 0.01 for all other wells.

[3] Wells were invovled in a multi-well aquifer test, but transmissivity values were unreasonably high due to distance from

cific capacity of a well is calculated by dividing the discharge rate of a well by the drawdown measured in the well during pumping, ideally after the pumping water level has stabilized. I gleaned the required discharge and drawdown information from logs of wells in Cedar Valley and the surrounding areas.

BASIN-FILL AQUIFER TEST ON TWO LARGE-DIAMETER IRRIGATION WELLS

In cooperation with a local farmer at the start of the 2005 irrigation season, UGS personnel conducted a three-week drawdown test involving two large irrigation wells on the western side of Cedar Valley near the community of White Hills (figure 2). The drawdown test was extended to 69 days by measuring water levels on most of the 12 observation wells near the end of the irrigation season. The two pumping wells are completed in the principal basin-fill aquifer near the western edge of a near-surface clay confining unit and near the eastern edge of alluvial fans coming off the Oquirrh Mountains to the west. The wells are 456 feet (139 m) apart from each other and have UGS identifiers 44 [or "North Well," cadastral location (C-6-2)17dcc- 2] and 1035 [or "South Well," (C-6-2)17dcc-1]. They were drilled in the early 1960s as irrigation wells for M. K. White, and are therefore commonly named the White wells. Based on the well driller's log (appendix B), the North Well (ID 44) has a 16-inch-diameter (41 cm) casing and multiple short perforated intervals open to "conglomerate" and boulders between 170 and 587 feet (52–179 m) below land surface (table 2), giving the well 139 feet (42 m) of perforated casing over 417 feet (127 m) of wellbore. Above the top of the perforated intervals (from 0 to 170 feet [0–52 m] below surface), the lithology is primarily clay, which confines the aquifer. The producing units are separated by clay intervals ranging in thickness from 4 to 37 feet (1.2–11 m), which may separate the aquifer into hydrologically distinct zones; however, no potentiometric data are available to determine the degree of possible separation, so I treat the aquifer as one interbedded unit.

The South Well (1035) has a 16-inch-diameter (41 cm) casing and multiple perforated intervals open to gravel, boulders, and "conglomerate" between 150 and 555 feet (46–169 m) below land surface (table 2), giving the well 146 perforated feet (44 m) through 405 feet (123 m) of wellbore (appendix B). The lithology in 1035, including the thickness and position of clay intervals, correlates well with that of 44, increasing the possibility that what I treat as one confined basin-fill aquifer is made of distinct aquifer zones.

UGS personnel measured water levels at 12 observation wells located 0.7 to 2.2 miles (1.1–3.5 km) from the pumping wells, and a stream gage operated by the Utah Division of Water Rights monitored discharge at Fairfield Spring during the aquifer test. Well completion and aquifer information at locations involved in this test is listed in table 2. Water levels collected for this test are tabulated in appendix A, table A-1. The static water level in 44 before the start of the test was 19.9 feet (6.07 m) below land surface datum. Since 1984, there has been no access for water-level measurement in 1035, so the most recent water level measured by the U.S. Geological Survey was 14 feet (1.3 m) below land surface in 1983.

UGS personnel measured water levels using either electronic water-level sounders or steel tapes in pumping well 44 and 12 observation wells. Well-head configuration and downhole equipment limited the use of pressure transducers to measure water levels to only two of the observation wells. Pressure transducer data were compensated for atmospheric pressure changes and barometric efficiency of the wells. UGS personnel monitored water levels in the wells for two or more weeks before the start of the test. I noted rising water-level trends in the background water-level data in observation wells completed in an unconfined part of the aquifer west of the pumping wells, but declining water levels in most of the other observation wells, which are completed in the confined aquifer. To correct for antecedent trends, I calculated the water-level response to pumping (drawdown) by subtracting observed water levels from the heads projected from the antecedent trend as shown in figure 3.

I measured discharge rates from pumping well 44 during the aquifer test using a Controlotron clamp-on portable flow meter. The discharge from 44 was 3400 gallons per minute (208 L/s) at the start of the test but decreased throughout the first hour of pumping while the empty irrigation lines filled and pressurized (figure 4). At various times throughout the test, an in-line booster pump was used; discharge from 44 with and without the booster pump operating was approximately 2500 and 1600 gallons per minute (158 and 101 L/s), respectively. Discharge from pumping well 1035 was not metered. Flow estimates are based on the known volume of water that the irrigation sprinklers emitted, or 1200 gallons per minute (76 L/s). Pumping at variable rates from both wells was considered during aquifer-test analysis.

Neither pumping well had been pumped since the previous fall except to test the pumps for less than 1.5 hours the day prior to the test (figure 4). Water level had recovered from this short pumping period by the start of the aquifer test. The North Well (44) was pumped continuously from July 7 to July 24, 2005, except for a six-hour period on July 13. The South Well (1035) began pumping four days into the test and pumped intermittently throughout the test. Irrigation demand led to the use of various combinations of downhole and booster pumps, which produced variable discharge rates throughout the test, including two periods having zero discharge at 6 and 17 days into the test, as shown on figure 4. After July 26, both wells were pumped for irrigation, and although detailed pumping records are not available, the two wells together averaged approximately 1800 gallons per minute (114 L/s) for the duration of the summer (Utah Division of Water Rights, 2009a), which would be possible if one well was pumped at all times and the other supplemented the flow periodically.

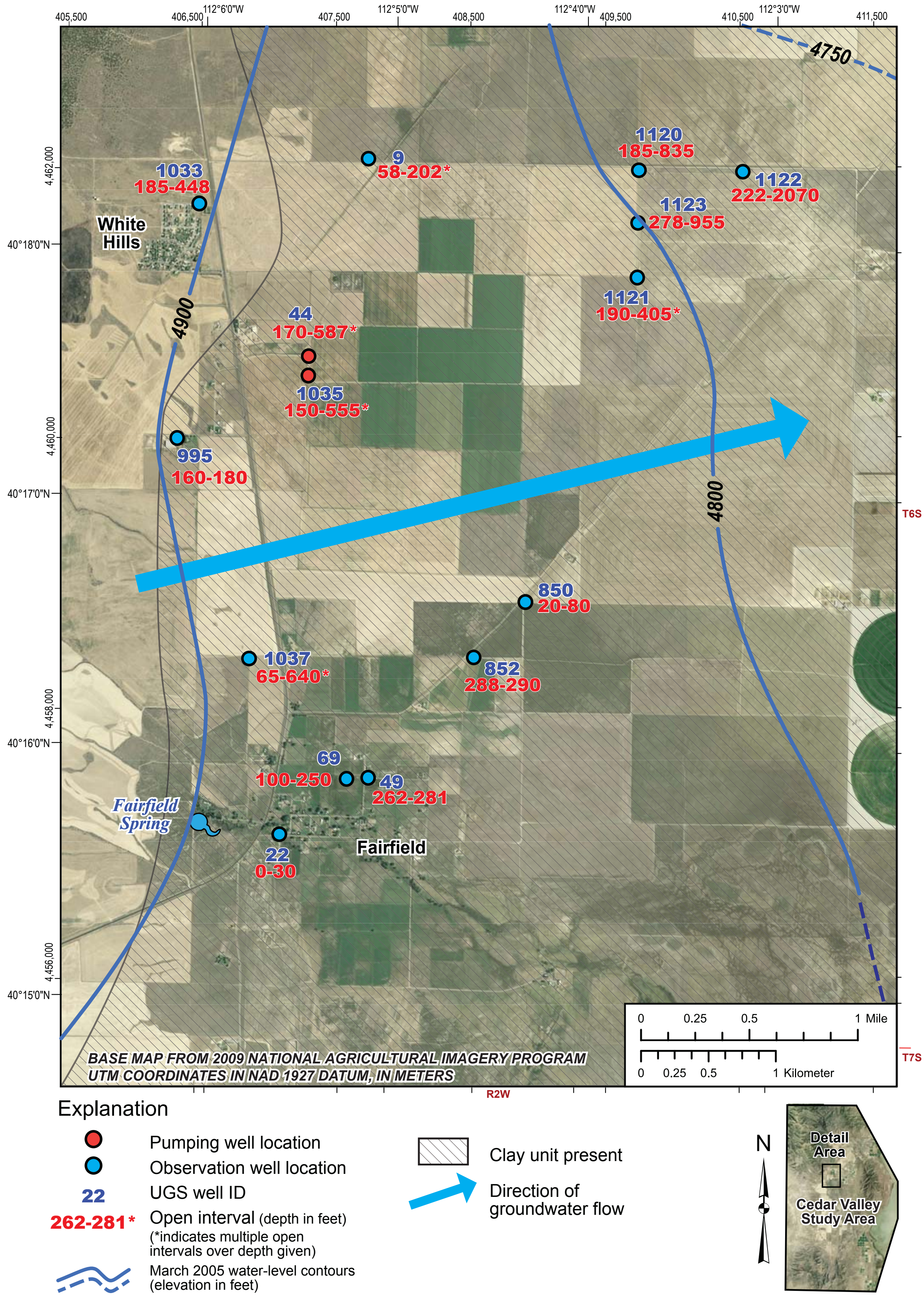


Figure 2. *Location and screen interval of pumping and observation wells involved in the White wells aquifer test.*

Table 2. *Aquifer and well information at locations involved in the White wells aquifer test.*

UGS ID or name	UTM easting (m)	UTM northing (m)	Land elevation (ft)	Dia-meter (in)	Distance from 44 (ft)	Maximum drawdown (ft)	Depth to top of aquifer (ft)	Screen depth (ft)
44 (North Well)	407288	4460602	4923.9	16	0	32	170	170–174, 238–248, 325–350, 365–371, 410–440, 465–481, 488–493, 530–544, 550–574, 582–587
1035 (South Well)	407285	4460461	4916.1	16	456	unknown	147	150–175, 237–246, 350–376, 422–432, 445–492, 525–555.
9	407734	4462061	4923.2	6	5040	0.69	42	58–70, 70–100, 100–140, 140–165, 165–202
22	407067	4457069	4879.1	8	11,611	conf.	conf.	0–30
49	407730	4457488	4869.5	6	10,299	1.2	36	262–281
69	407569	4457481	4870.0	6	10,260	conf.	conf.	100–250
850	408901	4458787	4856.0	9	7930	0	158	288–290
852	408517	4458378	4861.0	6	8317	conf.	conf.	20–80
995	406309	4459999	4959.0	6	3786	0.73	115	160–180
1037	406843	4458367	4891.7	16	7461	1.08	108	65–75, 105–130, 153–163, 214–224, 267–328, 329–338, 343–350, 408–420, 548–558, 590–640
1120	409744	4461976	4874.3	16	9233	0	85	185–835
1121	409734	4461185	4862.7	16	8248	0	190	190–293, 293–340, 395–405
1122	410520	4461964	4867.6	16	11,484	0	82	222–440, 985–995, 1045–1075, 1440–1485, 1844–2070
1123	409739	4461587	4867.7	16	8632	0	152	278–955
Fairfield Spring	406467	4457211	4895.0	na	11,446	0	na	na

UTM coordinates in meters, NAD27 datum.
conf. = Well is completed in the confining unit. No drawdown expected or observed.
na = not applicable

One other production well in the area large enough to cause potential interference in the drawdown response of the White wells is the White Hills community well (ID 1033 on figure 2), which pumped intermittently during the aquifer test. The actual pumping rate is not known, but the well served 112 mostly residential water connections in 2005 (Utah Division of Water Rights, 2009b) on ¼- to ½-acre (0.1–0.2 ha) lots (area estimated based on the 2006 aerial photograph). The total production from the White Hills community well in 2005 was reported as 79 acre-feet (0.097 hm^3) (Utah Division of Water Rights, 2009b)—a reasonable volume considering the number of households, average water used per household in Utah (Utah Division of Water Resources, 2010), and the lot sizes in White Hills. I assumed the volume produced by 1033 during the aquifer test (necessarily less than the 2005 total of 79 acre-feet [0.097 hm^3]) would have a negligible influence on drawdown in observation wells compared to the 721 acre-feet (0.89 hm^3) produced by the White wells from July through October 2005 (Utah Division of Water Rights, 2009a).

Irrigation lines discharged the water produced by the White wells during the test onto adjacent alfalfa fields downgradient (east) of the wells. The potentiometric surface is approximately 35 to 90 feet (11–27 m) below land surface under the fields and the aquifer is confined. Irrigation water may have leaked through the confining layer and recharged the aquifer, effecting less drawdown at the downgradient wells, as discussed below.

Drawdown Response

Drawdown response in wells was varied (figure 5). Pumping the White irrigation wells caused drawdown in pumping well 44 (figure 5A) and four of the observation wells (figure 5B and C) but not in most of the wells south and east of the pumping wells (figure 5D and E). Variations in pumping rate had a pronounced effect on drawdown in 44, where 45.5 feet (13.9 m) of drawdown was observed on day eight when both the downhole pump and the inline booster pump were operating (figure 5A), but only 26.5 feet (8.08 m) of drawdown was observed on day 69 when only its downhole pump was operating. Drawdown in the observation wells was between 0 and 1 foot (0–0.3 m) after 21 days of pumping and between 0 and 3 feet (0–1 m) after 69 days of pumping (figure 5B though E). Four of the observation wells, UGS IDs 9, 22, 49, and 69, are small-diameter domestic or stock supply wells. Each of these wells was pumped at least once during the test, which drew down the water level in that

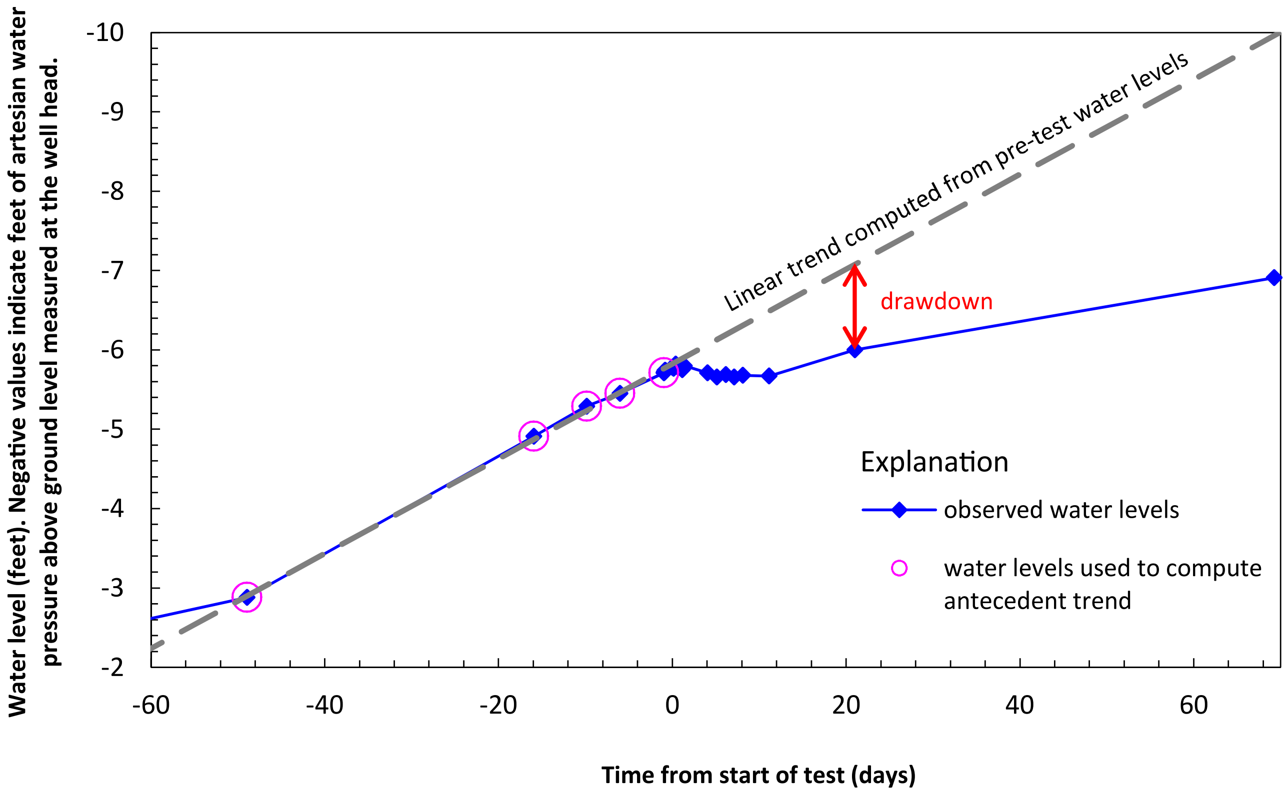


Figure 3. *Potentiometric head in well 1037 before and during aquifer test, showing method to correct for the antecedent water-level trend.*

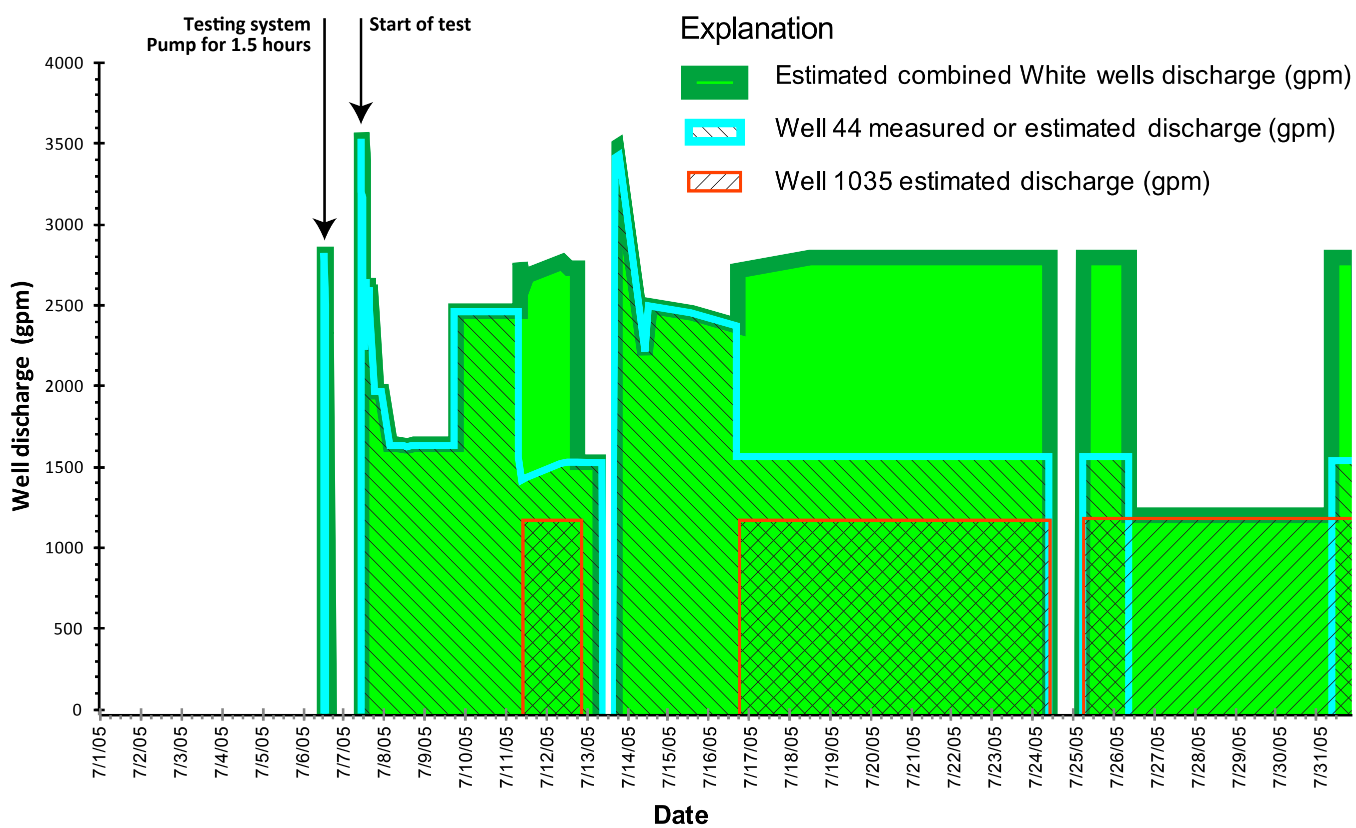


Figure 4. *Well discharge during the White wells aquifer test. The need for more or less irrigation caused variable discharge rates throughout the test, including two periods having zero discharge: (1) six hours on July 13 and (2) approximately 20 hours on July 24 through 25. Detailed flow records were not kept after July 31, 2005.*

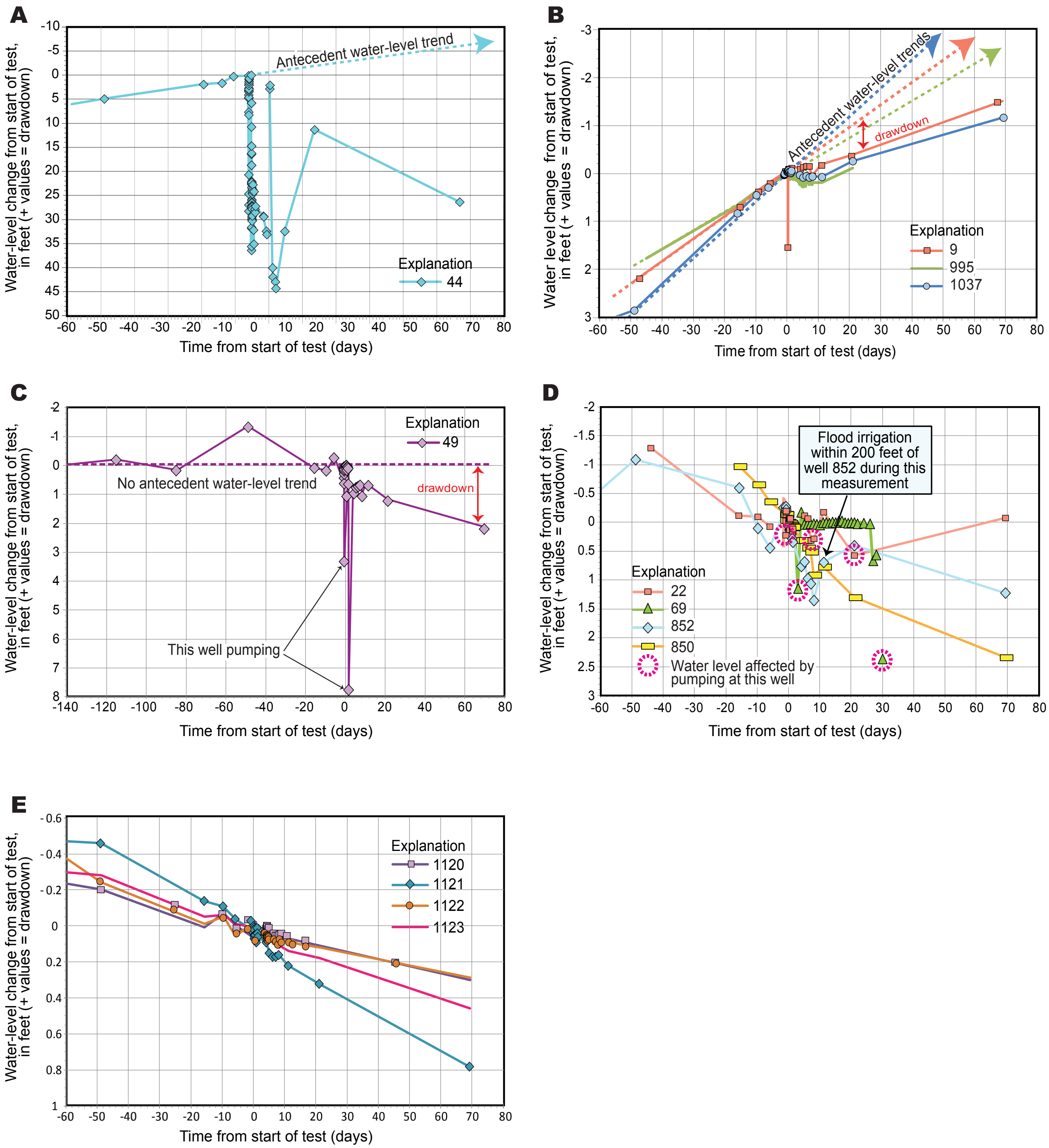


Figure 5. *Potentiometric responses in pumping (A) and observation wells (B–E) during the White wells aquifer test showing antecedent water-level trends, if present, drawdown, and irregularity of water levels. Water level in pumping well 44 (A) is affected by shut down periods, operation of a booster pump, and well 1035 pumping. Drawdown in wells 9, 995, and 1037 (B) ranged from 0.7 to 1.0 foot as calculated from extrapolated upward antecedent water-level trend. Well 9 pumped intermittently during the test. Calculation of drawdown in a private well located in Fairfield (C) is complicated by pumping the well. Water-level trends in wells south (D) and east (E) of the White wells do not clearly show the effects of pumping the White wells. Wells 22, 69, and 852 (D) are wholly or partially completed in the clay confining unit, and well 852 likely is affected by nearby flood irrigation. Wells east of the pumping wells (E) have long perforated intervals, which could conceal small amounts of drawdown.*

specific well. Because the duration and frequency of pumping are not precisely known, I omitted water levels measured during or immediately after the wells pumped (identified on figure 5) from analysis.

Observation wells 22 and 69 showed no drawdown from the White wells pumping because they are completed in the clay confining unit (figure 5D).

Observation well 852 is located approximately 1.5 miles (2.4 km) east-southeast of the pumping wells and is screened in the clay confining layer from 20 to 80 feet (6–24 m). The downward water-level trend observed in 852 before the test continued for 11 days into the test, after which time water level increased markedly (figure 5D). A field 200 feet (60 m) southwest of the well was being irrigated when the water level rose. This irrigation recharge to the shallow water table in the clay confining layer (depth to water 7 to 10 feet [2–3 m]) likely caused the rising water levels.

I interpret no drawdown at Fairfield Spring based on the lack of observed decrease in discharge at the spring over the period of pumping. Figure 6 shows that the discharge of Fairfield Spring was increasing before and for approximately one month into the period when the White wells were pumping, and again starting in October several weeks after pumping ceased. The generally consistent discharge of approximately 7 cubic feet per second (200 L/s) throughout August and September (figure 6) may be interpreted as a decrease in discharge compared to the expected discharge had the antecedent trend continued; however, natural variation in flow cannot be ruled out. In most years, springtime discharge of Fairfield Spring is usually 1.5 to 2 cubic feet per second (43–57 L/s) higher than fall-time discharge (Jordan and Sabbah, 2012). Jordan and Sabbah (2007) and Jordan and Sabbah (2012) analyzed the relationship of Fairfield Spring discharge to the White wells pumpage, water levels in wells located between the White wells and the spring, and precipitation over 25 years. In those studies, we concluded that Fairfield Spring discharge depends primarily on the potentiometric head in the confined aquifer, which is controlled primarily by recharge to the aquifer. We also concluded that the conditions for direct influence of pumping at the White wells on Fairfield Spring discharge are met only when storage in the aquifer is depleted, because depleted storage conditions allow the wells' radius of influence to reach the aquifer at the spring, decreasing the potentiometric head, and thus, the discharge at the spring. Historically, extended periods of lower than average precipitation and pumping greater than approximately 2000 acre-feet per year (2.5 hm^3/yr) from the White wells has depleted aquifer storage. In the years preceding the White wells aquifer test, precipitation over the study area was below average in the years 2001, 2002, and 2003 (by an average of 16%), but discharge from the White wells averaged only 785 acre-feet per year (0.97 hm^3/yr) (Jordan and Sabbah, 2012, appendices E and F). In 2005, precipitation was above average and discharge was 721 acre-feet (0.89 hm^3); therefore, conditions were not conducive for the White wells to affect Fairfield Spring discharge.

In addition, the sharp rise in spring discharge in mid-July, which occurred after the pumping started, and the nearly one-month lag time between either starting or stopping the pumps and the subsequent change in discharge trend are inconsistent with a negative discharge affect from pumping. Other factors, for example, natural variation in the amount of recharge to the spring area, are the likely cause of the change in spring discharge over summer 2005.

The cone of depression formed by pumping the White wells for 21 days is shown in figure 7. The cone of depression is elongated in the north-south direction and shows drawdown in some wells in the principal aquifer (wells 9, 995, 1037, and 49), but not others (wells 1120, 1121, 1122, 1123, and 850). The elongated shape of the drawdown cone may result from

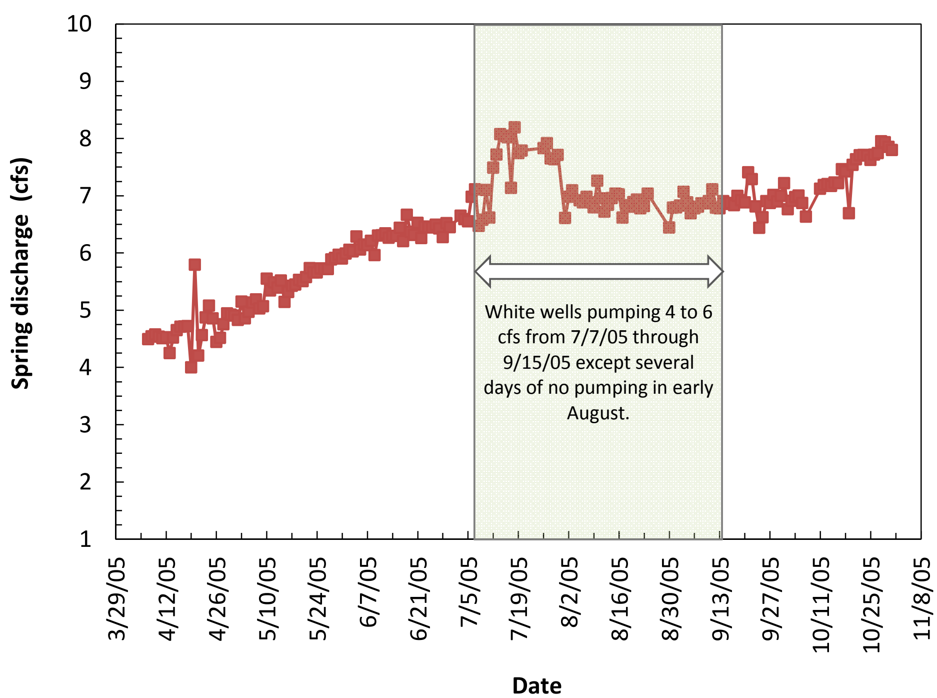


Figure 6. *Fairfield Spring discharge during the White wells aquifer test. Pumping at the White wells causes no apparent decrease in spring discharge during this time period.*

the similarity of aquifer characteristics in the north-south direction. The pumping wells are located in the principal alluvial aquifer, which is overlain by the continuous clay confining layer. The clay layer pinches out to the west, where coarser alluvial fan sediments dominate. Logs from wells 1120, 1121, 1122, and 1123, east of the pumping wells, show considerably more clay than wells farther west. Drawdown is greater in the north-south direction because the confined aquifer is elongated in that direction, becomes unconfined to the west, and becomes finer grained and of lower transmissivity to the east.

Analysis

The first step in aquifer-test data interpretation is to analyze the relationship of drawdown to time while considering the hydrogeologic setting. The characteristic shape of the curves generated by plotting drawdown versus elapsed time using log-log scale and log-linear scale provides clues about the aquifer flow characteristics.

In the earliest part of a drawdown test, the water level in the pumped well will not have drawn down far enough to create a hydraulic gradient into the well that is sufficient to produce the volume of water being pumped, and the well discharge will be almost entirely water that is stored in the wellbore (wellbore storage) (Papadopulos and Cooper, 1967). Wellbore storage effects are more pronounced the larger the pumping well diameter and the lower the aquifer transmissivity (van Tonder and others, 2002, part B, pg. 16; Papadopulos and Cooper, 1967; Moench, 1984). My analysis of the drawdown response in pumping well 44 used the Theis confined aquifer solution (Theis, 1935) with correction for wellbore storage (Papadopulos and Cooper, 1967). I used AQTESOLV PRO v. 4.5 computer software (Duffield, 2007) to match the analytical aquifer solution to the drawdown data. Curve matches from the software program are shown in figure 8.

Curve matching 44's drawdown data produced a transmissivity estimate of 9100 feet squared per day (850 m^2/d). The thickness of the aquifer over which 44 is screened is 150 feet (46 m), and the thickness of similar sediments according to the driller's log is 417 feet (127 m) (appendix B). I used these thicknesses to calculate a range of hydraulic conductivity for the aquifer at the White irrigation wells from 22 to 61 feet per day (6.7–19 m/d). Because of added drawdown in the pumping well due to well screen inefficiency and mechanical well loss, these values are probably lower than the actual transmissivity and hydraulic conductivity of the aquifer.

I used the Theis confined aquifer solution (Theis, 1935) with correction for partial penetration (Hantush, 1961) where necessary to analyze data from the four observation wells that showed drawdown. Transmissivity values ranged from 62,700 to 115,000 feet squared per day (5830–10,700 m^2/d) and hydraulic conductivity ranged from 150 to 280 feet per day (46–85 m/d) assuming an aquifer thickness of 417 feet (127 m). These values are much higher than the value calculated at the pumping well, and are in the typical range for well-sorted sand and gravel (Fetter, 1988, p. 80). Drillers' logs of each well show interbedded clay and gravel layers consistent with their location near the intersection of the alluvial fans on the western margin of Cedar Valley and the finer sediment found in the center of the basin. The high transmissivity values are due to the small amount of observed drawdown, which could be a result of (1) unknown sources of recharge to the aquifer in the vicinity of the observation wells, (2) a semi-permeable barrier to groundwater flow between the pumping and observation wells, which would result in more drawdown on the pumping side of the barrier and less drawdown on the non-pumping side as compared to a homogeneous and areally extensive aquifer, and/or (3) leakage from overlying or underlying units. Leaky aquifer solutions produced similar transmissivity values to non-leaky solutions.

Calculating average values of aquifer parameters for the western Cedar Valley aquifer using values that likely are too low, as in the case of well-efficiency influence at the pumping well, or too high, as suspected based on lithology in the distal observations wells, would be inaccurate. Rather, using the analyses from the pumping and observation wells as low and high brackets, I estimate aquifer transmissivity in the western Cedar Valley basin-fill aquifer to be between 18,000 feet squared per day (twice the value from the pumping well) and 32,000 feet squared per day (approximately half of the two lower estimates from the observation wells). The median value of this range, 25,000 feet squared per day (2300 m^2/d), is given on table 1. A rough estimate of the hydraulic conductivity using an aquifer thickness of 417 feet is 60 feet per day (18 m/d).

Storativity estimated from the four observation wells that showed drawdown ranged from 0.006 to 0.03. The lower end of this range is typical of confined aquifers (Fetter, 1988, pg. 107) and was determined from the two wells near Fairfield, where the aquifer clearly is confined as shown by artesian wells. The higher end of the storativity range is more typical of a fine-grained unconfined aquifer (Fetter, 1988, pg. 107), and was determined from two wells west and north of the pumping well, near the edge of the confining unit and where the confined or leaky nature of the aquifer is not well understood due to lack of subsurface data.

The calculated derivative of the drawdown can be useful in aquifer-test analysis to determine the type of aquifer, the time when radial flow dominates, and the presence of aquifer boundaries (Renard and others, 2009). Unfortunately, the derivative calculated from time and drawdown data for the White wells test had significant noise due to pumping rate variability and the small magnitude of drawdown and large time duration between water-level measurements in late-time data; therefore, I was not able to use derivatives to aid my analysis.

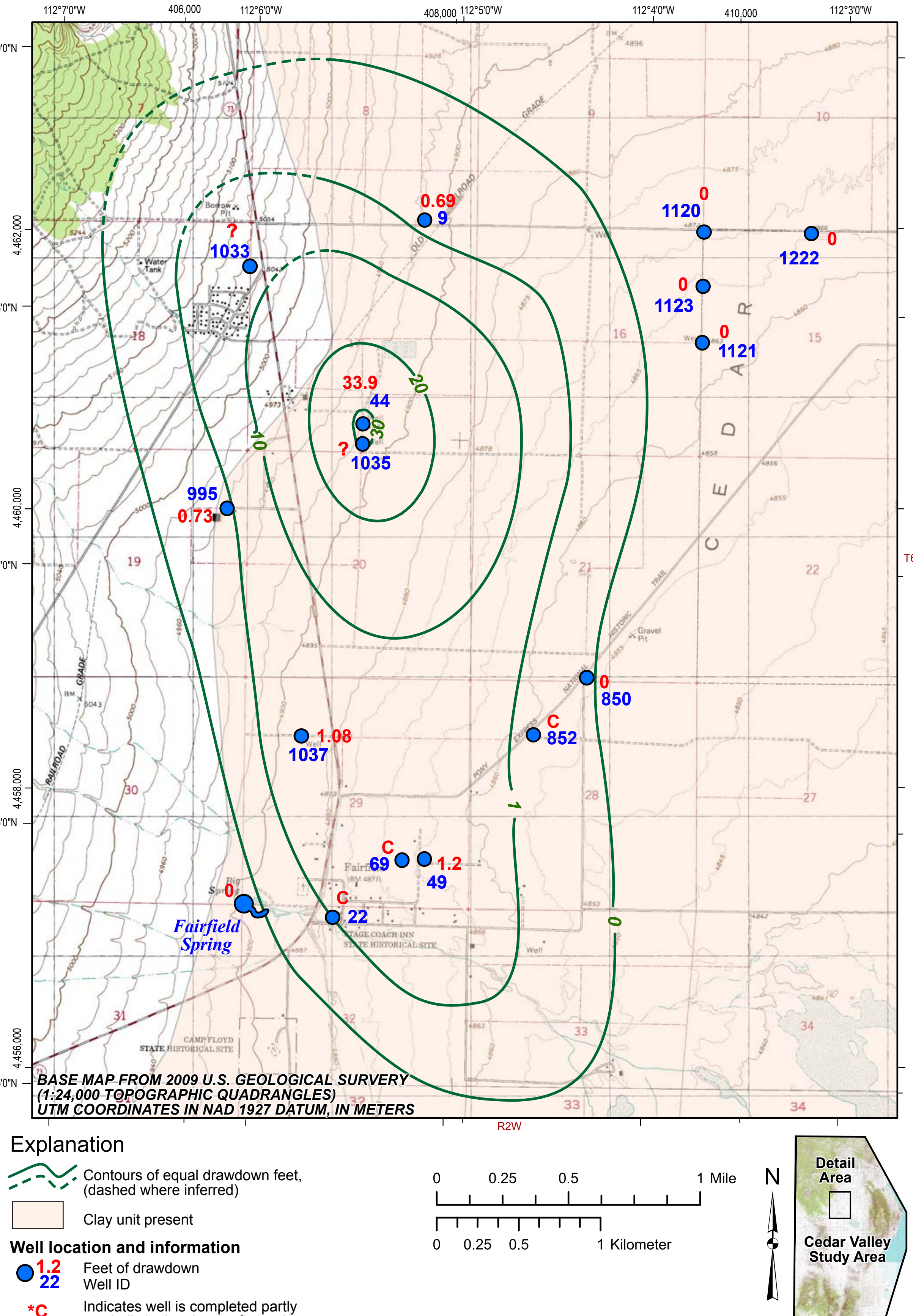
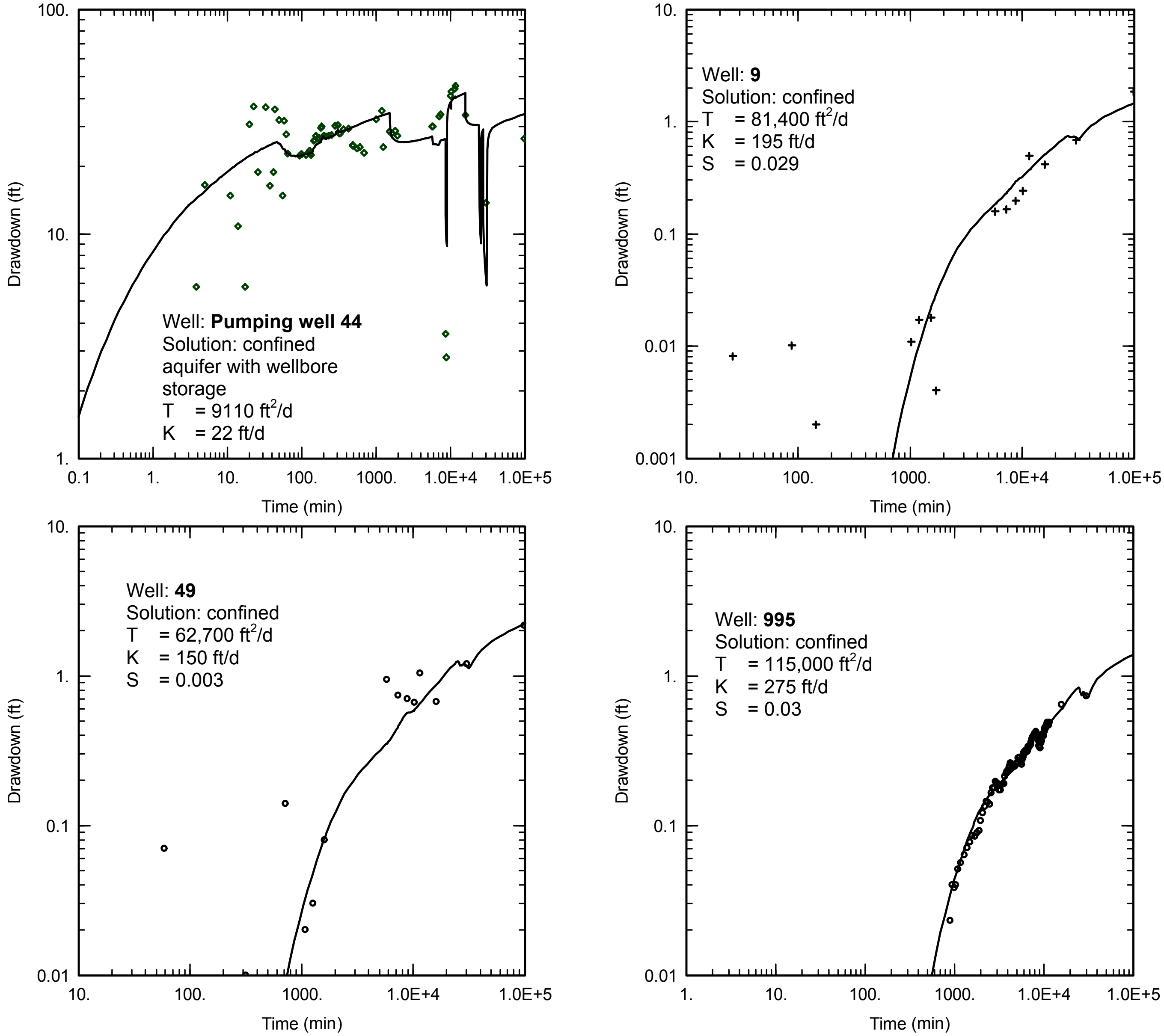


Figure 7. *Cone of depression formed during the White wells aquifer test, western Cedar Valley. Corrected drawdown at day 21 (day 16 in pumping well 44) is contoured.*



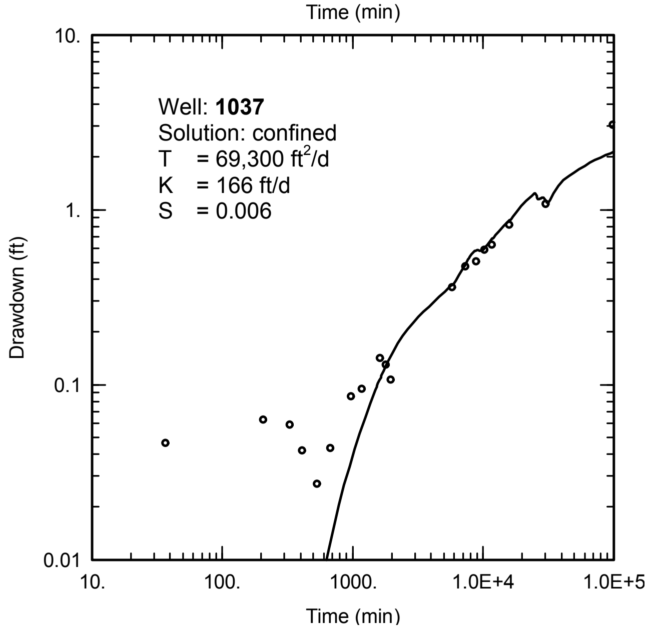


Figure 8. *Analytical solution type curves (solid lines) matched to aquifer-test data from the White wells aquifer test (symbols). Aquifer transmissivity and hydraulic conductivity estimated using data from observation wells are higher than can be reasonably expected for the interbedded gravel and clay aquifer. T = transmissivity, K = horizontal hydraulic conductivity, and S = storativity.*

BASIN-FILL AQUIFER TEST ON A MEDIUM-DIAMETER WELL

In September 2005, I coordinated a 48-hour single-well aquifer test on a 10-inch diameter (25-cm) privately-owned well located on a mink farm in north-central Cedar Valley. The well, UGS ID 991, is in the southeast quarter of section 23, Township 6 South, Range 2 West, Salt Lake Base Line and Meridian (SE1/4 sec. 23, T. 6 S., R. 2 W., SLB&M) (figure 9). Because the driller's log contains little stratigraphic information, stating that 991 is screened from 120 to 400 feet (37–122 m) below land surface in "interbedded layers," UGS staff ran a downhole gamma-ray geophysical log on the completed well before permanent pump installation to learn more about the aquifer surrounding the well. Both logs are included in appendix B. The gamma-ray log reveals a clay unit from the surface to approximately 70 feet (21 m), one thin sand unit from 70 to 78 feet (21–24 m), and mostly silty clay below 78 feet (24 m). The static water level in 991 is approximately at the bottom of the upper clay unit. The aquifer is confined during static conditions but dewaters near the well during pumping. The well partially penetrates the basin-fill aquifer, defined for aquifer-test analysis at this location as from the bottom of the clay unit at 70 feet (21 m) to the bottom of the screen interval at 400 feet (55 m).

Because the aquifer test was conducted at the end of the irrigation season, 991 had been in use for three weeks prior to the aquifer test to irrigate an alfalfa field 250 feet (76 m) south of the well (figure 9). A field northwest of the well was irrigated in the weeks prior to the test by a different well, and a sod farm located ½ mile southwest of 991 was irrigated all summer long and up to a week before the test by another well. Although none of these areas were being irrigated during the test, seepage from the summer irrigation season was likely entering the aquifer during the drawdown test through the continuous clay layer present from the surface to 70 feet (21 m) below surface. Discharge during the test was released onto the alfalfa field 250 feet (76 m) south of the well, which likely provided a constant head in the overlying confining unit. Although the driller's log for the test well (appendix B) only describes the lithology at the site to 430 feet (131 m), logs of wells in the vicinity indicate that the clay content increases with depth in the basin-fill aquifer. Leakage from permeable units upward through clay interbeds to the producing interval of the well is likely. At this location, the basin-fill aquifer acts as a confined aquifer with leaky confining units above and below the aquifer.

Although the pump in 991 was shut down 27 hours before the aquifer test to allow the water level to recover, an upward trend is apparent in the pre-test water-level data. The upward trend indicates that the water level in the well did not fully recover from pumping before I started the test and/or that the water level was being affected by irrigation seepage. Drawdown and recovery data were corrected for the upward trend by adding the difference between the water level at the start of the test and a projected water level at the time of observation. I calculated the projected water level using a linear regression calculated from the water levels measured in the two hours before the test began, as shown on figure 10. Other projection methods (exponential, logarithmic, and linear regression on all the pre-test water levels) did not fit the data as well as the linear trend. The chosen method assumes the factors affecting water level in the two hours prior to the test continued affecting water level at the same rate during the test, which would not be the case for recovery (water-level correction for recovery would lessen with time), but likely would predict background or season-long irrigation seepage effect. The correction amounts to an additional 5% drawdown in comparison to uncorrected data (2.56 feet [0.78 m] correction on 47.80 feet [14.57 m] of measured drawdown) over the 48-hour test.

A corrected maximum drawdown of 50 feet (15 m) was observed in 991 as it pumped at a nearly steady 215 gallons per minute (13.6 L/s). No observation wells were available for this test. Flow was measured periodically throughout the test by measuring the time it took to fill a 55-gallon barrel. Water levels collected for this test are tabulated in appendix A, table A-2.

Well discharge in the early part of an aquifer test has a component of wellbore storage. The effect of wellbore storage on drawdown, which manifests as less observed drawdown than theoretically predicted, is greater in situations of low aquifer transmissivity and large-diameter wells (Papadopulos and Cooper, 1967). Using a simple equation relating well diameter and the transmissivity of the aquifer to approximate length of time over which wellbore storage affects the drawdown curve (Papadopulos and Cooper, 1967), I estimate that the effect of wellbore storage lasted approximately an hour into the aquifer test, after which time I expect the drawdown response to conform to type curves for specific aquifer solutions.

In aquifer-test analysis, the derivative is a measure of the change in the rate of drawdown and is defined as the change in drawdown with respect to the change in the natural logarithm of the elapsed time (Renard and others, 2009). When examined with type curve plots for drawdown versus time, derivative versus time plots can aid in determination of the type of aquifer (confined, unconfined, leaky, fractured, etc.) involved in the test and if drawdown is affected by a boundary or wellbore storage (Renard and others, 2009) (figure 11). The shape of the derivative curve supports my interpretation of a leaky confined aquifer displaying wellbore storage.

I applied the leaky aquifer solution of Moench (1985), which accounts for leakage from irrigation seepage, wellbore storage, and wellbore skin (area around the well having altered

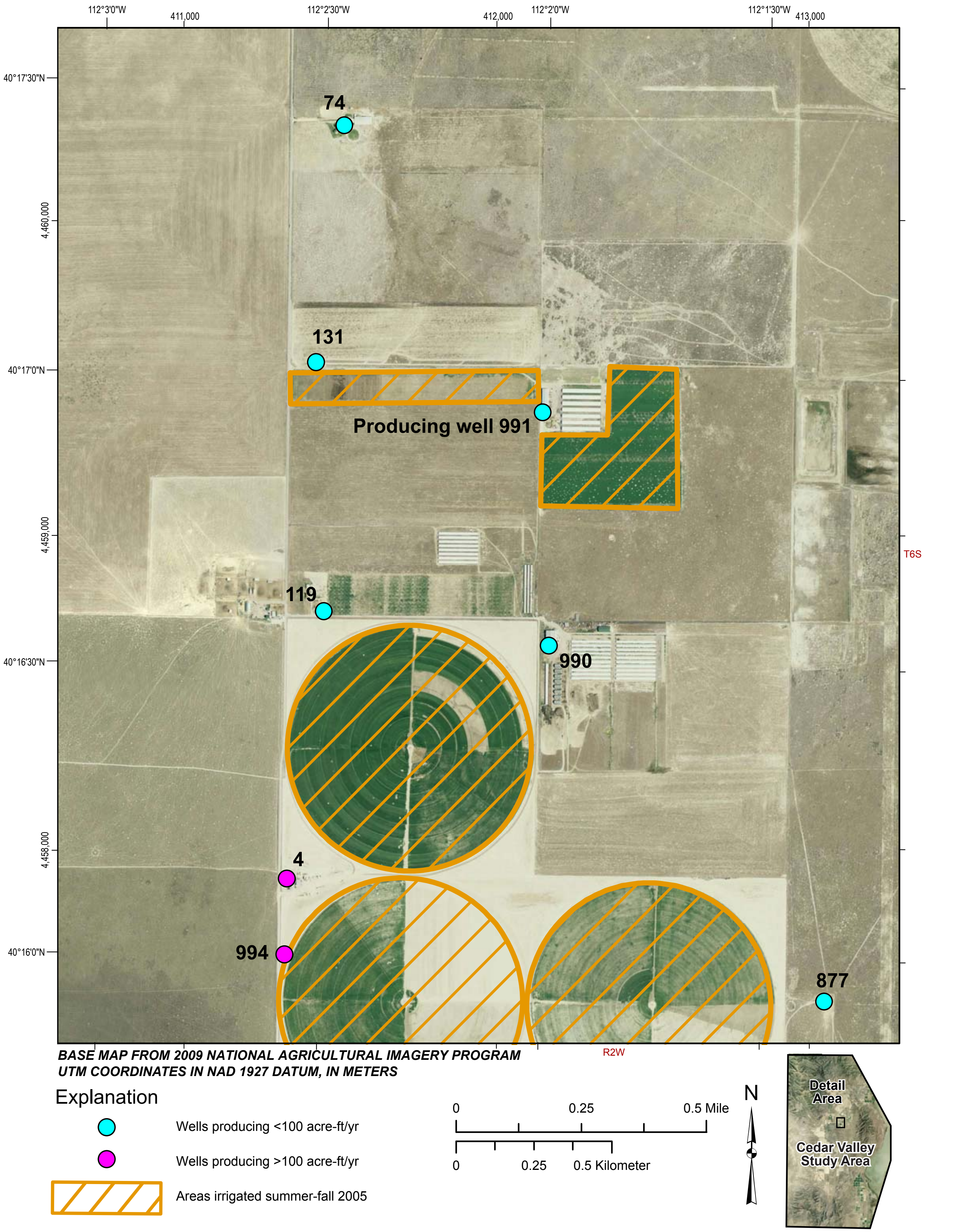


Figure 9. *Well locations and irrigated areas involved in an aquifer test on a medium-diameter well in basin fill.*

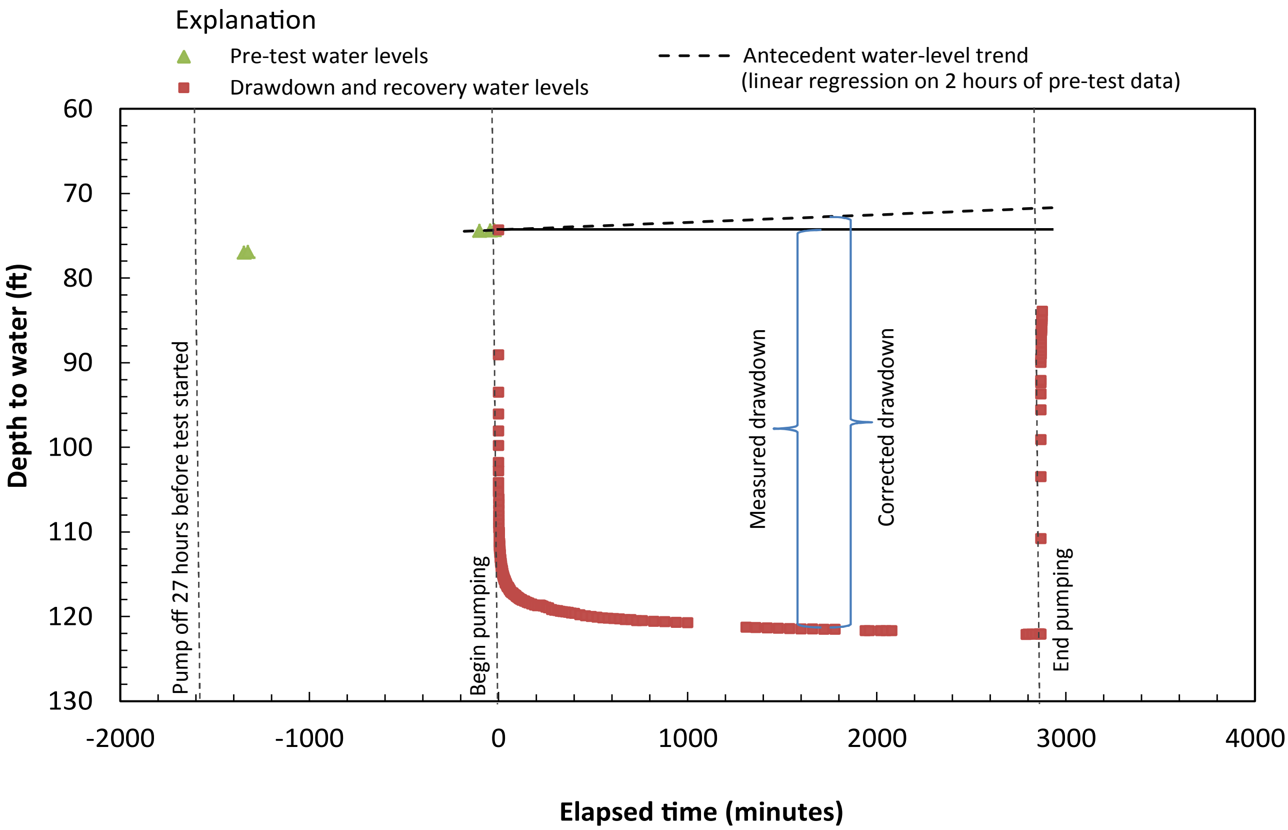


Figure 10. *Water-level response in the medium-diameter well before and during the aquifer test, showing method to correct for an antecedent water-level trend.*

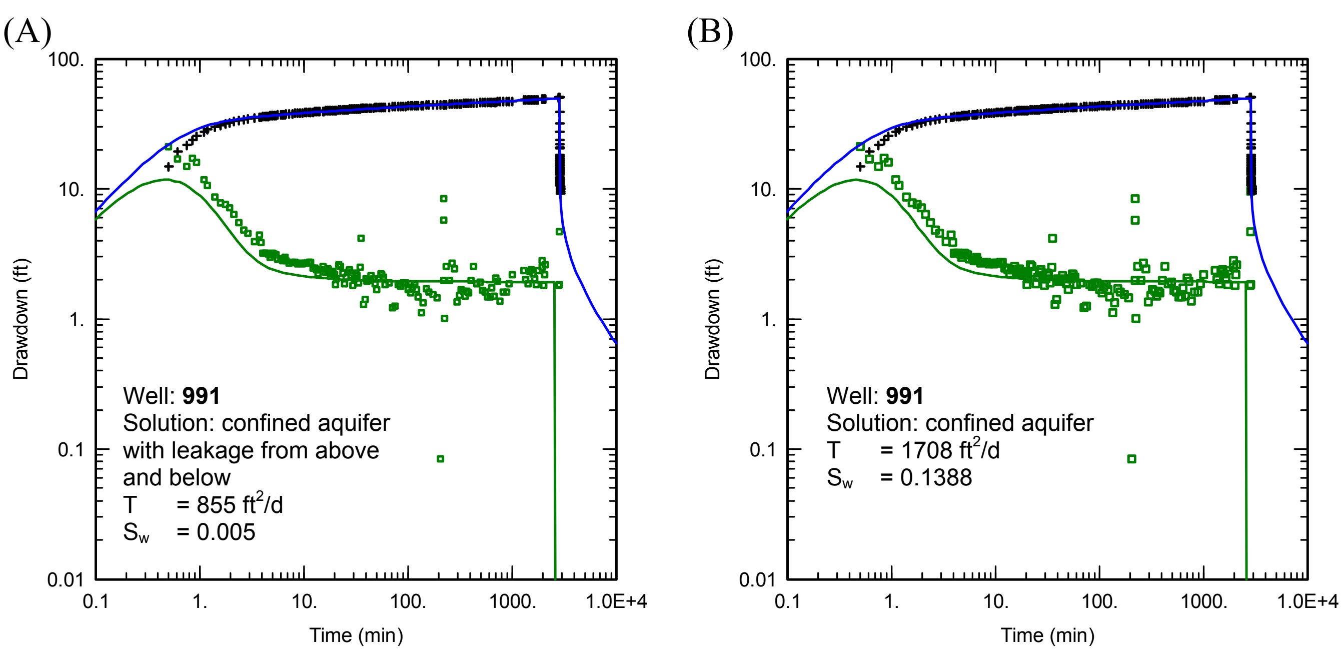


Figure 11. *Analytical solution type curves (blue curves) and derivative curves (green curves) matched to water-level data from a medium-diameter basin-fill well (black crosses) and derivative of the drawdown (green squares) using (A) a leaky aquifer solution with storage in the confining units (Moench, 1985) and (B) a confined solution accounting for partial penetration (Dougherty and Babu, 1984). T = transmissivity,* S_w *= wellbore skin factor.*

permeability), but does not account for partial penetration of the aquifer by the well. I used AQTESOLV (Duffield, 2007) computer software to apply case 3 of Moench's (1985) leaky aquifer solution, which configures the aquifer being tested as having confining units above and below (with storage in those confining units) and a constant head in an aquifer overlying the upper confining unit (to best simulate irrigation seepage from the near surface layers through the confining unit). The calculated transmissivity using the leaky aquifer solution is approximately 855 feet squared per day (79.4 m^2/d) (figure 11), which is likely lower than the actual transmissivity because the solution does not account for partial penetration. Partial penetration of a well into an aquifer results in increased drawdown in the well; however, the solution used assumes all observed drawdown is the response of the leaky aquifer in a fully penetrating well. Smaller true drawdown (without partial penetration) in the aquifer would yield a larger transmissivity estimate.

Next, I applied a confined-aquifer solution (Dougherty and Babu, 1984), even though the geologic model does not fit as well as a leaky aquifer solution, because the confined solution accounts for partial penetration of the pumping and observation wells, wellbore storage, and wellbore skin. The confined solution yielded an aquifer transmissivity of approximately 1700 feet squared per day (160 m^2/d) (figure 11), which is likely higher than the actual transmissivity because the solution does not account for leakage from a confining unit. Leakage from a confining unit would decrease observed drawdown; however, the solution assumes all drawdown is from a confined, non-leaky aquifer. Larger drawdown (as would be the case if there was no leakage) in the aquifer would yield a smaller transmissivity estimate.

I estimated the transmissivity of the basin-fill aquifer at this location as slightly higher than the 855 feet squared per day (79 m^2/d) predicted using the leaky aquifer solution, but lower than the 1700 feet squared per day (158 m^2/d) confined aquifer solution result, or probably in the range of 1000 to 1200 feet squared per day (90–110 m^2/d). Using an aquifer thickness measured from the bottom of the confining unit to the bottom of the screen interval, or 330 feet (100 m), the hydraulic conductivity is approximately 3.0 to 3.6 feet per day (0.9–1 m/d), a value typical of fine sand (Fetter, 1988; Anderson and Woessner, 1992; Kruseman and de Ridder, 2000; Singhal and Gupta, 2010).

The two solutions applied to these test data give wellbore skin factors (S_w) close to zero, which indicates the wellbore skin (the area surrounding the well) is neither much more nor much less conductive than the aquifer. Other parameters estimated by the solutions including storativity, vertical to horizontal anisotropy, and leakage factors are not accurate when using single-well test data.

BASIN-FILL AQUIFER TEST ON A SMALL-DIAMETER WELL

A single-well seven-hour constant-rate drawdown test followed by five hours of recovery data collection was performed on a 6-inch-diameter (15 cm) privately owned domestic well located in Fairfield (figure 12) on May 9, 2006. The duration of the test was shortened from 24 hours to seven hours because drawdown stabilized approximately five hours into the test and sand was entering the well during the last hours of the test. A maximum drawdown of 49.19 feet (14.99 m) was measured in the pumping well as it pumped 31 gallons per minute (2 L/s) (figure 13).

The tested well, UGS ID 49, was drilled in 1977 by cable tool method to a depth of 281 feet (86 m). The unperforated well casing extends to 262 feet (79 m), below which the borehole likely is fully or partially collapsed. The static water level is approximately 6 feet (2 m) below ground level. The well driller's log (appendix B) lists clay layers as much as 34 feet (10 m) thick interbedded with clay-sand-gravel layers from the surface to 248 feet (76 m). The well is open in a unit logged by the driller as containing clay, sand, and gravel from 248 to 281 feet (76–86 m), in which the driller noted "fine layers of gravel". No observation wells were available to monitor during this test. The aquifer is confined, and leakage from the interbedded sediments above the producing unit is likely. Even though the hole likely collapsed below the bottom of the casing, the collapsed portion likely has a transmissivity much larger than the undisturbed sediment, so I assumed the well to be partially penetrating the aquifer from 262 to 281 feet (80–86 m). I used an aquifer thickness from 248 (top of the aquifer unit) to 281 feet (bottom of the borehole) (76–86 m), or 33 feet (10 m), in my analysis.

This domestic well supplies water to one household and residential irrigation system. The well pump was operating to irrigate the lawn and garden six days prior to the test, at which time irrigation was discontinued until the test. Household use of water was limited but ongoing before and during the test. Discharge is the sum of the discharge measured at five outdoor hydrants by recording the time it took to fill a 5-gallon bucket at each hydrant. I measured discharge periodically, and never during periods when indoor water use would have been diverting flow from the hydrants. Water was discharged onto the lawn and garden areas surrounding the house during the test. It is unlikely that the small discharge of the well (31 gallons per minute [2 L/s]) would have infiltrated through 248 feet (76 m) of interbedded clay, sand, and gravel to the aquifer producing water to the well. No large-scale irrigation was present on adjacent residential lots or farm fields.

UGS personnel measured water levels using an electronic water-level sounder. Background water levels fluctuated by 0.68 foot (0.21 m) during the eight days prior to the test but no upward or downward antecedent trend was apparent.

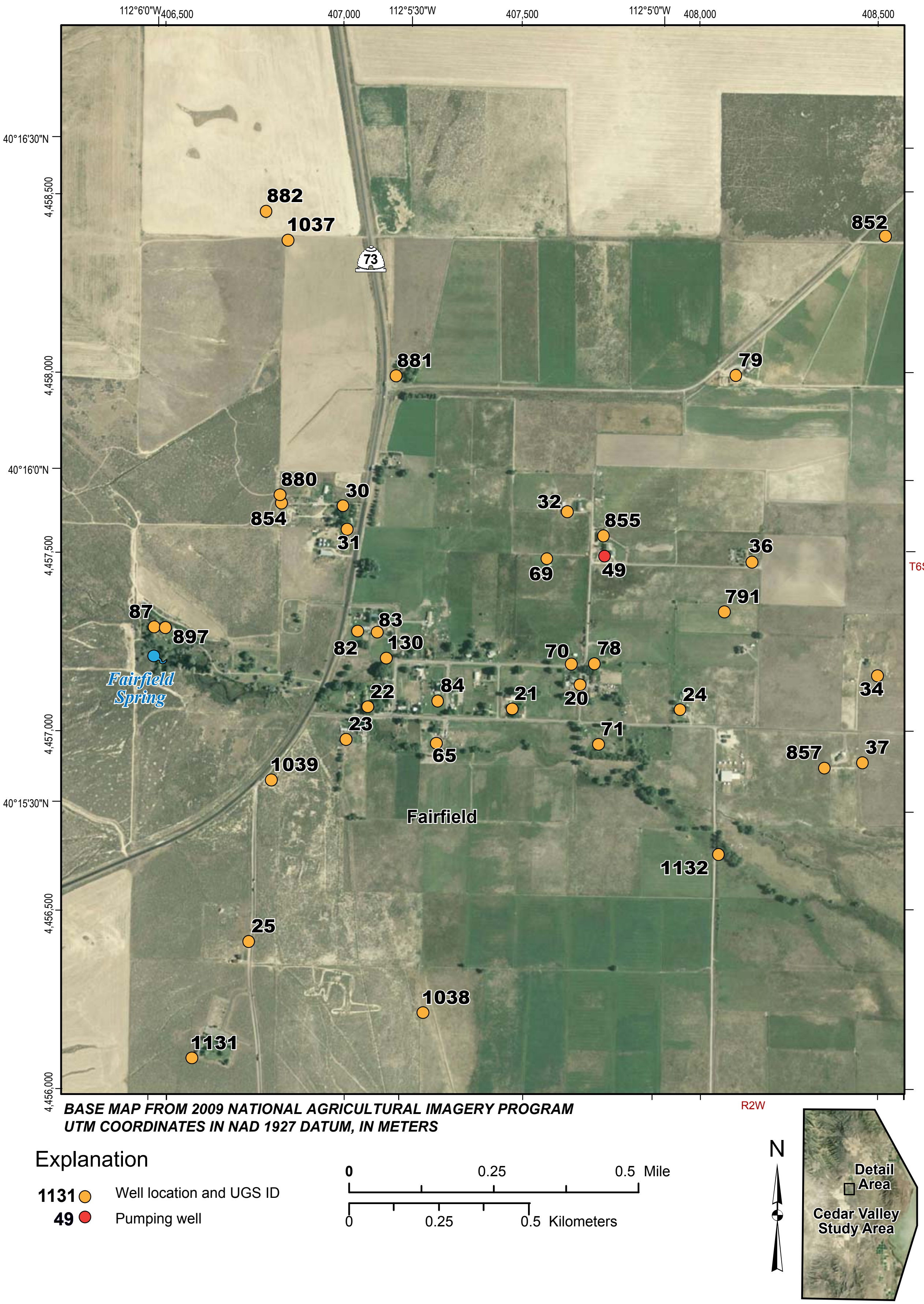


Figure 12. Location of the small-diameter domestic well involved in a basin-fill aquifer test.

Water-level response was typical for a leaky aquifer with the exception of a small rebound of the water level at approximately 100 minutes into the test (figure 13), which does not coincide with any sizable decrease in measured discharge. It is possible that the measurement technique did not detect a change in discharge or that a pressure tank in the domestic water supply system masked a change in discharge. The maximum drawdown of 49.19 feet (14.99 m) was observed at 267 minutes (4 hours 45 minutes) into the test, after which time drawdown varied by less than 1 foot for the remaining 2.5 hours. Water level recovered to within 1.2 feet (0.37 m) of background water level five hours after pumping stopped. Water levels collected for this test are tabulated in appendix A, table A-3.

I analyzed the aquifer-test data using AQTESOLV version 3.5 (Duffield, 2003) and two different leaky aquifer solutions, which compensate for the effects of partial penetration or wellbore storage (Hantush and Jacob, 1955; Hantush, 1964; Moench, 1985) (figure 14). Wellbore storage affected approximately the first 300 minutes of the test, based on analysis using Papadopulos and Cooper (1967). Partial penetration effects are also significant in this analysis because the well only penetrates 60% of the thickness of the aquifer. Well inefficiency may have contributed to drawdown observed in the well. The Hantush solution (Hantush and Jacob, 1955; Hantush, 1964) (figure 14A), while not correcting drawdown in the early time for wellbore storage, corrects for partial penetration and is a close match in late-time data. The Moench solution (Moench, 1985) (figure 14B) provided slightly higher aquifer parameter estimates. The calculated transmissivity, T, of the shallow part of the basin-fill aquifer near Fairfield using these aquifer test data is 70 feet squared per day (6.5 m^2/d) and the hydraulic conductivity, K, is 2 feet per day (0.6 m/d). Storage and leakage factor estimated by curve matching are not accurate when using data from a single well.

OQUIRRH GROUP FRACTURED-ROCK AQUIFER TEST

At the request of the Utah Division of Water Rights, the UGS and the city of Eagle Mountain conducted a five-month constant-rate aquifer test and a 35-day recovery test on Eagle Mountain Municipal Supply Well 3 (herein referred to as "Well 3" but also given UGS ID 992) over the summer of 2007. The purpose of the test was to determine the transmissivity, storativity, and anisotropy of the fractured-bedrock aquifer at Cedar Pass while pumping for an extended period of time (153 days) at this well's typical flow rate of approximately 1930 gallons per minute (122 L/s). The geologic formations comprising the bedrock aquifer at this location are the Butterfield Peaks Formation and West Canyon Limestone of the Pennsylvanian-aged Oquirrh Group. Jordan (2009) reported the results of the test to Eagle Mountain.

Well 3 is at the northern end of the Lake Mountains in northwestern Utah County (figure 15) in the NW1/4 sec. 30, T. 5 S., R. 1 W., SLB&M. The driller's log of Well 3 (appendix B) shows that the well has multiple, separate sections of 16-inch-diameter (41 cm), 0.08-inch-slot (0.2 cm), stainless steel, wire wrap screen between 700 and 1210 feet (213–369 m) below ground surface (figure 16, table 3). The observation wells for

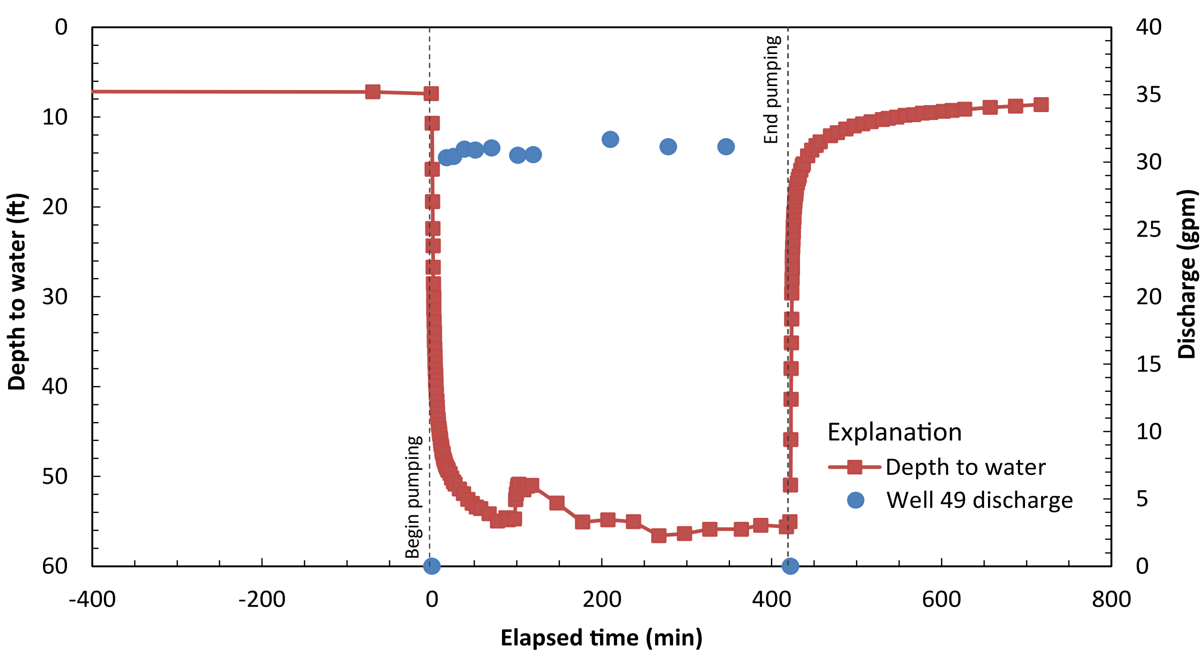


Figure 13. *Water-level response and measured discharge in the pumping well during the basin-fill aquifer test on a small-diameter well in Fairfield.*

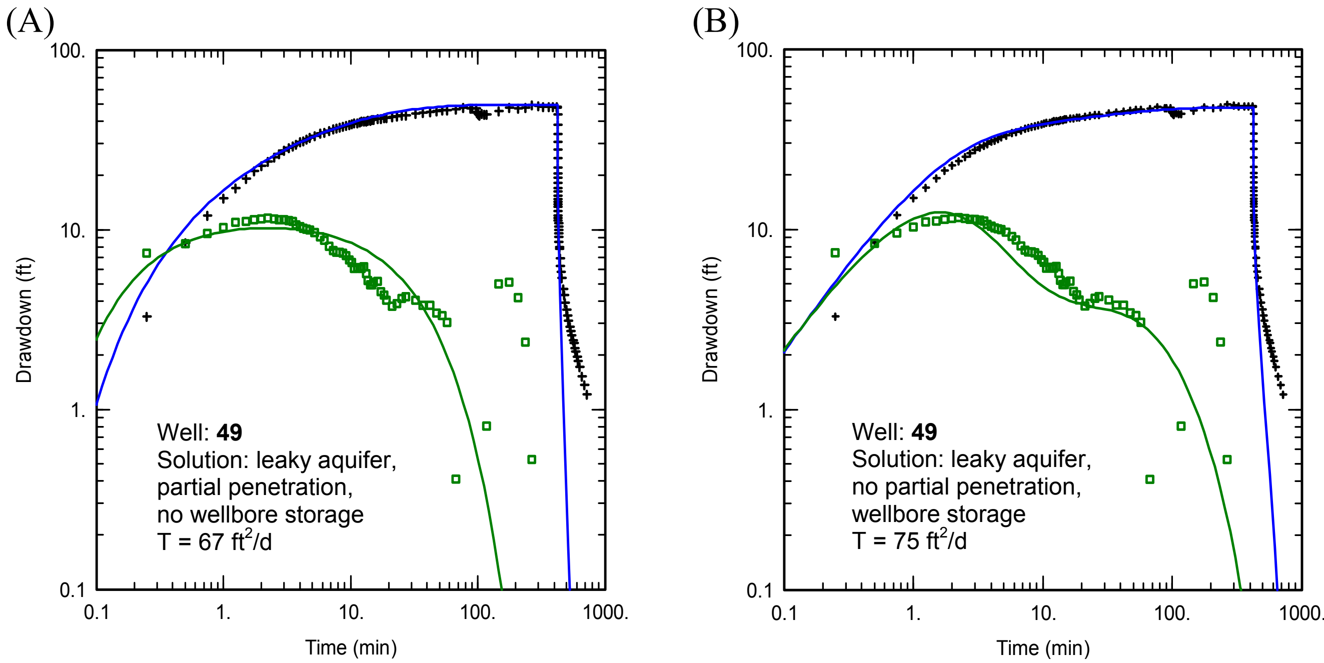


Figure 14. *Analytical solution type curves (blue curves) and derivative curves (green curves) matched to aquifer-test data (black crosses) and derivatives (green squares) from a small-diameter basin-fill well using (A) a leaky aquifer solution with no storage in the confining unit (Hantush, 1964; Hantush and Jacob, 1955), which accounts for partial penetration but not wellbore storage, and (B) a leaky aquifer solution with storage in the confining units (Moench, 1985) which accounts for wellbore storage but not partial penetration.*

this test were MW1 (UGS ID 901), MW2a&b (two separate wells in one borehole, UGS IDs 902 and 903, respectively), and a private well given the UGS identification number 807. Well 5 (UGS ID 983) is an Eagle Mountain production well, which was drilled after the aquifer test was completed. Well completion details and distances from the pumping well are provided in table 3 and locations are shown on figure 15.

Geologic Setting

The lower part of the Middle- to Lower-Pennsylvanian Butterfield Peaks Formation of the Oquirrh Group is exposed at the northern end of the Lake Mountains (Biek, 2004) (figure 15). The Butterfield Peaks Formation is composed of interbedded, fine-grained calcareous sandstone, medium-gray, fine-grained sandy limestone, and minor orthoquartzite. A large syncline runs the length of the Lake Mountains, but smaller folds are also present. Well 3 and MW1 were sited on the axis of a small anticline striking northwest-southeast. Lithology of the Butterfield Peaks Formation in Well 3, MW1, and Well 5 is tan to gray calcareous sandstone and silty limestone grading into gray to black limestone and shaly limestone of the Pennsylvanian West Canyon Limestone below approximately 700 to 800 feet (210–240 m) (appendix B; Utah Division of Water Rights, 2007; Jordan, 2008). The wells are each screened over this gradational contact and through the West Canyon Limestone (figure 16, table 3), and static water level in the wells before testing was approximately 490 to 500 feet (149–152 m) deep.

MW2, a double-completion monitoring well, is located almost 1 mile northwest of Well 3, where unconsolidated basin-fill deposits are present from the surface to 220 feet (67 m) below ground. The shallow completion, MW2a, monitors water level in volcanic rock, which I interpret to be part of the Tertiary flows and ash flow tuffs cropping out in the Traverse Range (Jordan, 2008) (figure 16). MW2b is completed in the underlying Butterfield Peaks Formation (figure 16).

Well 807 is an unused private water-supply well located 4540 feet (1384 m) northeast of Well 3. Based on the surrounding outcrop and dip of the strata (Biek, 2004) and a driller's log for this well documenting limestone, 807 is most likely completed in the West Canyon Limestone.

I estimated aquifer thickness at Well 3 based on the geologic log to be 530 feet (162 m). I calculated the thickness by adding the length of the screened intervals (360 feet [110 m], cumulative), thickness of units below the water table but above the screened interval of the well that likely have high conductivity based on their lithologic and geophysical properties (140 feet [43 m]), and thickness of limestone below the completed well that likely has similar properties as the lowest screen interval (30 feet [9 m]). The inclusion of strata above the screened interval as part of the aquifer thickness is not standard in aquifer-test analysis, but is justified in this case because observation wells MW2b and 807 displayed response to pumping and are completed above the screen interval of Well 3, demonstrating that the aquifer is well connected vertically.

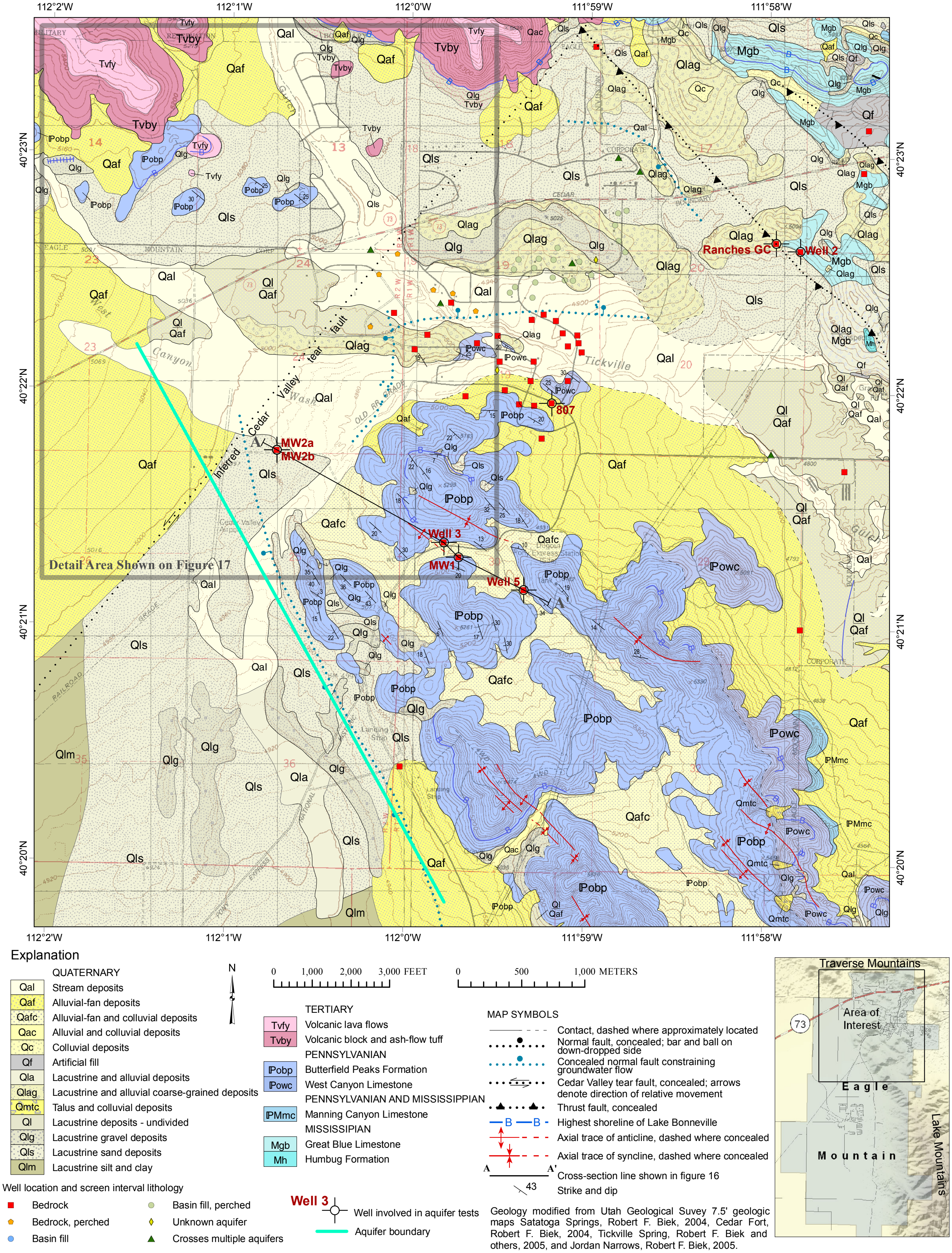


Figure 15. Location of wells and geologic setting of Eagle Mountain fractured-rock aquifer tests.

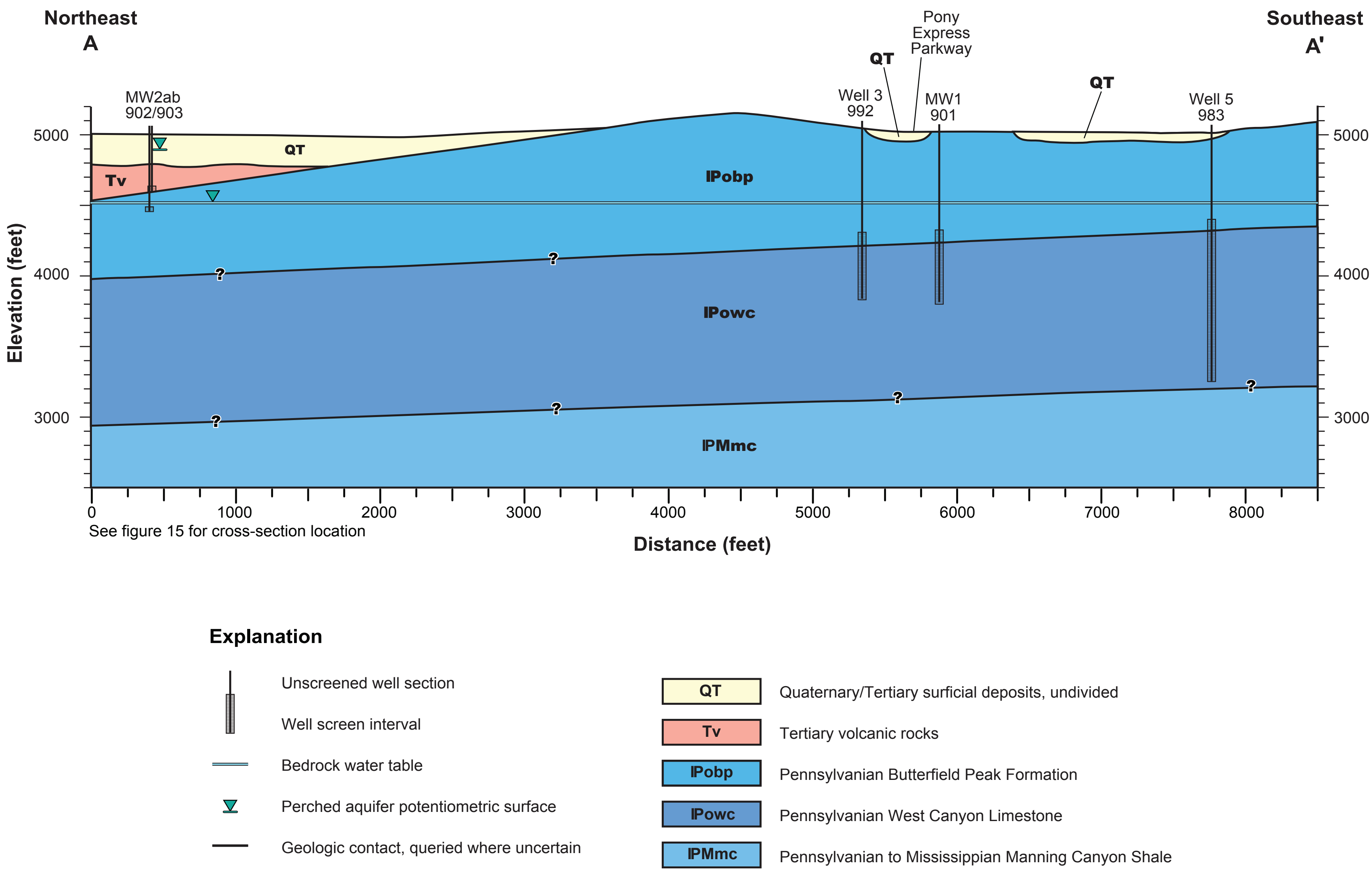


Figure 16. *Schematic geologic cross section through Eagle Mountain monitoring and production wells for an aquifer test on Well 3.*

Table 3. Aquifer and well information at locations involved in the Well 3 aquifer test.

Name (UGS ID)	UTM easting (m)	UTM northing (m)	Land elevation (ft)	Dia-meter (in)	Distance from Well 3 (ft)	Maximum drawdown (ft)	Screen lithology(1)	Screen depth (ft)
Well 3 (992)	415370	4467533	5024.7	16	0	41.2(note 2)	IPobp & IPowc	700–720, 740–780, 790–810, 830–870, 890–910, 930–1050, 1100–1160, 1165–1185, 1190–1210
MW1 (901)	415486	4467418	4999.6	2	537	10.4	IPobp & IPowc	695–705, 725–825, 845–865, 875–1025, 1035–1055, 1065–1205
MW2a (902)	414051	4468259	4989.4	2	4939	0	Tv	345–365
MW2b (903)	414051	4468259	4989.4	2	4939	8.65	IPopb	534–554
807	416223	4468623	4921.3	5	4540	5.95	IPowc	560–647
Well 5 (983)	416001	4467161	4998.3	20	2404	NA	IPobp & IPowc	600–680, 701–801, 843–1003, 1045–1245, 1247–1387, 1429–1469, 1522–1742

UTM coordinates in meters, NAD27 datum.

[1] Screen lithology: IPobp = Pennsylvanian Butterfield Peaks Formation of the Oquirrh Group, IPowc = Pennsylvanian West Canyon Limestone of the Oquirrh Group, Tv = Tertiary volcanic rocks.

[2] Drawdown in Well 3 actually increased to 45.5 feet in the last 1.5 minutes of the pumping period due to the increase in discharge as the well pumped to waste before shutting down.

Fracture permeability is more likely to be greater in the direction parallel to the axes of folds and parallel to bedding, in a northwest to southeast direction at this location. Observation wells MW1 and MW2b are located on the axis of the suspected higher permeability direction, whereas 807 is perpendicular to this axis (figure 15).

The aquifer lacks a well-defined confining unit below the potentiometric surface, which would simplify its classification as a confined aquifer. The significant depth to water and presence of fine-grained units above the water table that could limit atmospheric connection between the land surface and the water table lead me to classify the aquifer as a semiconfined, fractured, sedimentary-rock aquifer.

There is hydrogeologic evidence for aquifer boundaries. The potentiometric surface in the Cedar Valley basin-fill aquifer just west of Cedar Pass is approximately 200 feet (61 m) higher than the potentiometric surface in the Oquirrh Group fractured-rock aquifer (Jordan and Sabbah, 2012), suggesting that a barrier is hindering flow out of the basin fill to the bedrock. A wedge of Tertiary volcanic rock, logged at 230 feet (70 m) thick during UGS-directed drilling of MW2a&b (Jordan, 2008), is sandwiched between the underlying Paleozoic bedrock and the overlying basin fill, in a position to act as a barrier to groundwater flow (figures 16 and 17). After developing the monitoring well by air-lifting water from the well, the water levels in MW2a recovered slowly (rising 36 feet in six weeks), indicating the volcanic unit is not very transmissive. The volcanic unit is positioned at and below the elevation of the basin-fill water table, which forces groundwater to flow around it to the south or underneath it into the bedrock. Similarly, it is bounding the fractured-bedrock aquifer in which Well 3 is completed on the west. The subsurface extent of the volcanic rocks is unknown, but is likely as extensive as depicted on figure 17 based on its existence in the subsurface at the MW2 location and a well located between MW2 and the outcrop.

Geologic structures that could hinder groundwater flow out of the basin fill, creating a western no-flow boundary for the fractured-bedrock aquifer, are a Basin and Range normal fault on the western side of the Lake Mountains and/or an older transverse tear fault in the Paleozoic bedrock, which are both generally located where the basin-fill aquifer intersects the fractured-bedrock aquifer (figure 15). Although the exact position of the normal fault that bounds the western side of the Lake Mountains is unknown (Biek, 2004), the relatively shallow depth to Paleozoic bedrock in MW2b (414 feet [126 m]) suggests the fault is not located east of the borehole, but may project west of the borehole (figure 15). Fault gouge or discontinuity of geologic units can impede groundwater flow across a fault plane (Lachmar and others, 2002; Bense and Person, 2006).

Aquifer Test Setup

Well 3 had not been pumped for 76 days prior to beginning Well 3 aquifer test, and there are no other large production wells for several miles around. A few private supply wells in the vicinity of 807 and one south of MW2a&b may have been pumping; however, because most of the private wells are small diameter (≤ 8 inches [20 cm]) and equipped with small domestic pumps (maximum flow rate ≤ 25 gallons per

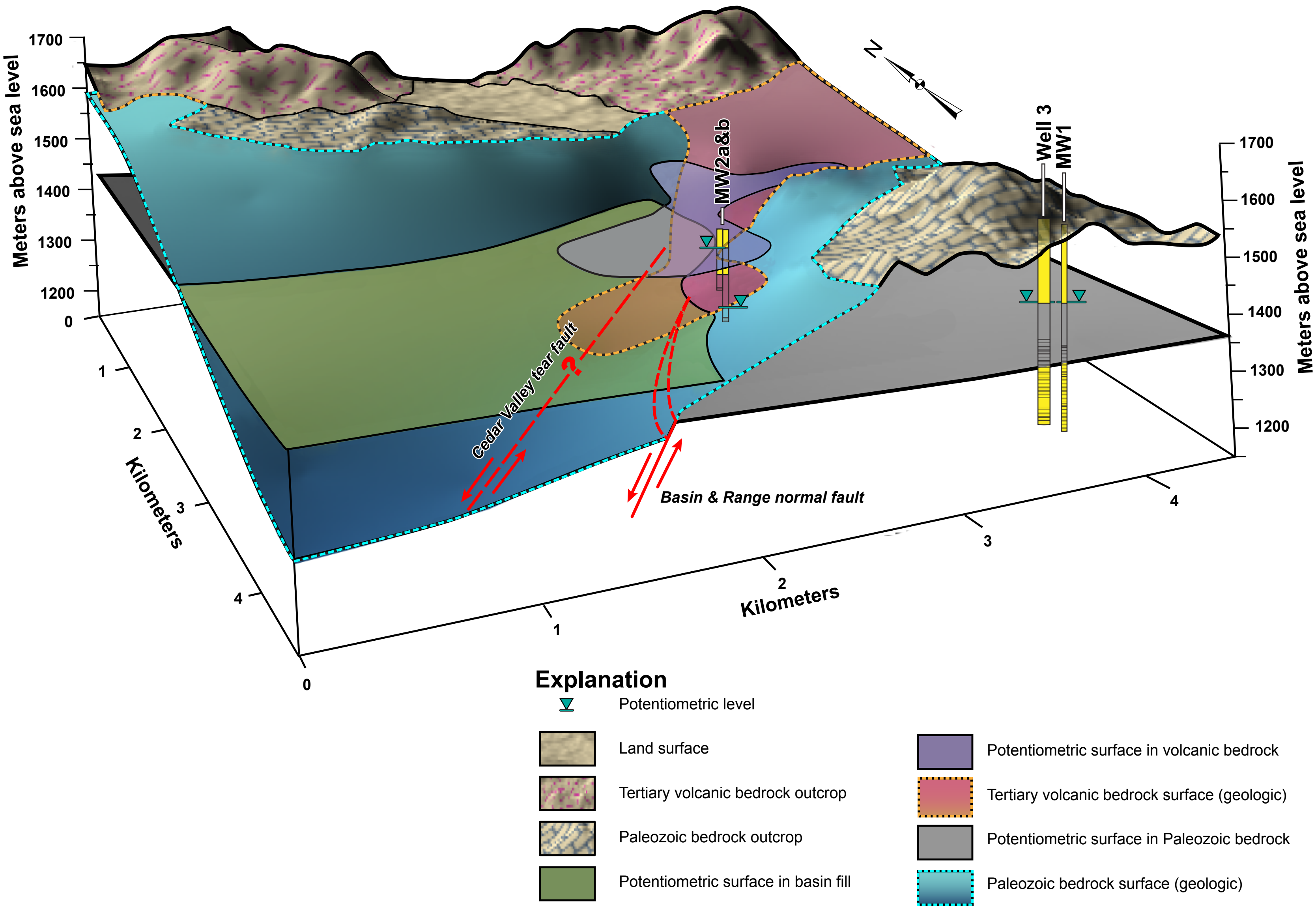


Figure 17. *Oblique view of the hydrostratigraphy on the western side of Cedar Pass. Area of detail is shown on figure 15. The surface of the basin fill has been removed to show the position of a wedge of Tertiary volcanic rocks (in pink) between the basin fill and the Paleozoic bedrock. The potentiometric surface of the Tertiary volcanic unit (in transparent purple) is high above the top of the volcanics. The water table in the basin fill (in transparent green) laps up onto the top of the volcanic unit and may intersect the surface of the Paleozoic bedrock at about the location of a buried normal fault. Both the volcanic unit and the fault may act as barriers to groundwater flow.*

minute [1.6 L/s]), I assumed that any effects from the private wells on wells 807 or MW2b were masked by the relatively greater magnitude of the drawdown resulting from Well 3 pumping.

Eagle Mountain water managers started pumping Well 3 at 1:50 PM on May 1, 2007. Flow and water level in the pumping well were measured by the city's in-line permanent magnetic flow meter and pressure transducer, respectively, and are tabulated in appendix A, tables A-4 and A-5. Eagle Mountain personnel estimated that the discharge in the first seven minutes of the test, when the water is routed “to waste” and is not metered, was approximately 2400 gallons per minute (151 L/s). Flow decreased throughout the five months of pumping from an average of 1960 gallons per minute to 1910 gallons per minute (124 L/s to 121 L/s) as depth to water increased. The well pump was off for a few hours several times during the test, but these periods occurred well into the test (day 49 and later), and effects on the interpretation of aquifer properties were inconsequential. The software program I used to analyze the aquifer-test data (Duffield, 2007) takes changes in discharge into account. Discharge was routed to the city water lines for use mostly outside the cone of depression. A few tens of acre-feet were used for irrigation at the surface within the cone of depression, but because the water table is over 500 feet (152 m) deep and the volume of water used for irrigation was small compared to total discharge, infiltration to the water table and an effect on drawdown were unlikely during the test.

Water-Level Observations

Data Collection Methods

The city’s data logger recorded water levels and discharge of Well 3 automatically every 1 hour 20 minutes for several months prior to the test. These background data showed that the well pumped intermittently depending on water demand between July 21, 2006, and October 15, 2006, and again between January 9, 2007, and February 13, 2007. Eagle Mountain personnel shut off Well 3 on February 13, 2007, to allow water levels to recover for the aquifer test. Between February 14 and the start of the aquifer test on May 1, water level in Well 3 rose approximately 2 feet (0.6 m).

Background water-level data collection on observation wells MW1 and MW2b was restricted to the two to three months between when the wells were completed (February 2007) and the start of the aquifer test on May 1, 2007. Water levels in both wells trended upward by approximately 1.2 feet (0.37 m) in the two months prior to the test (figure 18). My attempt to correct drawdown data for this antecedent trend is described below. Water levels collected for this test are tabulated in appendix A, table A-4.

The city’s pressure transducer measured water levels in Well 3 throughout the test until the transducer malfunctioned three days into the recovery period. City officials were unsure of the type of transducer, absolute or vented, in the well. Water

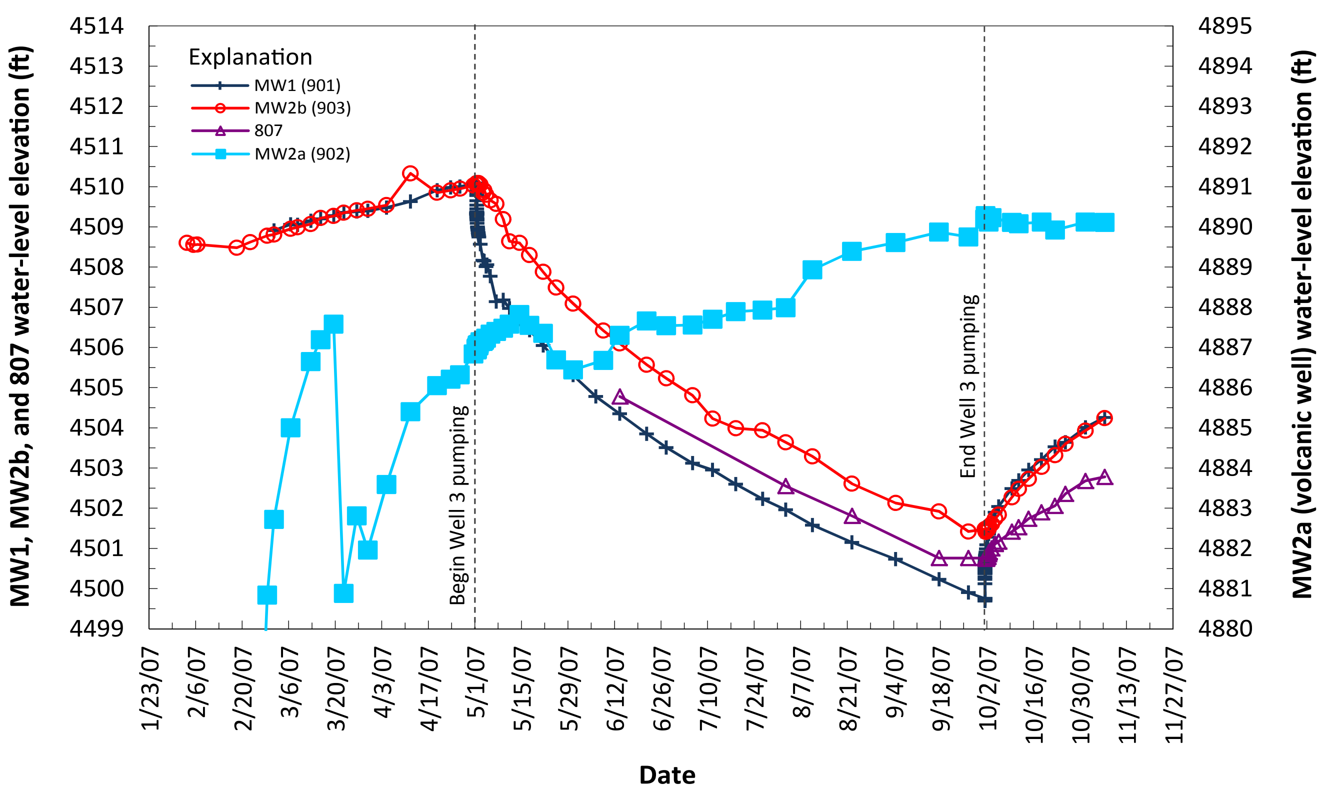


Figure 18. *Water-level response in observation wells during Well 3 aquifer test.*

levels in the Eagle Mountain observation wells were measured manually using electronic water-level sounders. MW1 was also equipped with a pressure transducer, although the later portion of the drawdown test was not recorded on the transducer.

I found observation well 807 43 days into the test and immediately began measuring water levels using an electronic water-level sounder (figure 18). In order to calculate drawdown, a pre-test static water level is required. I calculated the pre-test static water-level elevation in 807 as 4506.70 feet (1373.64 m) above mean sea level (AMSL) by extrapolating a straight line through the late-time drawdown data back to the start of the test. Because drawdown is logarithmic in nature (quicker rate in early time), this method most likely underestimates the amount of drawdown in 807. Another method to estimate static water level is to mirror the early recovery water-level curve onto the drawdown. Static water-level elevation by this method (4507.10 feet AMSL [1373.76 m]) is higher by 0.4 foot (0.1 m) than the water level I calculated by the straight-line method. Recovery is not always the reverse of drawdown, so to evaluate the applicability of using the reverse of the recovery curve, I analyzed the shape of the drawdown and recovery curves in the other distal observation well, MW2b. The drawdown curve was steeper (quicker rate of change) than the recovery curve in MW2b. If 807 had a similar difference in the shapes of the drawdown and recovery curves, the static water-level elevation would be as much as 0.8 foot (0.2 m) higher (4507.50 feet [1373.89 m]) than the water level calculated by the straight-line method. Given the uncertainty in estimating the static water level of 807, I estimate the error on the magnitude of drawdown due to the missing pre-test static water-level measurement in 807 is +1.0 to -0.2 feet (+0.3 to -0.06 m). The effect of erroneous drawdown on the results of aquifer-test analysis is discussed with those results.

Water levels in all wells were measured for 36 days during the recovery portion of the aquifer test.

Water-Level Corrections

Barometric pressure correction: Water-level data collected for this test needed two types of adjustments to correct for changes in barometric pressure. The first correction is mandatory when using absolute pressure transducers to measure water levels. Absolute pressure transducers, like the transducer used in MW1, measure the total pressure on their sensors; this includes the pressure exerted by the water column in the well and the atmospheric pressure on the surface of that water column. The change in barometric pressure relative to a constant must be measured at or as near as possible to the well and subtracted from the absolute pressure transducer readings to determine the water level in the well. Water levels collected using vented transducers or manually with tape measuring devices do not need barometric pressure change subtracted. The second type of correction is to remove the real change in a well's water level caused by the differential response of the well water to the change in atmospheric pressure versus the response from the aquifer matrix and pore water. This second type of correction is more complicated and may be necessary on water levels collected with any type of equipment.

The water level in a well may display an inverse relationship to barometric pressure; that is, an increase in barometric pressure may cause the observed water level in a well to decline, and a decrease in pressure may cause the water level to rise (Freeze and Cherry, 1979, p. 233). The mechanisms that determine the magnitude of water-level change with respect to barometric pressure change are described in detail by Hare and Morse (1997), Rasmussen and others (1997), Spane (2002), Halford (2006), Toll and Rasmussen (2007), and Butler and others (2011) and depend on the type of aquifer, nature of the unsaturated zone or confining unit, depth to water, and the size and condition of the wellbore. Typically, confined aquifers respond quickly to barometric pressure change, whereas the response in unconfined aquifers is delayed because barometric pressure takes time to transmit through the unsaturated zone (Rasmussen and Crawford, 1997). If a change in barometric pressure (expressed in feet of water units) produces the same change in water level (also measured in feet), the well is said to have a barometric efficiency (BE) of 1 (Jacob, 1940; Freeze and Cherry, 1979, p. 233). Similarly, a well's barometric response function (BRF) characterizes a well's water-level response over time to a change in barometric pressure (Butler and others, 2011). For aquifer-test analysis, especially when the magnitude of drawdown is of the same order of magnitude as the potential water-level change due to barometric pressure fluctuations, it is important to consider barometric pressure effects on water level. In the Well 3 aquifer test, the rate of drawdown in the observation wells was such that the daily variation in barometric pressure of approximately 0.35 inch of mercury (0.4 foot of water) was larger than the daily drawdown, and attention to the wells' responses to barometric pressure fluctuations was necessary. Large high and low pressure weather systems during the test produced barometric pressure changes of as much as 0.8 foot (0.24 m) of water.

I attempted to simultaneously remove the measured barometric pressure and identify and remove real water-level changes induced by barometric pressure fluctuation from MW1 unvented (absolute) pressure transducer data using a method described by Rasmussen and Crawford (1997) and implemented in BETCO, a computer program by Toll and Rasmussen (2007). The program subtracts the barometric pressure, which was measured at the Lehi, Utah, weather station located approximately 9 miles (15 km) northeast of MW1, from transducer readings, and then allows the user to systematically vary the BE and lag time (to produce a BRF for the well) until the data are smooth. I was unable to process the data into a smooth curve, possibly because of unidentified water-level trends in addition to the barometric pressure-induced fluctuations. Instead, I simply corrected MW1 absolute transducer data for changes in barometric pressure by subtracting the change in barometric pressure. The variability in water levels observed in MW1 after barometric pressure was subtracted, is approximately 0.2 foot (0.06 m) based on a visual comparison of the corrected water levels to a smooth drawdown curve. If these variations are due to barometric pressure changes, then the BE or BRF is approximately 0.5.

I did not correct water levels that were collected with electronic water-level sounders for BE or BRF because the observations were not temporally dense enough to correlate with barometric pressure data. Assuming similar barometric response in observation wells 807 and MW2b as in MW1, I estimate that the error on each water-level measurement may be as much as 0.2 foot (0.06 m). While errors from the "true" water level of 0.2 foot (0.06 m) will affect the degree of fit of the analytical type curve match, the final position of the matched curve to the average late-time water level trend, and thus the aquifer parameters, will not be significantly different from a curve match with properly corrected water levels.

I did not apply barometric pressure corrections (either subtraction of barometric pressure change from absolute pressure transducer readings or correction for BE or BRF) to Well 3 data because (1) I was not certain of the type of transducer used, (2) well and pump hydraulics caused the water level in the well to fluctuate on an hourly basis by more than the amount of possible barometric pressure correction, and (3) the magnitude of drawdown was so much larger (41 feet [12.5 m]) than the amount of possible barometric pressure correction that correction would have a negligible effect on aquifer-test analysis.

Antecedent trend: Aquifer-test drawdown data should be corrected for any antecedent water-level trend. The upward trends observed in wells MW1 and MW2b during the three months prior to the aquifer test were likely caused by recovery from well development (by air lifting water from the wells after well completion) and recovery from pumping of Well 3, which was discontinued almost three months prior to the aquifer test. In addition to the recovery trend, additional trends, such as barometric well function, regional groundwater recharge, or seasonal trends, may have affected the water level in the wells (Hare and Morse, 1997; Rasmussen and Crawford, 1997; Spane, 2002; Halford, 2006). In order to identify these types of trends, long-term water-level data collected in periods devoid of pumping are required. Since the first wells drilled into this aquifer in this area were production wells, these historical data do not exist.

A simple and objective method to correct for an antecedent trend, such as removing a linear or logarithmic background trend as described by Spane (2002), was not applicable to Well 3, or observation wells MW1 or MW2b data because the observed antecedent water-level trends are at least partially due to recovery. Using a linear or logarithmic trend correction would have over-corrected late-time data because the change in water level due to recovery becomes less in late time as compared to the extension of the linear trend (A and A′ on figure 19). I attempted to superimpose and match the trend observed in the wells during the October 2007 recovery portion of the aquifer test onto the background data to extrapolate the recovery curve into the test and calculate a correction, but the differences in volume of water removed and duration of pumping created a dissimilar recovery response (B and B′ on figure 19).

I attempted to correct MW1 water-level data for the antecedent trend plus barometric effects (barometric well function), earth tides, and regional water-level trends before and during the aquifer test using an advanced spreadsheet program created by Halford (2006). This method requires that the conditions affecting water levels during the background data collection period exist during an aquifer test, and that the background data collection period is long enough to observe regional and seasonal trends. In the four examples given by Halford (2006), the error in corrected water levels during an aquifer test was acceptably small only when the background data collection period was at least as long as the aquifer test and best when the background period was four times as long as the test and collected immediately antecedent to the test. Background data should also be collected when there is no residual recovery or drawdown due to pumping. MW1 was completed only two months prior to the start of the test, and Eagle Mountain could not spare the use of Well 3 for long enough to collect background data for a five-month aquifer test. Therefore, the time available for background water-level collection to use in the spreadsheet program was too short to identify seasonal trends or the length of the recovery from the January–February 2007 pumping for this five-month aquifer test.

I quantified the potential error in aquifer parameter estimation due to not correcting aquifer-test water levels for the apparent recovery that occurred before the test. I assumed the entire observed upward trend was due to recovery, not a seasonal recharge trend, and visually extrapolated a hypothetical recharge curve (C on figure 19). For wells MW1 and MW2b the additional drawdown at late time was approximately 0.9 foot and 1.1 feet (0.27–0.34 m), respectively. I analyzed the hypothetical corrected drawdown curves and noted no significant difference in any of the aquifer parameters except for MW2b hydraulic conductivity, which was approximately 10% lower than the value returned from analysis of the non-corrected data.

Partial penetration: The pumping well is screened from 140 feet to 500 feet (43–152 m) below the top of the 530-foot-thick (162 m) aquifer, or over approximately 68%

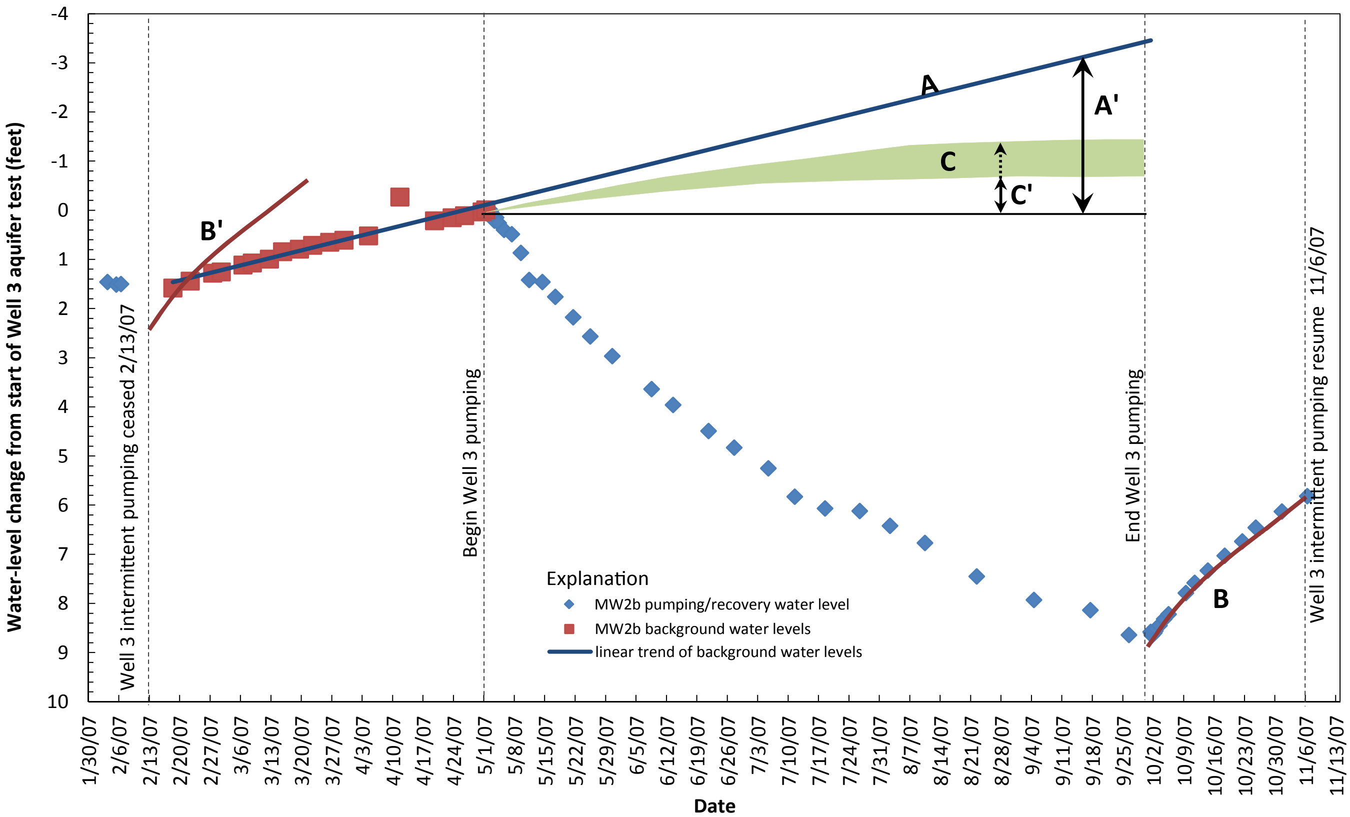


Figure 19. *Possible antecedent water-level trend correction methods applied to well MW2b. Linear extrapolation of antecedent trend (A) leads to overcorrection of late-time data (A'). Superimposing the recovery curve (B) onto background data (B') is not a good fit. Continuation of antecedent trend as a hypothetical recovery curve (C) leads to a correction to late-time data of an additional 0.75 to 1.5 feet (C').*

of the aquifer thickness. Drawdown is more in a partially penetrating pumping well than if that well was fully penetrating, and must be corrected in order to calculate the true aquifer transmissivity (Hantush, 1961, 1964). Nearby observation wells screened in the same part of the aquifer (i.e., MW1) will also experience more drawdown and should be corrected (Kasenow, 2006). I chose to use a vertical hydraulic conductivity to horizontal conductivity ratio (symbolized as K_v/K_h, K_z/K_h, K_z/K_r, or β, depending on the analytical solution) between 0.25 and 1 for reasons explained below in the discussion on curve matching. The relatively high K_v/K_h I used reduces the effect of partial penetration on wells as compared to an aquifer with a very low vertical to horizontal conductivity ratio (Fetter, 1988, pg. 189; Kruseman and de Ridder, 2000, pg. 159). However, drawdown correction still is necessary in Well 3 and MW1, and was done by the curve matching software I used (Duffield, 2007). Observation wells 807 and MW2b are far enough from the pumping well for the effects of partial penetration to be negligible (Hantush, 1964; Kruseman and de Ridder, 2000, pg. 159).

Water-Level Response to Pumping

Water-level elevations in the bedrock aquifer before the test were within 1 foot of each other over the approximately 1-square mile monitored by wells MW2b, MW1, and Well 3, making the potentiometric surface remarkably flat. After five months of pumping, the maximum recorded drawdown was 41 feet (12.5 m) in the pumping well and between approximately 6 and 11 feet (1.8–3.4 m) in the observation wells (table 3). The drawdown cone is relatively flat and wide as shown by only slightly more drawdown (10.4 feet [3.17 m]) in MW1, located 537 feet (164 m) from the pumping well, as compared to the 8.65 feet (2.64 m) of drawdown in MW2b, located 4939 feet (1505 m) away. A wide, flat drawdown cone indicates the aquifer may have a relatively high transmissivity (Ferre and Thomasson, 2010). Anisotropy is shown by less drawdown in 807 (5.98 feet [1.81 m]) than in MW2b (8.65 feet [2.64 m]), even though 807 is located 399 feet (122 m) closer to the pumping well than MW2b. Relatively greater drawdown in MW2b indicates that MW2b is located closer to the high conductivity fracture zone intersecting the well than 807, as drawdown is always greatest in the observation well closest to a pumped fracture, regardless of radial distance from the well (Jenkins and Prentice, 1982). The water level in MW2a did not respond to pumping in the bedrock aquifer because MW2a is completed in a volcanic rock unit overlying and hydraulically separate from the sedimentary bedrock aquifer.

The drawdown responses in the observation and production wells are shown on a log-log plot on figure 20. The roughly 20% greater discharge rate from the pumping well in the first seven minutes of the aquifer test generated 40.98 feet (12.49 m) of drawdown in Well 3 and 0.78 foot (0.24 m) in MW1 in those first minutes, after which the discharge rate eased and water levels actually recovered (7.3 feet [2.23 m] in Well 3 and 0.03 foot [0.01 m] in MW1) in response to the lower flow rate. Drawdown in all affected wells continued to increase over the following five-month pumping period. In August and September, drawdown at the affected wells was increasing at a nearly steady rate of 1 foot (0.3 m) every 24 to 28 days.

Analysis

My analysis of the Well 3 aquifer-test data involved identification of wellbore storage, determination of the type of flow in the aquifer, curve matching to analytical solutions, evaluation of aquifer boundaries and their effects on drawdown, and calculation of potential anisotropy in the transmissivity of the aquifer.

Wellbore Storage

Drawdown data from large pumping wells and wells in lower transmissivity aquifers are more likely to be affected by wellbore storage (Papadopulos and Cooper, 1967; Moench, 1984; van Tonder and others, 2002, part B, pg. 16). Well 3 is a 16-inch-diameter (41 cm) well in a 22-inch-diameter (56 cm) borehole, which provides adequate volume for wellbore storage, given the moderate transmissivity of the aquifer. A slope of one on the early-time data drawdown and derivative curves on a log-log plot of drawdown versus time is an indication of wellbore storage (van Tonder and others, 2002, part B, pg. 15; Renard and others, 2009). Early-time data from Well 3 and MW1 plotted on log-log scale have unit slopes, indicating wellbore storage (figure 20). I calculated that the first 15 to 30 minutes of the Well 3 aquifer test should have been affected by wellbore storage using a simple equation given by Papadopulos and Cooper (1967); however, a decrease in discharge at approximately seven minutes into the test affects the shape of the drawdown curve and obscures the change in slope between the unit slope characteristic of wellbore storage and the true aquifer response.

The effect of wellbore storage on drawdown decreases with increasing distance from the pumped well, and the early-time observation well water-level data have a slope increasingly greater than one the farther away the observation well is from the pumping well (van Tonder and others, 2002, part B, pg. 15). Wells MW2b and 807 are located too far from the pumping well to see the effect of wellbore storage.

Curve Matching

I used AQTESOLV (Duffield, 2007) to match theoretical curves generated mathematically from published analytical solutions to the aquifer-test data. My choice of input parameters and definition of aquifer flow characteristics necessary for curve matching are defined below.

Aquifer physical properties: I defined the following parameters describing the physical nature of the aquifer for use in the analytical solutions: aquifer thickness, vertical to horizontal hydraulic conductivity ratio, and the shape of the aquifer matrix blocks separated by fractures. I chose an aquifer thickness of 530 feet (162 m) as defined in Geologic Setting, above. The ratio of K_v to K_h may vary from 10^2 in fractured rocks, in which the permeability is mainly due to vertical fractures, to 10^{-3} in stratified sedimentary rocks, where horizontal bedding planes form the easiest passage for water (Singhal and Gupta, 2010).

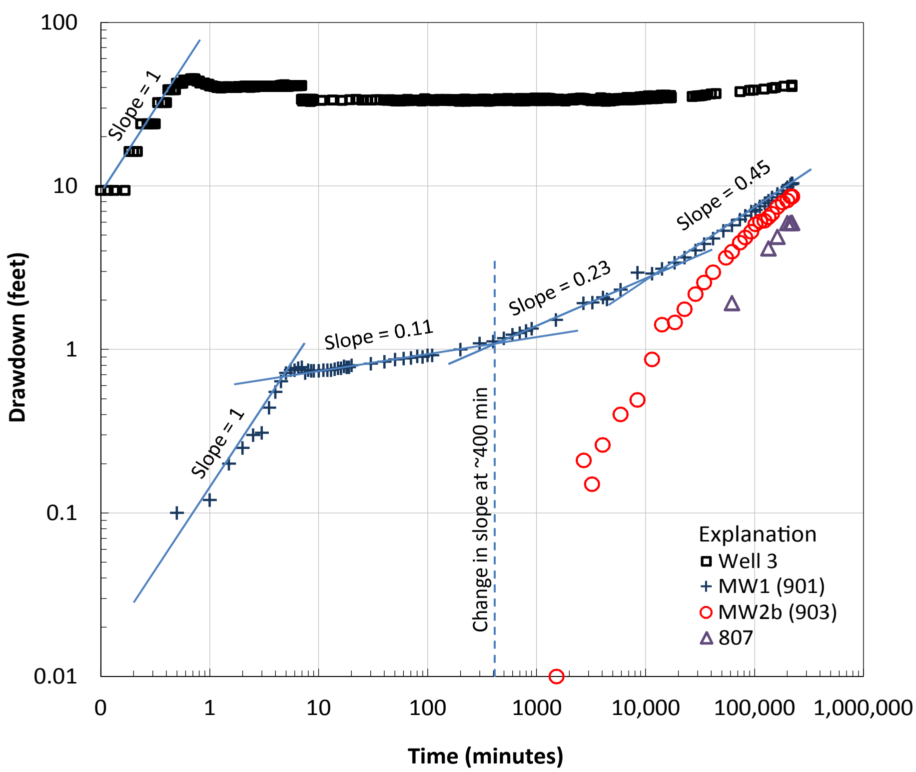


Figure 20. Log-log plot of drawdown in the production and observation wells during the pumping portion of the five-month constant rate aquifer test on Well 3. Early drawdown data are affected by the high pumping rate in the first seven minutes of the aquifer test. The effect of wellbore storage in Well 3 and MW1 is indicated by a unit slope in early-time data. A change in slope at 400 minutes, a doubling of the slope in late time, and a slope of nearly 0.5 in late time may be indicative of aquifer boundaries.

I chose a vertical to horizontal hydraulic conductivity ratio of between 0.25 and 1 because the prevalence of fractures encountered during drilling in this consolidated rock (figure B-6; Jordan, 2008) would tend to reduce the normally much higher tendency of water to flow along the limestone bedding planes. I determined a block diameter of 3 feet by examining outcrop near MW1, which had joints roughly perpendicular to bedding planes every 2 to 3 feet (0.6–0.9 m).

Linear flow: Numerous researchers (e.g., Gringarten and Witherspoon, 1972; Gringarten and others, 1974; Jenkins and Prentice, 1982; Smith and Vaughan, 1985; Streltsova, 1988; Gernand and Heidtman, 1997; Allen and Michel, 1998) have detected linear flow in fractured-rock aquifers. Linear flow occurs when flow is laminar and linear toward the plane of a highly transmissive fracture intersecting a well. I suspected linear flow may occur along a set of interconnected fractures on the axis of the anticline in the early part of the Well 3 aquifer test. If linear flow is present, the data should fall on a straight line on a plot of drawdown versus the square root of time with arithmetic axes (Jenkins and Prentice, 1982), and the slope of early-time drawdown versus time data on a log-log plot should fall on a straight line having a slope between 0.25 and 0.5 (Streltsova, 1988; Gernand and Heidtman, 1997). Data for Well 3, MW1, and MW2b passed the first test for most of the duration of the aquifer test (figure 21), but because early-time data were affected by wellbore storage as shown in figure 20, the early-time data have slopes of 1, and therefore the second test for 0.25 to 0.5 slope is not applicable. I interpret these results to show some characteristics of linear flow in the aquifer.

Choice of solutions: I used the analytical solution described by Moench (1984) for a double-porosity fractured-rock aquifer to match time versus drawdown data from MW1 and Well 3, and radial flow solutions (Theis, 1935; Neuman, 1972, 1974) for wells MW2b and 807. The use of different solutions is necessary because of the distance to each observation well—the difference in scale affects the drawdown response at each well. Double (or dual) porosity is characterized by linear flow to the pumping well from the fracture network at the beginning of pumping followed by water released from storage in the matrix blocks at later time. Linear flow from the fractures can be detected only at early time (generally the first few to tens of minutes, depending on aquifer and pumping characteristics) and near the pumping well (Jenkins and Prentice, 1982; Moench, 1984; Kruseman and de Ridder, 2000; Singhal and Gupta, 2010). Bourdet and Gringarten (1980) in their method for analyzing drawdown response to pumping in a double-porosity fractured aquifer, provide an interporosity flow coefficient (λ) that is dependent on the number of fracture sets, typical distance between the fracture sets, distance to the observation well, and the ratio of the hydraulic conductivity in the fractures to that of the matrix. For wells with λ less than 1.78, the effect of double porosity should be observable in the drawdown response (Bourdet and Gringarten, 1980; Kruseman and de Ridder, 2000). For MW1, I calculated λ to be approximately 1, so the response from the double porosity of the fractures and the matrix should be detectable in MW1 water-level observations. In most reported cases, observation wells distant from the fracture will have a water-level response typical of radial flow to the pumping well and can be interpreted by radial flow methods (Jenkins and Prentice, 1982; Moench, 1984; Smith and Vaughan, 1985; Gernand and Heidtman, 1997). Wells MW2b and 807 are far enough from the pumping well that the initial response of fracture flow is not detectible and radial flow solutions can be used.

Solutions using aquifer boundaries: The effect on drawdown of one or more aquifer boundaries can be incorporated into aquifer-test analysis. A lateral boundary that inhibits groundwater flow, called a no-flow boundary, will increase the drawdown in wells between the pumping well and the boundary when the cone of depression around a pumping well extends radially outward far enough to intersect the no-flow boundary. As discussed in the Geologic Setting section above, the wedge of Tertiary volcanic rocks, the Basin and Range normal fault, and possibly an older transverse fault west and northwest of the well (figure 17) likely inhibit flow from the

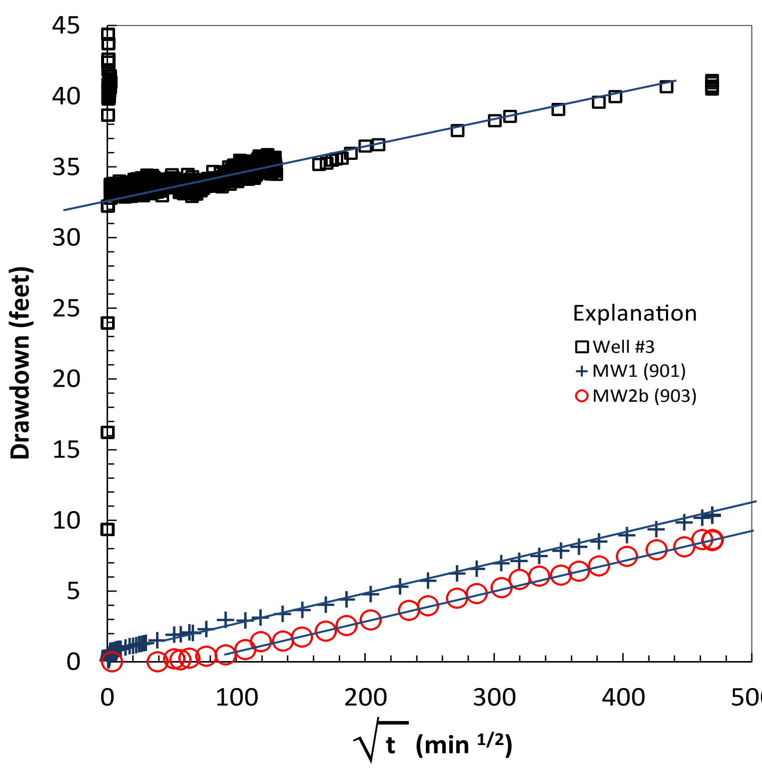


Figure 21. *Linear flow in the aquifer may be indicated by a straight line fit to the plot of drawdown versus the square root of time on arithmetic scale (Jenkins and Prentice, 1982).*

basin fill to the Paleozoic bedrock to an unknown degree, and may inhibit flow through the bedrock underlying the basin fill. To incorporate this possibility in my analysis, I placed a no-flow boundary roughly parallel to the normal fault bounding the west side of the Lake Mountains (figure 15) and ran automatic curve matching for each well with and without the boundary in place. In reality, these geologic features appear to be acting as a semi-permeable boundary, so running the two scenarios is a way to bracket the range of aquifer parameters.

Double-porosity matches—MW1 and Well 3: Double porosity is characterized at early time by flow to the pumping well from water stored in the fracture network, followed at late time by water released from storage in the matrix blocks (Moench, 1984; Kruseman and de Ridder, 2000). The drawdown curve of the response of a double-porosity aquifer plotted on log-log scale may have an early-time drawdown curve representative of flow to the well from the higher transmissivity fracture network, followed by a nearly-horizontal plateau formed by the transition to matrix flow, and finally, a late-time drawdown curve representative of flow from the matrix to the fissures and through the fissures to the well (Moench, 1984; Gernand and Heidtman, 1997; Kruseman and de Ridder, 2000; Singhal and Gupta, 2010). The drawdown curve of MW1 shows this general pattern except that wellbore storage dominates the first 15 to 30 minutes of the test and straightens the early-time drawdown curve produced by the initial fracture flow (figure 20). Once corrected for wellbore storage, the drawdown in the early-time data is indicative of the fracture flow response. The next segment of the drawdown curve is semi-horizontal (slope = 0.1) on a log-log plot and this likely represents the transition from fracture to matrix flow, during which time the well is receiving water from both systems. The increase in the rate of drawdown after a few hundred minutes may represent transition to water released from storage in the matrix blocks, but the rate increase may also be compounded by the existence of an aquifer boundary, as discussed later in this report.

Figure 22 is a compilation of the curve matches produced from curve matching using AQTESOLV curve-matching software (Duffield, 2007). Curve matching for MW1 with no aquifer boundary produced a good match of the double-porosity fractured-rock solution of Moench (1984) to the observed data. I compared the ideal type curve and derivative curve solutions for different types of aquifers presented by Renard and others (2009) to the observed drawdown and derivative plots for MW1. The shape of the curves indicates that (1) wellbore storage is affecting the well response in the early part of the test, and (2) flow during late time is radial to the well but has a non-integer flow dimension less than two. A non-integer flow dimension less than two can result from combination of one-dimensional flow as is the case to a linear fracture and two-dimensional flow as is typical for radial flow to a well in a uniform layer (Cook, 2003, pg. 47). In the fractured carbonate aquifer at Cedar Pass, this combination is possible as linear flow to interconnected vertical fractures on the anticlinal axial plane intersecting the well and typical radial flow through the carbonate rock to the well. The values of aquifer parameters derived from the curve matches represented on figure 22 are given in table 4 and were mostly within reasonable values for a fractured-rock aquifer, and are discussed later in this report concurrently with the aquifer parameters derived from curve matching data from wells MW2b and 807.

Data from pumping wells are often difficult to analyze because of well loss (loss of head due to well construction), wellbore storage, and the difficulty of obtaining manual water-level measurements to check the accuracy of pressure transducers installed by pump contractors. Variability in discharge also has an immediate and sometime large effect on the drawdown in the pumping well. In the Well 3 test, we observed much higher discharge in the first seven minutes of the test because there was less backpressure on the pump while the water pumped to waste (appendix A, table A-5). Nevertheless, a double-porosity solution correcting for wellbore storage and variable flow rate (Dougherty and Babu, 1984; Moench, 1984, 1988) provided a reasonably good match to the time-drawdown data (figure 22).

I ran MW1 and Well 3 automatic curve matching analyses a second time with a no-flow boundary located as shown on figure 15. AQTESOLV uses image well theory (Ferris and others, 1962) to simulate the effect of a boundary on drawdown. The aquifer parameters resulting from matching Well 3 data showed hydraulic conductivity to be larger and storage to be smaller when a boundary was present, whereas the parameters derived from matching MW1 data were very similar in the boundary and no-boundary matches (table 4).

Radial flow matches—MW2b and 807: Two pieces of evidence suggest the aquifer at MW2b may be unconfined: (1) the Oquirrh Group bedrock above the water table at this well location was unsaturated during drilling (Jordan, 2008), and (2) water level in MW2b, located 4939 feet (1505 m) northwest of the pumping well, began to decline after more than one day of pumping. Distal wells in unconfined aquifers typically take longer to respond to pumping than those in confined aquifers because the storativity of unconfined aquifers is usually much larger than confined aquifers (Singhal and Gupta, 2010, pg. 166). However, the aquifer may show response more typical of a confined aquifer because the presence of weathered clay and tight volcanic rock above the bedrock limit the connection between the deep (~500 feet) water table and the atmosphere. I performed curve matching using confined (Theis, 1935) and unconfined (Neuman, 1972; Theis [1935] with the Cooper-Jacob [1946] correction to drawdown) solutions. The resulting curve matches and aquifer parameters were similar with the exception of variation in the storativity of approximately one and one-half orders of magnitude (table 4). A representative curve match is shown on figure 22.

AQTESOLV allows the user to control the variation of some parameters and set others to constant values when using these radial flow solutions. I chose to allow the ratio of vertical hydraulic conductivity to horizontal hydraulic conductivity (K_v/K_h) to vary between 0.05 and 1 in order to test my assumption that fractures allow more permeability in the vertical direction than is typical for non-fractured sedimentary rock. (The K_v/K_h used in the double-porosity solutions was set at 0.25. I also matched the data with the K_v/K_h set to 1 and the results were very similar, indicating the double-porosity curve matching is not sensitive to K_v/K_h in this case.)

Including a boundary located approximately ¼ mile west of MW2b in my aquifer-test analysis had a tremendous effect on the numerical results of curve matching MW2b data. While the type and derivative curves fit almost as well in the aquifer-boundary scenario as in the no-boundary scenario (typical match shown on figure 22), the calculated transmissivity values were more than double because of the proximity of a boundary that limits recharge to the cone of depression (table 4).

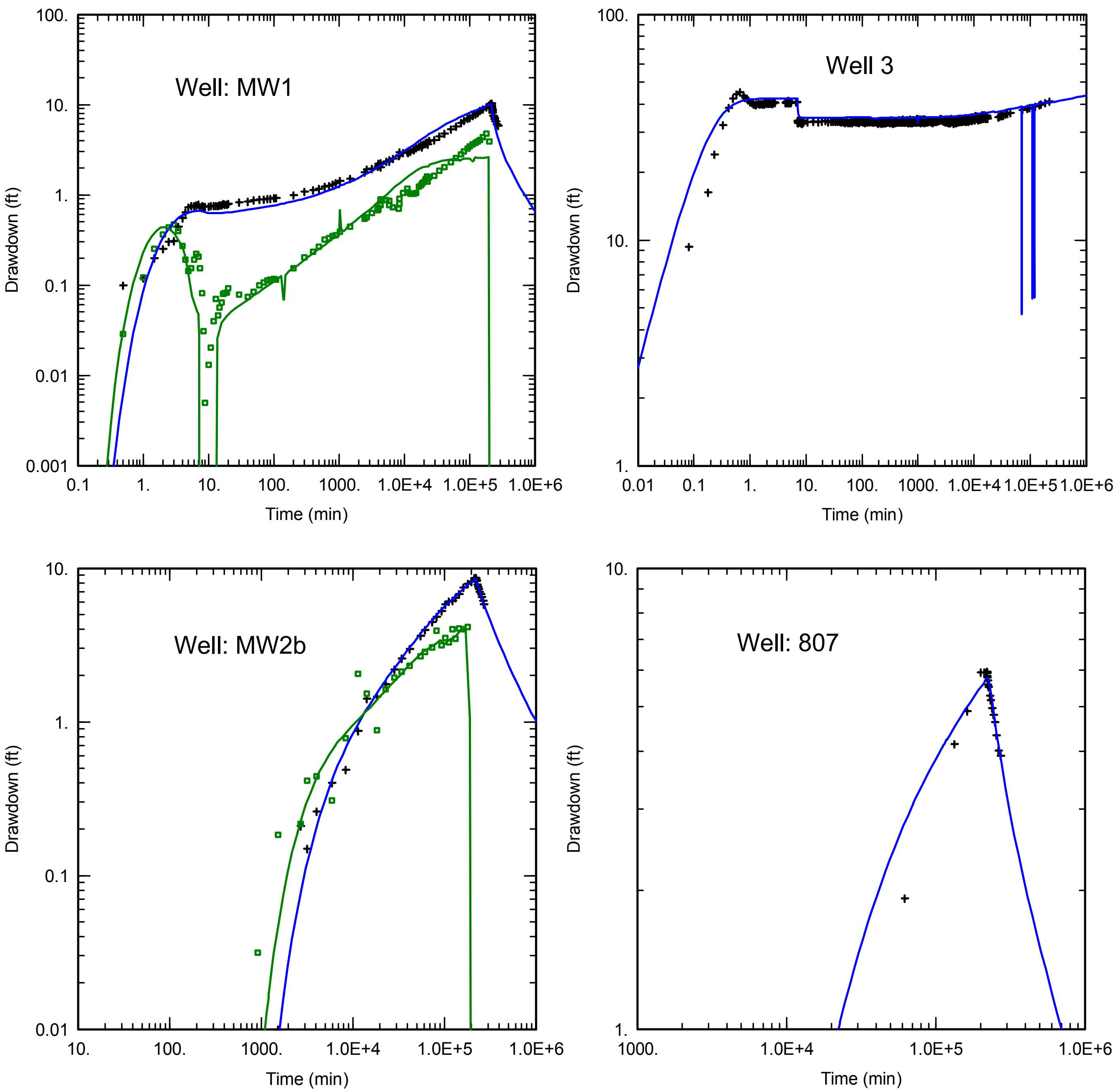


Figure 22. *Typical curve matches of aquifer-test data from pumping and observation wells involved in the Eagle Mountain Well 3 five-month constant-rate aquifer test. Analytical solution type curves (solid blue lines) are matched to water-level change (black symbols), and derivative curves (green curves) are matched to the derivative of the drawdown (green squares). Curve matches obtained using the various analytical solutions described in the text (confined, unconfined, double porosity, bounded, and unbounded scenarios) were visually similar to the typical matches shown here. Aquifer parameters derived from curve matching are listed in table 4.*

Table 4. *Aquifer parameters estimated from type curve matching at individual wells involved in the Well 3 aquifer test.*

Well	Solution (note 1)	Boundary condition	K (ft/d)	K' (ft/d)	K_v/K_h	T (ft^2/d)	S_s (ft^{-1})	S_s' (ft^{-1})	S_y	S	S_w	S_F	$r_{(w)}$ (ft)
Well 3	double porosity	none	25	2×10^{-5}	0.25 (note 2)	1.3×10^4	6×10^{-9}	3×10^{-4} (note 3)	n/a	1.4×10^{-1} (note 3)	0 (note 3)	0.9 (note 3)	3 (note 3)
Well 3	double porosity	1	39	8×10^{-7}	0.25 (note 2)	2.0×10^4	2×10^{-10}	1×10^{-5} (note 3)	n/a	7.8×10^{-3} (note 3)	0.3 (note 3)	0.6 (note 3)	1 (note 3)
MW1	double porosity	none	21	5×10^{-5}	0.25 (note 2)	1.1×10^4	2×10^{-7}	6×10^{-4} (note 4)	n/a	3×10^{-1} (note 4)	20	0.2	0.8
MW1	double porosity	1	22	4×10^{-5}	0.25 (note 2)	1.2×10^4	4×10^{-7}	6×10^{-4} (note 4)	n/a	3×10^{-1} (note 4)	20	0.2	0.8
MW2b	confined and unconfined	none	14	n/a	0.1 to 1	7.5×10^3	n/a	7×10^{-6} to 2×10^{-5}	4×10^{-3}	4×10^{-3} to 1×10^{-2}	n/a	n/a	n/a
MW2b	confined and unconfined	1	28	n/a	0.05 to 0.1	1.5×10^4	n/a	2×10^{-5}	5×10^{-3}	1×10^{-2}	n/a	n/a	n/a
807	confined and unconfined	none	19	n/a	0.01 to 0.7	9.9×10^3	n/a	2×10^{-5} to 5×10^{-5}	3×10^{-2}	9×10^{-3} to 3×10^{-2}	n/a	n/a	n/a
807	confined and unconfined	1	28	n/a	0.25 to 1	1.5×10^4	n/a	2×10^{-5} to 5×10^{-5}	2×10^{-2}	9×10^{-3} to 2×10^{-2}	n/a	n/a	n/a

Abbreviations: K = horizontal hydraulic conductivity, K' = horizontal hydraulic conductivity of the rock matrix in double porosity solutions, K_v/K_h = vertical to horizontal hydraulic conductivty ratio (also denoted in formulas as K_z/K_r, K_z/K_h, or β), T = transmissivity, S_s = specific storage or specific storage of the fractures in double porosity solutions, S_s' = specific storage of the matrix in double porosity solutions, S_y = specific yield, S = storativity, S_w = wellbore skin, S_F = fracture skin, $r_{(w)}$ = effective well radius.

[1] Analyical solutions include the Moench double porosity solution, Theis confined, Theis confined with correction to drawdown for unconfined conditions, and Neuman unconfined with delayed gravity yield. See text for full discussion and references.

[2] Vertical to horizontal hydraulic conductivity ratio, an input parameter for these solutions, was varied from 0.1 to 1 with little change in the results.

[3] Storage, fracture skin, wellbore skin, and effective well radius are less accurate when calcualted from pumping well data than from observation wells.

[4] Specific storage was set to maximum of 0.0006 in the automatic curve matching to keep storativity in a reasonable range.

The shape of the derivative curve match for MW2b is consistent with an aquifer having radial flow and a flow dimension less than two (similar to MW1's analysis) or a no-flow boundary (Renard and others, 2009). In this case, the derivative analysis is supporting the existence of a flow-limiting boundary.

Because geologic strata overlying the aquifer at 807 are not well defined in the driller's log, I analyzed the drawdown and recovery data using the same confined and unconfined solutions I used for MW2b. The resulting curve matches and aquifer parameters for the Theis confined and unconfined solutions were similar, but the Neuman solution produced a transmissivity that was approximately 25% lower than the Theis solutions. Representative curve matches are shown on figure 22 and the numerical results are listed in table 4.

I added the boundary used in the other wells' analyses to the hydrogeologic setting for 807 and matched the same Theis and Neuman solutions again. The effect of a no-flow boundary located west of MW2b on the drawdown at 807 is theoretically less than the effect at MW2b because of 807 is farther from the boundary, and indeed, the transmissivity value calculated at 807 in the boundary scenario was only 30% to 70% greater than the no-boundary scenario (table 4), as opposed to more than doubling estimates at MW2b.

I estimated an error on the magnitude of drawdown from missing the pre-test static water-level measurement in 807 of +1.0 to -0.2 foot (+0.3 to -0.06 m). I evaluated the effect of the error by shifting the curve in AQTESOLV up by 1 foot or down by 0.2 foot in each of the analytical solution curve matches and comparing the resulting transmissivity values. If I underestimated drawdown by 1 foot (0.3 m), the transmissivity of the aquifer was overestimated by approximately 15% to 22%; if I overestimated drawdown by 0.2 foot (0.06 m), the transmissivity was underestimated by 2% to 3%.

Boundary Analysis

When the cone of depression intersects a no-flow aquifer boundary, the time versus drawdown curve becomes steeper because the well can no longer pull water from an expanding cylinder of aquifer, but instead must draw the water from within its boundaries, which makes the rate of drawdown increase. Van Tonder and others (2002, part B, p. 26) showed how to use the Cooper-Jacob equation to estimate either the distance to a boundary by noting the time at which the drawdown curve steepens or, vice versa, the time at which we expect the drawdown curve to steepen if we know the distance to a boundary.

I perceive the rate of drawdown in MW1 to change three times, as shown by the intersection of straight lines on figure 20: once at seven minutes when the pumping rate changed; once at about 400 minutes, although the change around this time was gradual; and finally again at approximately 12,000 minutes. Inserting the transmissivity and storativity of the fracture system into the rearrangement of the Cooper-Jacob equation explained by van Tonder and others (2002, part B, pg. 22) and using 400 minutes as the time at which the drawdown cone reaches the aquifer boundary, I calculate that an aquifer boundary is at approximately 4400 feet (1340 m) from Well 3. The location of the boundary I used in type curve matching, which I based on the inferred position of a buried normal fault (figure 15), is approximately 4300 feet (1310 m) from Well 3. The change in the rate of drawdown, which creates the inflection point in the time-drawdown curve, sometime happens more quickly than the gradual change observed at 400 minutes in MW1's data (Kruseman and de Ridder, 2000, p. 52). I interpret the gradual change as due to the semi-permeable and probably non-vertical nature of the boundary, which would lengthen the period of time over which the drawdown cone (actually a cylinder) intersects the boundary. This analysis supports the existence of a boundary.

We would expect to have observed drawdown in MW2b earlier than 400 minutes if the drawdown cone intersected a semi-permeable boundary located more distant from the pumping well than MW2b; however, we did not detect drawdown in MW2b until after 1500 minutes. The reason drawdown was not observed in the distal wells until later may be because the aquifer is partially unconfined and anisotropic. Using the values of transmissivity and storativity at the distal wells, which were calculated from curve matching and which are larger than at MW1, the time calculated for the cone of depression to reach a boundary located at approximately 4300 feet (1310 m) with the Cooper-Jacob equation is on the order of 5000 to 7000 minutes. The drawdown curves of the distal wells do not show the distinct inflection point sometimes noted in a bounded aquifer response, but instead, are generally very steep in late time. This lack of a definite inflection point in MW2b's and 807's data and the gradual nature of the inflection point at 400 minutes in MW1's data could be a result of the semi-permeable and probably non-vertical nature of the boundary.

Many authors (Jenkins and Prentice, 1982; Moench, 1984; Gernand and Heidtman, 1997; van Tonder and others, 2002; Renard and others, 2009; Singhal and Gupta, 2010) have shown that aquifers having boundaries often show a characteristic slope of the late-time drawdown data. Renard (2005) states that the slope of the drawdown curve doubles in late time in aquifers having no-flow boundaries, a characteristic observed in MW1's data where the slope doubles after 400 minutes and again after 12,000 minutes (figure 20). Also, a slope of 0.5 is indicative of parallel no-flow boundaries or three equidistant no-flow boundaries (van Tonder and others, 2002, part B, pg. 12), and the slope of MW1's drawdown curve in late time on a log-log plot (figure 20) is nearly 0.5. The extent and position of the Tertiary volcanic wedge is not well constrained, but the volcanic unit could easily be acting as a barrier to flow from the north or northwest. Well 3's Oquirrh Group bedrock aquifer thins to the east as less permeable Manning Canyon Shale rises to the surface on the east limb of the Lake Mountains syncline; this likely acts as an aquifer boundary. Additional boundaries in these two possible positions were included in AQTESOLV analyses, and the resulting transmissivity and storativity values were three to four times higher than with one boundary. The fact that the rate of drawdown continues to increase throughout the test supports the existence of one or more no-flow or semi-permeable aquifer boundaries.

Anisotropy

I applied a method proposed by Heilweil and Hsieh (2006) to the drawdown data from observation wells MW1, MW2b, and 807 to measure horizontal anisotropy. The method uses a simplification of Papadopulos' (1965) solution for non-steady flow to a well in an anisotropic aquifer. The method assumes the wells are placed along and perpendicular to the axis of higher permeability, which I assume at this location is the direction parallel to the axis of the small anticline on which the wells are sited, i.e., northwest to southeast. MW1 and MW2b are located along the axis of higher permeability and 807 is located perpendicular to the axis. The method uses change in drawdown over one log cycle of time (figure 23) to simultaneously solve for transmissivity in the x and y directions (T_{xx} and T_{yy}, respectively) and storativity (S). Calculations are given in appendix C. The method produced reasonable values of T_{xx} (14,000 ft^2/d [1300 m^2/d]) and T_{yy} (6000 ft^2/d [660 m^2/d]) when applied to the two distal observation wells (MW2b and 807), but did not produce reasonable results when data from the near observation well (MW1) were used. The analysis produced reasonable values of horizontal hydraulic conductivity in the high transmissivity direction (K_x = 26 ft/d [7.9 m/d]), and perpendicular to high transmissivity direction (K_y = 11 ft/d [3.4 m/d]) and aquifer storativity (0.015). The result of this anisotropy analysis is that the transmissivity in the direction parallel to the axis of folding is approximately 2.3 times higher than perpendicular to that direction.

The effect of underestimating drawdown in 807 by 1 foot (0.3 m) due to missing the static water-level measurement was slight. Aquifer parameters are different by less than 10% and the ratio of transmissivity in the x and y directions is 2.1:1.

Summary of Aquifer Parameters

Aquifer parameters estimated from analysis of the individual well responses to pumping Well 3 for five months are given in table 4 and are summarized in table 5. Table 4 presents the geometric mean of the results from up to three curve matches using different aquifer solutions (double porosity, confined, unconfined) separately for each well in an unbounded aquifer scenario and for a scenario having a no-flow boundary at the approximate location of the normal fault on the western side of the Lake Mountains. These values were evaluated for quality and applicability as discussed below and compiled into table 5.

Values for hydraulic conductivity from 807 and storage from Well 3 and MW1 were not calculated into the geometric means presented in table 5 for the following reasons. The values determined from curve matching 807's data are not as sound as the other well data because 807 was not monitored for the full duration of the drawdown test, and the solutions produced imperfect curve matches to the observed data. Storage values calculated from pumping well data are not as reliable as values calculated from observation wells (Cook, 2003, pg. 48). Original automatic curve matching to MW1 data produced aquifer storage values that were greater than 1, an impossible situation. To force the solution to have a plausible storativity value of less than 0.30 (at the top of the the range typical for an unconfined aquifer [Fetter, 1988, p. 107]), I used the software to hold the solution to a maximum specific storage of 6 x 10^{-4} per foot (2 x 10^{-4} m^{-1}).

The curve matching analysis produced fracture network hydraulic conductivity (K for MW1 and Well 3) and whole aquifer hydraulic conductivity (K for MW2b and 807) of between 19 and 39 feet per day (5.8–12 m/d) (table 4)—values within the range of reported values for fractured carbonate (Driscoll, 1986; Anderson and Woessner, 1992; Kruseman and de Ridder, 2000; Cook, 2003; Singhal and Gupta, 2010). The fracture permeability is the most applicable value to consider when evaluating a fractured aquifer because at long pumping times, the aquifer behavior is equivalent to that of a homogeneous porous medium having a permeability equal to the fracture permeability (Kruseman and de Ridder, 2000, pg. 252; Singhal and Gupta, 2010, pg. 125). The hydraulic-conductivity estimates of the matrix system (K′ for MW1 and Well 3) are on the low end of, or lower than, the typical range of hydraulic conductivity for unfractured crystalline limestone (Driscoll, 1986; Domenico and Schwartz, 1990; Anderson and Woessner, 1992; Cook, 2003).

Specific storage of the fracture system (S_s for MW1 and Well 3) was very small, as expected for a fracture network. Specific storage of the matrix or whole aquifer (S_s' on table 4) ranged from 7 x 10^{-6} to 6 x 10^{-4} per foot (2 x 10^{-6}–2 x 10^{-4} m^{-1}); however, as noted earlier, I set the specific value of 6 x 10^{-4} per foot (2 x 10^{-4} m^{-1}) at that maximum in analysis of MW1 data to keep the storativity within a realistic range. Specific yield (S_y) from the unconfined aquifer analysis ranged from 4 x 10^{-3} to 2 x 10^{-2}. Specific storage over the thickness of the aquifer combined with specific yield produced storativity (S) of between 7 x 10^{-3} to 3 x 10^{-2}, which is within the range of values typical for con-

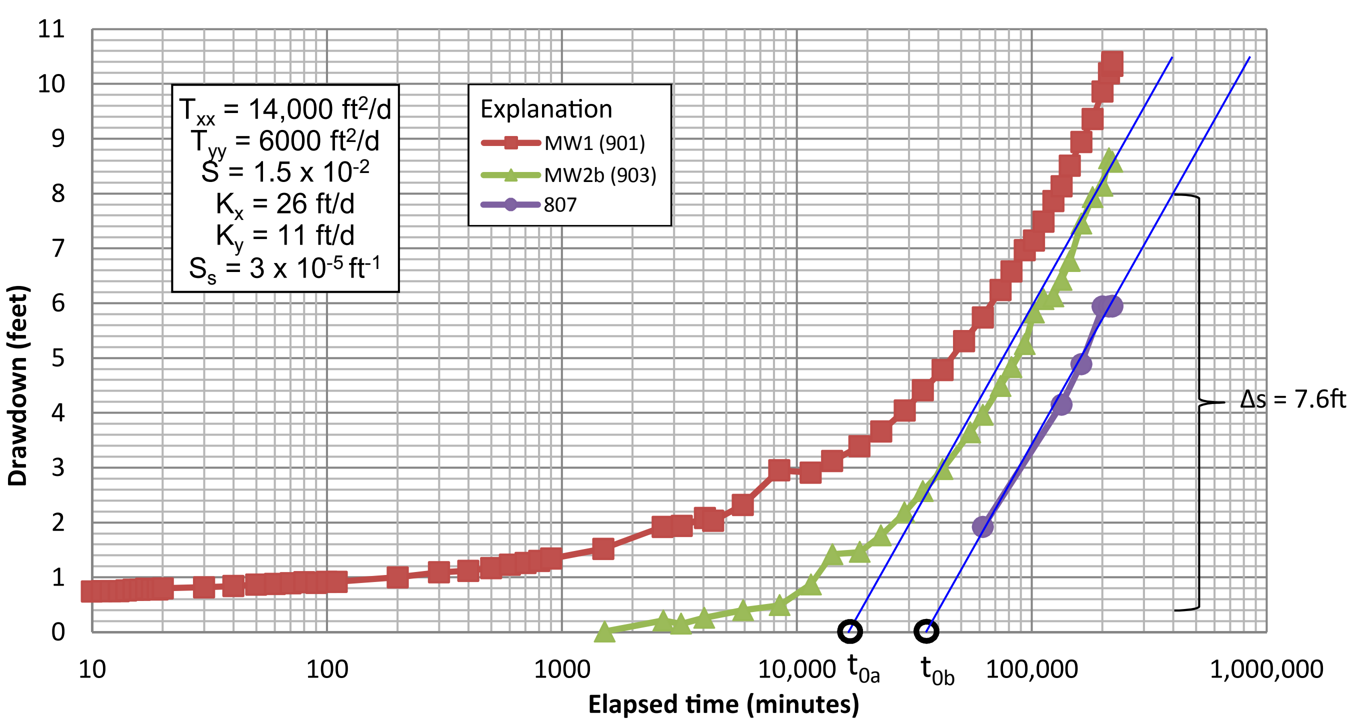


Figure 23. *Horizontal anisotropy determined from a semi-log plot of drawdown versus time in observation wells during the Well 3 aquifer test. Parallel lines are fitted to the straightest part of the late-time data. The value of drawdown over one log cycle (Δs) and x-axis intercepts of the fitted lines (t_{0a} and t_{0b}) are used as input into a simplified method for determining anisotropic transmissivity in fractured-rock aquifers (Heilweil and Hsieh, 2006). See appendix C for calculations and definitions.*

fined or semi-confined aquifers (Driscoll, 1986; Domenico and Schwartz, 1990; Anderson and Woessner, 1992; Cook, 2003; Singhal and Gupta, 2010), with the exception of the analysis from MW1, which is an artificial result.

The aquifer characteristics, choice of solutions, and method of curve matching were identical in the unbounded and bounded scenarios except that a no-flow boundary at the approximate location of the normal fault west of the Lake Mountains was added to the bounded aquifer characteristic. The effect is to increase the transmissivity in the bounded aquifer by up to two times and the storativity by approximately one order of magnitude.

Table 5. *Summary of the Oquirrh Group bedrock aquifer characteristics.*

Parameter	Estimated value
Aquifer thickness (b)	530 ft
Boundary condition	semi-permeable to the west
Hydraulic conductivity of fractures (K)	20 to 30 ft/d
Hydraulic conductivity of matrix blocks (K')	4×10^{-6} to 4×10^{-5} ft/d
Vertical to horizontal K ratio	0.1 to 0.7
Transmissivity in NW-SE direction (T_{xx})	11,000 to 16,000 ft^2/d
Transmissivity in NE-SW direction (T_{yy})	6,000 ft^2/d
Specific storage of the fractures (S_s)	3×10^{-7} ft^{-1}
Specific storage of the matrix blocks (S_s')	1×10^{-5} to 3×10^{-5} ft^{-1}
Storativity (S)	7×10^{-3} to 3×10^{-2}

The curve matching using the radial flow solutions used on MW2b and 807 test data is affected by the ratio of the vertical to horizontal hydraulic conductivity, and AQTESOLV allows the user to vary this parameter. I allowed the ratio to vary between 0.05 and 1 in order to test my assumption that fractures allow more permeability in the vertical direction than is typical for non-fractured sedimentary rock. The resulting hydraulic conductivity ratios covered the entire input range, producing inconclusive results about the nature of this aspect of the aquifer.

Additional parameters estimated with the double-porosity solutions are wellbore skin, fracture skin, and effective radius of the pumped well. Wellbore skin (S_w) is a dimensionless term indicating the resistance to flow of a lower or higher permeability zone around the wellbore, for example, decreased permeability from residual drilling mud in the formation or increased permeability due to structural fracturing (Moench, 1984). Fractured-rock aquifers often have negative wellbore skin factors, which indicate an increase in permeability around the wellbore due to fracturing (van Tonder and others, 2002, part B, pg. 19). The test data indicate a wide range for S_w of zero to 20 (table 4), providing inconclusive information about the fracture skin, but suggesting that permeability around the well is not enhanced compared to the rest of the aquifer.

Fractures may have mineral coatings that have a lower permeability than the matrix blocks, which was defined mathematically by Moench (1984) as a dimensionless fracture skin, or S_F. A zero value of S_F means the skin is not present or not hindering the flow of water out of the matrix block into the fractures; a large S_F (~10) indicates the skin is significantly hindering flow from the matrix to the fractures (Moench, 1984). The results of the Well 3 aquifer-test analysis showed a S_F of 0.8 to 1, or a slight hindrance of flow from matrix to fractures. The lack of a tight fracture skin may allow water to flow from the matrix blocks to the fractures in early time, inflating the hydraulic-conductivity value. For this reason, the K value of the fracture network derived from Well 3 and MW1 data are more likely to be slight overestimates.

The effective radius of the borehole ($r_{(w)}$) can be increased by natural fractures that intersect the wellbore or by well development (Gringarten and others, 1974; van Tonder and others, 2002, part B, pg. 19; Singhal and Gupta, 2010, pg. 172). The drillers reported many borehole caving problems during drilling Well 3, using a 22-inch-diameter (56 cm) drill bit (appendix B). The well has a 16-inch-diameter (41 cm) screen and coarse filter pack (appendix B). The effective radius determined from aquifer-test data curve matches ranged from 0.8 to 3 feet (0.2–0.9 m), indicating that caving and interconnected fractures near the wellbore may have increased the well radius slightly.

Table 5 presents a summary of the aquifer characteristic, including the determination of anisotropy in the direction parallel to the structural fold axes. The values on table 5 are compiled by taking the geometric mean of the most reliable values in table 4. Both unbounded and bounded results are included in the geometric mean to produce an intermediate result consistent with my interpretation of a semi-permeable boundary at the fault and a limited-extent no-flow boundary at the location of volcanic rock wedge northwest of the well. Transmissivity and storativity for the Oquirrh Group fractured carbonate and sandstone aquifer are that of an aquifer that can produce as much as a few thousand gallons of water per minute given the proper well design, but the existence of one or more boundaries limits the longevity of production.

Implications for Groundwater Flow

My analysis of data from the Well 3 aquifer test shows that the fractured-rock aquifer in the Pennsylvanian-Mississippian-aged carbonate and clastic rocks of the Oquirrh Group (Butterfield Peaks Formation and West Canyon Limestone) has moderate to high transmissivity and storativity, and that the transmissivity is greater in the northwest to southeast direction, parallel to the axis of structural folding. The aquifer has fractures that are well enough connected along that axis to yield both linear flow to the well and radial-type drawdown across at least a 2-square-mile (≥5 km^2) area. The extent of the aquifer, however, is limited by geologic units (a volcanic rock wedge) and/or geologic structures (most likely the normal fault on the western side of the Lake Mountains), which restrict the flow from the Cedar Valley basin-fill aquifer to the west. Although I am unable to show through aquifer-test analysis the exact location and nature of aquifer boundaries and thus the path of recharge to the aquifer, potentiometric data presented in Jordan and Sabbah (2012) show that the area coinciding with the wedge of volcanic rocks and what may be a northern extension of the inferred normal fault bounding the western side of the Lake Mountains is where basin-fill groundwater is forced to find another path to the area having lower potentiometric levels east of Cedar Pass. Possible pathways are (1) north or south around this area, (2) downward into bedrock underlying the basin fill, or (3) slowly through the barrier. The water level in the volcanic wedge is higher than in either the basin fill or the bedrock (Jordan, 2008), so no basin-fill water is moving through it. Other limited potentiometric and water-quality data (Jordan and Sabbah, 2012) suggest that at least some of the flow goes south of the area, through a more permeable section of the fault. Without better potentiometric data to help understand this complex hydrogeologic setting, my opinion is that groundwater from the basin fill travels around the volcanic wedge and the northern part of the fault in north, south, and downward directions to reach the bedrock aquifer.

WEST CANYON LIMESTONE FRACTURED-ROCK AQUIFER TEST

Introduction

In March and April 2007, UGS personnel, with cooperation from Eagle Mountain employees, conducted a constant-rate aquifer test on Eagle Mountain Municipal Supply Well 2 (herein referred to as "Well 2" but also given UGS ID 156). The Utah Division of Water Rights requested the test to provide information on the fractured-bedrock aquifer in the Mississippian-age Great Blue Limestone in this area of recent groundwater resource development at Cedar Pass. The purpose of the test was to determine the transmissivity and storage properties of the fractured-bedrock aquifer while pumping for a long period of time (35 days) at this well's typical flow rate of 2500 gallons per minute (158 L/s). We conducted an 11-day recovery test after the drawdown test.

Well 2 is located near Cedar Pass in northwestern Utah County, Utah (figure 15). The well is on the golf course of the Ranches Golf Club in the SE1/4 sec. 17, T. 5 S., R. 1 W., SLB&M. Well 2 has a 14-inch-diameter (36 cm), 0.08-inch-slot (0.02 cm) louvered screen from 510 to 810, 820 to 860, and 900 to 940 feet (155–247, 250–262, and 274–287 m) below ground surface (figure 24). I used a production well operated by the Ranches Golf Club as the observation well. This well, given UGS well identification number 997, is located 660 feet (201 m) west-northwest of Well 2. Observation well 997 has perforated 12-inch-diameter (30 cm) casing from 650 to 700 feet (198–213 m) and perforated 8-inch (20 cm) casing from 920 to 1040 feet (280–317 m), below which the hole was left open to the total depth drilled of 1308 feet (399 m).

Geologic Setting

Geologic mapping by Biek (2004, 2005) and Biek and others (2005) shows folded Mississippian-aged Great Blue Limestone outcrop surrounding the well (figure 15). Bedding dips southwest at approximately 30 degrees (Montgomery Watson Harza, 2001) to 45 degrees (Biek, 2005). The Great Blue Limestone is as much as 2500 feet (760 m) thick (Biek, 2004) in this area and is present at the surface or under shallow unconsolidated Quaternary deposits at least 1 mile horizontally in all directions from the well, except to the southwest where it dips beneath the Pennsylvanian to Mississippian-aged Manning Canyon Shale and Quaternary sediments approximately ½ mile from the well. Biek (2004, 2005) and Biek and others (2005) inferred a northwest-southeast striking thrust fault near the well; previous mapping identified this structure as an anticline (Moore, 1973). Drillers' logs of Well 2 (appendix B), 997, and a water exploration well drilled for Eagle Mountain in 1999 approximately ¼ mile west of Well 2 show gray limestone with shale units of varying thickness to at least 2000 feet (610 m). Based on outcrop, well logs, and interpreted subsurface structure (Montgomery Watson, Inc., 2000; Montgomery Watson Harza, 2001; Biek, 2004, 2005), both Well 2 and 997 are completed in the Mississippian Great Blue Limestone.

Minimum thicknesses of the portion of the aquifer supplying water to Well 2 and 997 are 500 feet (152 m) and 525 feet (160 m), respectively. The aquifer thickness at Well 2 was calculated from the bottom of a 90-foot-thick (27 m) shale unit at 440 feet (134 m) deep in the well to the top of a 550-foot-thick (168 m) carbonaceous limestone and shale section of the Great Blue Limestone identified at 940 feet (287 m) in the deep test hole for Well 2. I calculated aquifer thickness at 997 from the bottom of a 315-foot-thick (96 m) shale unit at 500 feet (152 m) deep in the well to the bottom of the perforated interval at 1040 feet (317 m), not including 15 feet (4.6 m) of shale from 685 to 700 feet (209–213 m). The well driller indicated that 997 was left as an open hole from 1040 to 1310 feet (317–399 m), but he believed, based on during-drilling water production, that the open hole section of the well was contributing less than approximately 150 gallons per minute (10 L/s) (R. Peterson, Aqua Design well drilling, verbal communication, May 21, 2008).

Based on my review of the well logs, I classify the aquifer as a confined, fractured, mostly homogeneous limestone aquifer. Fracture permeability is more likely to be greater in a northwest to southeast direction, which is parallel to the strike of bedding and axes of structural folding and thrust faulting. Without multiple observation wells, however, the suspected anisotropy cannot be verified (Papadopulos, 1965; Jenkins and Prentice, 1982; Moench, 1984; Heilweil and Hsieh, 2006).

Aquifer Test Setup

Neither Well 2 nor 997 had been pumped in the four months preceding the test. The next nearest production well to Well 2, located in a gravel pit ⅔ mile (3770 feet [1149 m]) northeast of the aquifer-test site, may have been pumped at as much as 300 gallons per minute (19 L/s) intermittently during the test; however, production records for this well were not available. To test the effect of pumping the gravel pit well on drawdown at 997, I simulated a well pumping 300 gallons per minute (19 L/s) for half of each day at the gravel pit in my aquifer-test data analysis. Type curve matching using AQTESOLV (Duffield, 2003) using observed water levels and adding the gravel pit well produced a primary hydraulic conductivity 3% higher and specific storage 5% higher than without the gravel pit well pumping (because the aquifer is producing more water given the observed drawdown); these values are within the error associated with normal aquifer-test analysis.

Eagle Mountain's permanent in-line magnetic flow meter measured the discharge from the pumping well. I checked the accuracy of the city's meter using a portable Controlotron Stormmeter 1010 Uniflow Universal flow meter, and measurements

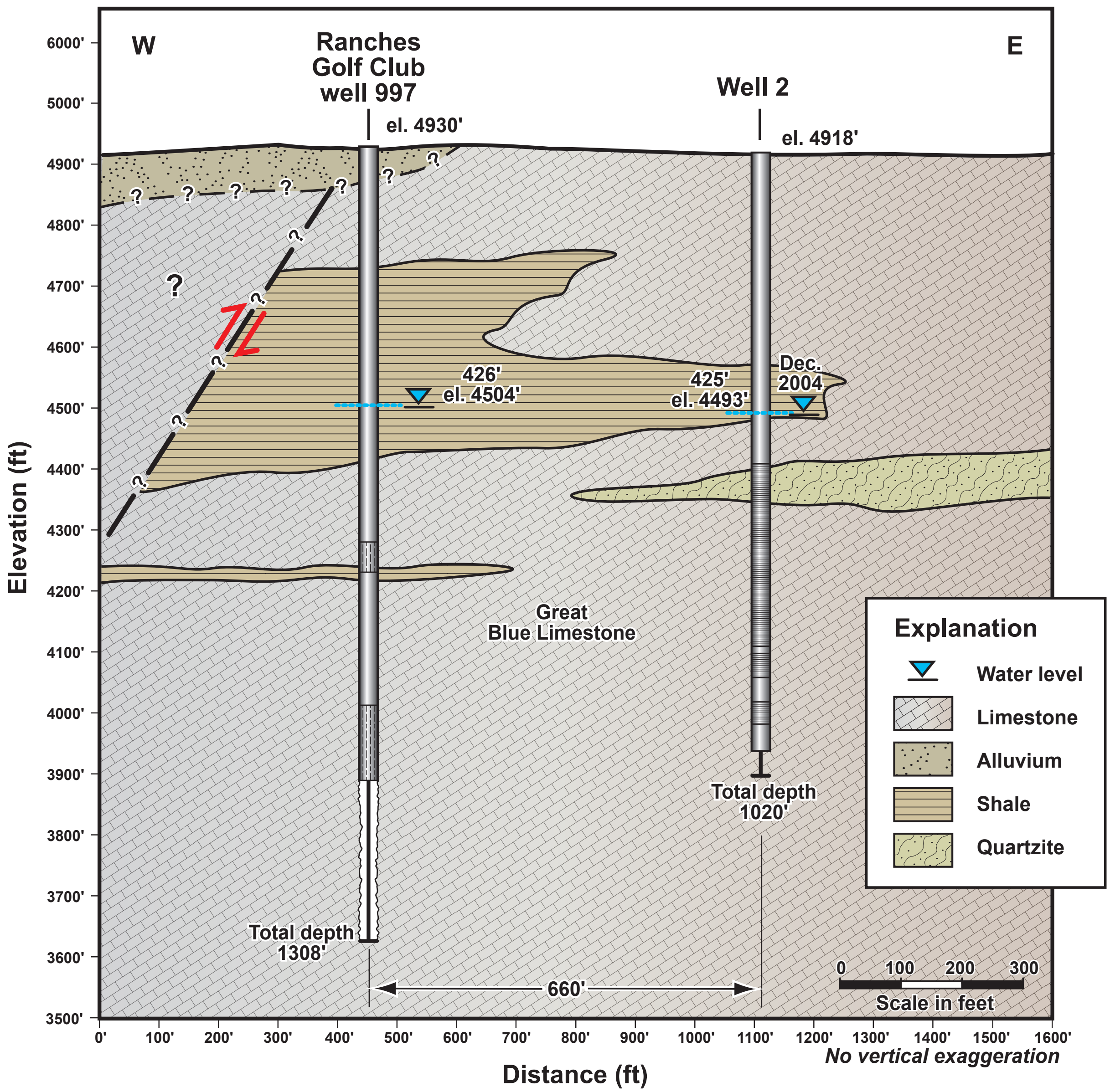


Figure 24. *Geologic cross section through production and observation wells for the aquifer test on Well 2.*

were less than 5 gallons per minute (0.3 L/s) different at a flow rate of 2060 gallons per minute (130 L/s), indicating an error of less than 1%. Discharge was recorded manually from the permanent flow meter at approximately 10-minute intervals at the start of the test and daily after the first day. For most of the test, approximately 1500 gallons per minute (95 L/s) of the discharge was routed to city water lines and 1000 gallons per minute (63 L/s) was discharged to a permanent clay-lined 6.5-million-gallon (24,600 m^3) retention pond located 1600 feet (490 m) south of the well. During approximately half the days of the test, the golf course used 200,000 to 300,000 gallons per day (760–1140 m^3/d) from the pond to irrigate the course, including two fairways running between the two wells. Pond and irrigation seepage is unlikely to have affected the water level in the observation well or pumping rate from Well 2 because the clay-lined pond is down gradient from the test site and irrigation seepage must penetrate a 428-foot-thick (130 m) unsaturated zone, including at least 90 feet (27 m) of shale at both wells, to reach the aquifer.

Eagle Mountain personnel started the pump in Well 2 at 9:54 AM on March 29, 2007. Discharge varied by approximately 100 gallons per minute (6 L/s) for the first 600 minutes of the test while water filled pipelines and flowed to the retention pond (appendix A, table A-6). Average discharge was approximately 2533 ±10 gallons per minute (160 ±0.6 L/s) for the majority of the test. The well pump was off for approximately 2.5 hours on day 13, but resumed steady flow for the next 17 days. During days 30, 31, and 32, water was intermittently routed completely into the city water lines to meet demand, which decreased the well discharge to approximately 2075 gallons per minute (131 L/s) when pumping into the city lines. For the last three days of the test, all water was routed to the retention pond, which increased the flow to approximately 2480 gallons per minute (157 L/s). Changes in discharge were incorporated by the AQTESOLV software (Duffield, 2003) into curve matching analysis of the aquifer-test data.

Water-Level Observations

Water levels in 997, measured manually using an electronic water-level sounder twice per week for 31 days prior to the test, showed no antecedent trends (figure 25). Static water level at the start of the test was 426 feet (130 m) below land surface or 4492 feet (1369 m) AMSL. Water levels in 997 (appendix A, table A-7) were measured during the 35-day test and for an 11-day recovery period following the test until the pump in 997 was turned on to supply water to the golf course. Water-level measurement was not possible in Well 2 during the aquifer test due to well-head configuration. Historically, 45 feet (14 m) of drawdown in Well 2 was recorded during an aquifer test in 2000 after three days of pumping 2200 gallons per minute (139 L/s) (Montgomery Watson Harza, 2001).

Maximum drawdown during the UGS aquifer test in 997 located 660 feet (201 m) from the pumping well was 17.45 feet (5.33 m), and, although the test did not reach steady state, the rate of water-level decline had slowed to approximately 0.04 foot (0.01 m) per day until the last few days of the test when changes in discharge affected the water level. Water levels in the observation well were not corrected for possible barometric well function because the magnitude of drawdown was

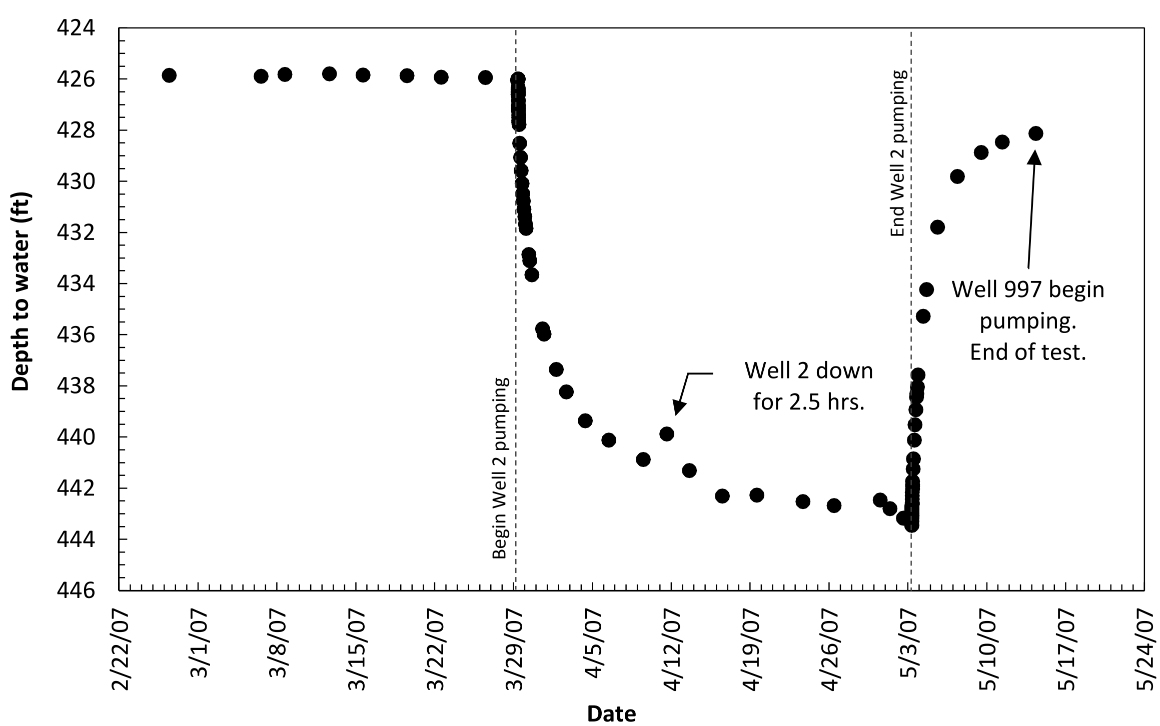


Figure 25. *Water-level response in the Ranches Golf Club well (997) during Well 2 aquifer test.*

much larger than the effects of barometric efficiency, which is typically less than 0.2 foot (0.06 m) per day or 1 foot (0.3 m) per week with a large weather system (Hare and Morse, 1997; Rasmussen and Crawford, 1997; Spane, 2002; Halford, 2006).

Data Interpretation

The drawdown response observed in 997 (figure 26) is similar to the response expected in an unconfined aquifer having delayed yield or a double-porosity aquifer (in this case, a fractured, consolidated-rock aquifer with high permeability in the fracture system and low permeability in the matrix blocks)—that is, a Theis-type drawdown curve in early time joined by a period of slower drawdown to another Theis-type drawdown curve in late time (Boulton, 1963; Neuman, 1972; Bourdet and Gringarten, 1980; Moench, 1984; Kruseman and de Ridder, 2000). Geologically, the aquifer is a confined, fractured-bedrock aquifer. I interpret the early-time data (approximately 0 to 300 minutes plotted on log scale on figure 26) to be the drawdown response of the fracture system. Wellbore storage, if it was a factor in this test, would only affect the first 10 minutes of the drawdown data as calculated by a simple formula relating well diameter and the transmissivity of the aquifer (Papadopulos and Cooper, 1967). The slope of the early-time data, however, is 0.33, not 1 as would be the case if the wellbore storage was a factor (van Tonder and others, 2002, part B, pg. 15; Renard and others, 2009). I interpret the late-time data (after 1000 minutes) to be the response of the aquifer as a whole, in which water comes from storage in both the fractures and the matrix blocks and is transmitted to the well via the transmissivity of the fracture system.

I applied a double-porosity aquifer model (Moench, 1984) having 3-foot-diameter (1 m) spherical blocks to the 997 drawdown data (figure 26). I chose a matrix block diameter of 3 feet (1 m) because, in outcrop near Well 2, vertical jointing and shallowly dipping bedding planes appeared roughly equally distributed, dividing the outcrop into blocks 2 to 3 feet (0.6–1 m) wide. I estimate the ratio of horizontal hydraulic conductivity, K_h, to vertical conductivity, K_v, in this fractured limestone aquifer composed of roughly cubic blocks to be close to 1. If K_h/K_v is less than 1.5, 997 is far enough away from Well 2 that drawdown at 997 is not affected by Well 2 only partially penetrating the aquifer (Hantush, 1961), so the data were analyzed without taking partial penetration into account.

Because the drawdown curve showed no inflection point typical of the cone of depression encountering an aquifer boundary, the extent of the aquifer appears to be large enough that no boundary was encountered over the 35 days of pumping at this well. Therefore, the thrust fault and the thinning of the aquifer to the southwest as Great Blue Limestone dips under the Manning Canyon Shale do not appear to be affecting drawdown during pumping periods of moderate length, and I assumed the aquifer to be of infinite areal extent for the purposes of data analysis.

Bourdet and Gringarten (1980) showed that with increased distance of the observation well from the pumping well, the effects of double porosity are not detectable in the drawdown response. Double-porosity effects should be observed if λ (a function of the block geometry, the distance from the pumping well, and the hydraulic conductivities of the fracture system and blocks) is less than 1.78 (Bourdet and Gringarten, 1980). For this setting, λ is in the range of 1 to 2, which would indicate that the drawdown observed in 997 could show a double-porosity response.

Aquifer parameters estimated using the double-porosity solution (Moench, 1984) are given in table 6 and on figure 26. As expected, the hydraulic conductivity of the fractures (K) is high and the hydraulic conductivity of the limestone blocks (K′) is low. The values fall near the top and bottom, respectively, of the range of reported values for limestone (Driscoll, 1986; Domenico and Schwartz, 1990; Anderson and Woessner, 1992). I expected higher hydraulic conductivity (K) and lower specific storage (S_s) of the fractures as

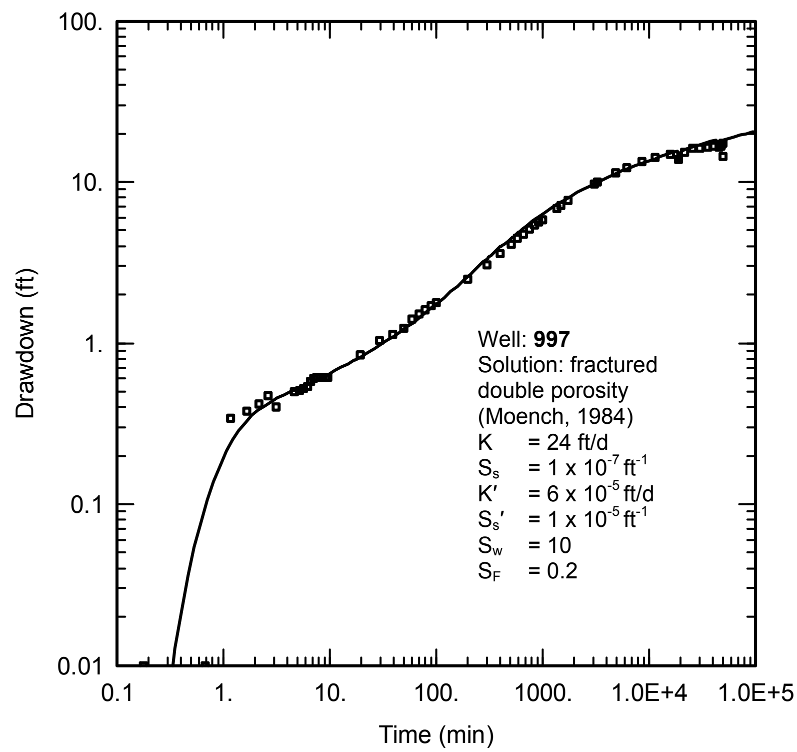


Figure 26. *Curve match to observation well 997 data using a double-porosity aquifer analytical solution. Parameters are defined in the text.*

compared to the matrix blocks (K′ and S_s′, respectively) (Moench, 1984; Kruseman and de Ridder, 2000), and my estimates confirmed this relationship. Additional parameters estimated with this method are wellbore skin (S_w) of 10, indicating a skin around the wellbore (such as from residual drilling mud or well fouling) which significantly hinders flow into the well, and a fracture skin (S_F) of 0.2 indicating a thin skin on the matrix blocks that slightly inhibits flow from the matrix to the fractures (Moench, 1984).

Table 6. *Great Blue Limestone bedrock aquifer characteristics estimated from an aquifer test on Well 2.*

Parameter	Value
Aquifer thickness (b)	525 ft
Hydraulic conductivity of fractures (K)	24 ft/d
Hydraulic conductivity of matrix blocks (K')	6×10^{-5} ft/d
Transmissivity (T)	1.3×10^{4} ft^2/d
Specific storage of the fractures (S_s)	1×10^{-7} ft^{-1}
Specific storage of the matrix blocks (S_s')	1×10^{-5} ft^{-1}
Storativity (S)	7×10^{-3}

I applied several other analytical solutions to the drawdown data to test the assumption that the aquifer is a double-porosity system. Curve matches using the Cooper-Jacob straight line method (Cooper and Jacob, 1946) applied to the late-time drawdown data (>1000 minutes) and Theis recovery method (Theis, 1935) applied to the late-time recovery data produced transmissivity estimates that are in good agreement with the estimates of the aquifer as a whole from the double-porosity model. Matching the Cooper-Jacob solution (1946) to early-time data is not appropriate because wellbore storage effects, double porosity, and well skin all effect early-time data, and the Cooper-Jacob assumption is not met in early time. The Theis (1935) solution for homogeneous isotropic aquifer did not produce satisfactory curve matches. The drawdown response in 997 did not fit the response expected from a well drawing water from one large fracture experiencing linear flow to the fracture plane (Jenkins and Prentice, 1982; Smith and Vaughan, 1985).

REANALYSIS OF SEVEN AQUIFER-TEST DATA SETS

To supplement the results from UGS aquifer testing, I analyzed or reanalyzed data sets from seven aquifer tests conducted by others mostly in the late 1990s and early 2000s. I reanalyzed available single-well aquifer-test data for wells near Cedar Fort and Cedar Pass using AQTESOLV (Duffield, 2003, 2007) and appropriate solutions (Theis, 1935; Moench, 1984). One data set was for a six-day test and the others were 24-hour tests. The quality of the data sets was generally as good as UGS-conducted single-well aquifer tests, and more accurate than specific-capacity data. I included the hydraulic conductivity and transmissivity values calculated from these tests in my evaluation of hydraulic properties of the Cedar Valley study area.

Eagle Mountain Well 1 in Basin Fill and Paleozoic Bedrock (ID 1018)

I reanalyzed data from a step drawdown test that was conducted in 1980 on Eagle Mountain Municipal Supply Well 1, also known as the John Walden well and given UGS ID 1018. The well is located in sec. 18, T. 6 S., R. 1 W, SLB&M (figure 1). The well was drilled to 475 feet deep (145 m) through 349 feet (106 m) of unconsolidated sediment and 126 feet (38 m) of fractured limestone. The well has un-perforated 16-inch-diameter (41 cm) casing to 355 feet (108 m) and open hole in bedrock to the total depth of 475 feet (145 m). A gravel pack is installed around the un-perforated casing from 120 feet (37 m) to the bottom of the casing. In theory, groundwater is able to enter the annular space from the unconsolidated aquifer and the top of the fractured-rock aquifer and flow down through the gravel pack to the open interval of the well.

The pumping rates during the 24-hour step test were 1560, 2590, and 3880 gallons per minute (98, 163, and 245 L/s) with corresponding drawdowns of 2, 6, and 12 feet (0.6, 1.8, and 3.7 m). Another large production well located approximately 60 feet (18 m) southwest of Well 1 experienced no drawdown during this test. Using the Theis method for analysis of step-test data (Theis, 1935; Rorabaugh, 1953; Hantush, 1961), I estimated a transmissivity of 32,000 feet squared per day (3000 m^2/d). If I use an aquifer thickness of 264 feet (81 m), which is the thickness from the water table to the bottom of the well and includes 138 feet (42 m) of unconsolidated boulders, cobbles, gravel, and clay, the average hydraulic conductivity of the combined basin fill and fractured limestone is 122 feet per day (38 m/d). If I assume all discharge comes from the bedrock, the calculated hydraulic conductivity of the bedrock is 254 feet per day (77 m/d).

The consulting geologist on this test calculated the transmissivity of the fractured bedrock at this well at 93,600 feet squared per day (8700 m^2/d) based on its specific capacity, but used a transmissivity value of 21,500 feet squared per day (2000 m^2/d) in travel time calculations for a drinking water source protection plan (DWSPP) report (Montgomery, 1995). The lower transmissivity was derived by averaging the transmissivity of the Cedar Valley basin-fill aquifer taken from Feltis (1967) and the fractured-bedrock transmissivity he calculated from specific capacity.

Cedar Fort Community Center Well in West Canyon Limestone (ID 847)

A DWSPP required by the Utah Division of Drinking Water contained data from a single-well step drawdown test conducted in 2001 on the Cedar Fort Community Center Well, UGS ID 847, located in Cedar Fort in the NE1/4 sec. 7, T. 6 S., R. 2 W., SLB&M (figure 1) (ESI Engineering, Inc., 2003). Based on the driller's log and surface geology in the foothills of the Oquirrh Mountains ¼ mile west of the well, the well is completed in the Pennsylvanian-Mississippian West Canyon Limestone. The 6-inch-diameter (15 cm) PVC well casing has saw-cut perforations from 380 to 410 feet (116–125 m) below ground and an open hole below that to a depth of 465 feet (142 m).

A transmissivity of 441 feet squared per day (41 m^2/d) derived from a constant-rate pumping test was reported in the DWSPP (ESI Engineering, Inc., 2003). The test data were not available to me, so I analyzed the step drawdown test data provided in the DWSPP using automatic type curve matching (Duffield, 2007). The maximum drawdown after nine hours of pumping, during which the pumping rate was increased in six steps to a maximum rate of 42 gallons per minute (2.6 L/s), was 60 feet (18 m). Each time the pumping rate increased, the water level quickly drew down to a new level and stabilized. This response is typical of an inefficient well in an aquifer that is not fully stressed. Curve matching produced poor matches to the step test data and a transmissivity of 130 feet squared per day (12 m^2/d) and hydraulic conductivity of 1.5 feet per day (0.5 m/d). Because of the poor match, I consider the transmissivity and hydraulic conductivity of 441 feet squared per day (41 m^2/d) and 7.7 feet per day (2.3 m/d), respectively, reported from the constant rate test (ESI Engineering, Inc., 2003) to be better estimates, and have included them in table 1.

Cedar Fort Artesian Well in Paleozoic Bedrock (ID 832)

A well given UGS ID 832 was drilled in 2001 in the mountains northwest of Cedar Fort (figure 1). Based on the driller's log, the well penetrated Tertiary volcanic rocks, Oquirrh Group strata, the Mississippian Manning Canyon Shale, and is open to the Mississippian Great Blue Limestone from 818 to 1280 feet (249–390 m) below surface. However, Jordan and Sabbah (2012) conclude, based on water chemistry data, that the well is more likely completed in Oquirrh Group bedrock. There may be concealed geologic structures that position Oquirrh Group bedrock below Manning Canyon Shale, or the unit identified as Manning Canyon Shale on the driller's log may be a shale within the Oquirrh Group.

Well 832 had an artesian pressure head of 409 feet (125 m) above ground after completion. I analyzed data from a shut-in test conducted after the well was allowed to flow for approximately six days. The discharge during the six-day flowing period decreased from 680 to 608 gallons per minute (43 L/s to 38 L/s). Water level was 178 feet (54 m) above land surface two minutes after the well was shut in and had recovered to within 40 feet (12 m) of full recovery after six days of being shut in, but took 33 days to fully recover. I analyzed the data as a simple recovery test and a constant head test using AQTESOLV (Duffield, 2007) and solutions for confined aquifers. I estimated the transmissivity for the Paleozoic bedrock aquifer in this location to be between 200 and 470 feet squared per day (19–44 m^2/d) and the hydraulic conductivity to be between 0.4 and 1 foot per day (0.1–0.3 m/d) from these data.

Eagle Mountain Well 2 in Great Blue Limestone (ID 156) 24-hour Test

Before the UGS worked with Eagle Mountain to conduct long-term aquifer tests on Well 2 and Well 3, which I have reported on above, I analyzed available data from single-well pump tests on both wells. A constant rate test conducted on Well 2 and reported by a consulting firm (Montgomery Watson Harza, 2001) showed drawdown of 43 feet (13 m) after pumping 2200 gallons per minute (139 L/s) for approximately two days. The consultants reported a transmissivity value of 15,500 feet squared per day (1440 m^2/d) derived by the Cooper-Jacob straight-line method (Cooper and Jacob, 1946). When I analyzed these data, I matched the Cooper-Jacob straight-line method to late-time data and also applied a fractured-rock aquifer solution (Moench, 1984). These analyses produced transmissivity estimates of 18,000 and 23,000 feet squared per day (1700–2100 m^2/d), which are slightly higher than the value of 13,000 feet squared per day (1200 m^2/d) determined from the long-term test conducted by UGS in 2007.

Eagle Mountain Well 3 in Oquirrh Group Bedrock (ID 992) 24-hour Test

Lang Exploratory Drilling (written communication, January 19, 2005) provided me with data from a 24-hour test conducted after their completion of Well 3. The test was conducted using a temporary pump discharging 3800 gallons per minute (240 L/s) (almost double the rate the permanent pump would discharge later). The drawdown after 24 hours was approximately 140 feet (43 m) (as compared to 41 feet [13 m] after five months of pumping at 1960 gallons per minute [124 L/s]). I used the Cooper-Jacob straight line method (Cooper and Jacob, 1946) to match late-time data after radial flow assumptions had been met and determined an aquifer transmissivity of approximately 15,000 feet squared per day (1440 m^2/d), similar to what I derived for the high transmissivity direction from the long term test conducted in 2007. At approximately 400 minutes into the test, the rate of drawdown increased slightly, an indication that the drawdown in the pumping well at this high rate may have been affected by the semi-permeable boundary described in the long-term test discussion.

Harvest Haven Irrigation Well in Great Blue Limestone (ID 1003)

A DWSPP contained data from a single-well aquifer drawdown and recovery test conducted in 1997 on the Harvest Haven Irrigation Well (UGS ID 1003) located north of Cedar Pass in the NE1/4 sec. 18, T. 5 S., R. 1 W., SLB&M (figure 1) (Mower and Allred, 1997). According to the well driller's log, the well is completed in interbedded shale, quartzite, and limestone of the Manning Canyon Shale, but according to the DWSPP it is completed in the Great Blue Limestone (Mower and Allred, 1997). The well has perforated casing from 671 to 683 feet (204–208 m), from 720 to 740 feet (219–226 m), and may be uncased from 760 to 830 feet (232–253 m). Manning Canyon Shale and Tertiary volcanic rocks confine the aquifer (Mower and Allred, 1997). The 24-hour test consisted of pumping the well at a rate of 501 gallons per minute (32 L/s) for 85 minutes before the rate was reduced to 309 gallons per minute (20 L/s) to avoid drawing the water level below the pump intake. The water level at the end of 24 hours was approximately 188 feet (57 m) below the static water level and was declining at a rate of more than 4 feet (1 m) per hour. The consultant analyzed the recovery portion of the test data and reported a hydraulic conductivity of 1 foot per day (0.3 m/d) and an aquifer transmissivity estimate of 98 feet squared per day (9.1 m^2/d), given an aquifer thickness of 95 feet (29 m) (Mower and Allred, 1997). I reanalyzed the drawdown and recovery data together by type curve and derivative analysis using AQTESOLV (Duffield, 2007). I used three different solutions to attempt to fit the atypical drawdown curve: (1) a double-porosity fractured aquifer (Moench, 1984), (2) a confined aquifer with variable rate pumping (Dougherty and Babu, 1984), and (3) a single vertical fracture (Gringarten and Witherspoon, 1972). None of the solutions provided an ideal fit to the drawdown data, but all produced aquifer transmissivity estimates that were lower than reported by Mower and Allred (1997) from the recovery data. The best matches were made using the fractured-rock solutions, which gave hydraulic-conductivity estimates of 0.3 to 1 foot per day (0.09–0.3 m/d). I used a more conservative aquifer thickness of 62 feet (19 m) based on the driller's log and calculated that the transmissivity of the Great Blue Limestone aquifer at this location may be in the range of 20 to 70 feet squared per day (1.9–6.5 m^2/d). The lower of these estimates is reported in table 1.

The last data set examined was not a specific aquifer test. I reviewed water-level and production data from the Harvest Haven Irrigation Well (ID 1003) recorded over two seasons of seasonal well use (A. Allred, Harvest Irrigation Co., written communication, September 5, 2008) and analyzed the water level recorded over the winter recovery period as a recovery test. Between July and November 2004, the well pumped intermittently at rates between 9 and 78 gallons per minute (0.6–4.9 L/s), averaging 30 gallons per minute (1.9 L/s) over that time period. The well had approximately 140 feet (43 m) of drawdown at the end of the 2004 irrigation season. The well operator recorded water levels throughout the winter, and by mid April 2005, after 5½ months of not pumping, the well had only recovered to within 51 feet (16 m) of the pre-2004 season static level. I used the straight-line Theis solution for a recovery test in a confined aquifer (Theis, 1935; Birsoy and Summers, 1980) and matched the late-time recovery data to yield an aquifer transmissivity estimate of approximately 4 feet squared per day (0.4 m^2/d). This estimate is lower than the estimate from the 24-hour variable rate test and considerably lower than the transmissivity reported in the DWSPP (Mower and Allred, 1997), but given the extremely slow and incomplete recovery observed in the well over two irrigation seasons, this low transmissivity is not unreasonable.

TRANSMISSIVITY AND HYDRAULIC-CONDUCTIVITY ESTIMATES FROM SPECIFIC CAPACITY

Theis and others (1963) derived equations for confined and unconfined aquifers to approximate the transmissivity using the specific capacity of a well given an estimated storativity. The specific capacity, SC, of a well is an expression of the productivity of the well and the aquifer near the well; it is calculated by dividing the discharge rate, Q, by the drawdown, s, in the well. I evaluated specific-capacity data from 95 wells in Cedar, Goshen, and northern Utah Valleys, 70 of which are within the Cedar Valley groundwater basin, using pumping and drawdown information from well drillers' logs. I input the specific-capacity data into a computer spreadsheet program (Cobb, 2005) based on a computer program developed by Bradbury and Rothschild (1985) that estimates hydraulic conductivity and transmissivity from specific-capacity data using the Cooper-Jacob (1946) approximation of the Theis (1935) equation for transient radial flow to a well (equation 1).

$$T = \frac{Q}{4\pi s}\left[\ln\left(\frac{2.25Tt}{r^2 S}\right)\right] \qquad \text{Eq. 1}$$

Where:

T = transmissivity
Q = pumping rate
s = drawdown
t = duration of test
r = radius of well
S = aquifer storativity

Aquifer thickness is needed to calculate hydraulic conductivity, K, from transmissivity, T. I used the length of the well screen or open interval as the aquifer thickness unless there was some indication that water was entering the well from intervals above or below the well screen. The equations and spreadsheets require a storativity value to work. I assumed values of storativity based on the few aquifer-test derived values available and generally accepted literature values (0.01 to 0.3 for unconfined aquifers, 0.00005 to 0.005 for confined aquifers, and 0.005 to 0.01 for leaky aquifers), which are generally accurate within one order of magnitude (Mace, 2001).

Estimates of hydraulic conductivity and transmissivity from specific capacity are affected by partial penetration, well loss, decrease in saturated thickness, and well development (Mace, 2001) and are generally lower than hydraulic-conductivity values calculated from aquifer tests due to friction loss and other well losses (Fetter, 1988, p. 172). However, Huntley and others (1992) concluded that the reverse is true for fractured-rock aquifers because the fractures often increase the effective radius of the well. Although estimates of transmissivity derived from specific capacity are not as accurate as those derived from aquifer tests, when evaluated with other hydrologic data, they provide reasonable estimates of aquifer characteristics. Table 1 lists the transmissivity and hydraulic conductivity calculated from specific-capacity data.

The transmissivity values calculated from specific capacity for the 95 wells in the Cedar Valley study area ranged from 0.3 to 1.2 x 10^5 feet squared per day (3 x 10^{-2}–1.1 x 10^4 m^2/d) with a geometric mean of 270 feet squared per day (25 m^2/d). The hydraulic-conductivity values from specific capacity for the wells in the Cedar Valley study area ranged from 3 x 10^{-3} to 530 feet per day (9 x 10^{-4}–1.6 x 10^2 m/d) with a geometric mean of 3 feet per day (0.9 m/d). The ranges and means for the 70 wells within Cedar Valley proper were similar, indicating that the data from wells outside the valley can be used to determine regional aquifer properties.

SUMMARY AND CONCLUSIONS

Aquifer tests and specific-capacity data in the Cedar Valley study area provide valuable information from which I was able to derive estimates of the hydraulic properties of the principal basin-fill aquifer and the fractured-bedrock aquifer. Aquifer testing was conducted on the two most important aquifers in the study area, the principal basin-fill aquifer and the fractured-bedrock aquifer. Specific capacity data were available from these and lesser-used aquifer units, and were combined with aquifer-test results to characterize the transmissivity and storativity of the aquifers in Cedar Valley.

Summary of the Three Basin-Fill Aquifer Tests

The first test conducted on the principal basin-fill aquifer in Cedar Valley used the combined pumping from two large irrigation wells on the western side of the valley during the summer of 2005. The need for irrigation water dictated the discharge from the wells, and because observation wells were also pumped private wells, the test conditions were less than ideal. Some of the 12 observation wells did not experience drawdown, either because the distance to the pumping well was too large or the wells were completed in the clay unit that confines the eastern half of the aquifer being tested. The cone of depression formed during this test and the aquifer parameters estimated from the test data show a confined aquifer that has moderately high transmissivity (25,000 ft^2/d [2300 m^2/d]) and moderately high storativity (0.02), and is elongated in the north-south direction. The edge of the confining unit and the depositional environment of the alluvial fan comprising the aquifer are the likely factors constraining the drawdown in the north-south direction. The flow at Fairfield Springs was not affected by the seasonal pumping in 2005.

The second test on the basin-fill aquifer was a single-well test on a medium-diameter well in the center of the valley. The leaky confined aquifer experienced 50 feet (15 m) of drawdown after 48 hours of pumping 215 gallons per minute (13.6 L/s). I factored antecedent water-level trend, leakage from irrigation, and wellbore storage into my analysis of drawdown. The aquifer, which is composed of silty clay at this location in the center of a closed basin, has transmissivity and hydraulic-conductivity values (1000 to 1200 ft^2/d [93–110 m^2/d] and 3.0 to 3.6 ft/d [0.9–1.1 m/d], respectively) in the moderate range for unconsolidated sediments.

The last basin-fill test I conducted lasted seven hours and involved a single 6-inch-diameter (15 cm) domestic well in Fairfield. The water level stabilized at approximately 49 feet (15 m) only five hours into the test while the well discharged 31 gallons per minute (2 L/s). Drawdown was affected by wellbore storage, partial penetration, and, probably, well inefficiency. The hydraulic conductivity of the confined leaky aquifer sediments is approximately 2 feet per day (0.6 m/d) and the transmissivity is approximately 70 feet squared per day (6.5 m^2/d) at this location.

Summary of the Oquirrh Group Fractured-Rock Aquifer Test

A five-month aquifer test on Eagle Mountain Municipal Supply Well 3 involving existing and new observation wells in ideal positions allowed for an excellent opportunity to determine the nature of the Pennsylvanian–Mississippian-aged carbonate and clastic rocks of the Oquirrh Group fractured-rock aquifer. Based on potentiometric evidence, flow out of the Cedar Valley basin-fill aquifer into the bedrock aquifer west of Cedar Pass is occurring, but is limited somehow. Monitor-well drilling revealed a wedge of low-transmissivity volcanic rocks positioned between the fractured bedrock and basin fill. The wedge and/or a fault on the western side of the Lake Mountains likely act as the barriers to groundwater flow. Another possible barrier may exist east (down gradient) of the well where the aquifer is truncated as the strata rise to the surface in the Lake Mountains syncline.

Complications in the test included a long-term (recovery?) water-level trend prior to the test, a variable flow rate in the first seven minutes of the test, not discovering an existing well that could be used as an observation well until after the start of the test, and malfunctioning equipment. Despite these issues, after I compensated for antecedent water-level trends, I collected a set of very good long-term drawdown and recovery data.

While pumping for five months at an average of 1930 gallons per minute (122 L/s), the maximum drawdowns in the pumping well and a well 539 feet (164 m) away from the pumping well were 41 feet (12.5 m) and 10.4 feet (3.2 m), respectively. The maximum drawdown in an observation well 4939 feet (1505 m) away, which aligned with the suspected high conductivity direction, was 8.65 feet (2.64 m), but another closer well located 4540 feet (1384 m) away and perpendicular to the suspected high conductivity direction showed only approximately 6 feet (2 m) of drawdown, indicating aquifer anisotropy. A fourth observation well showed no response because it was completed in the perched volcanic rock wedge above the sedimentary bedrock aquifer.

Aquifer-test analysis consisted of manual and automatic curve matching aided by computer software (Duffield, 2007) to double-porosity fractured-rock solutions and radial flow solutions for confined aquifers and unconfined aquifers with a correction to drawdown and accounting for delayed gravity yield. Concurrent drawdown derivative curve matching (Renard and others, 2009) confirmed that wellbore storage was affecting the drawdown in two wells and supported the existence of linear flow close to the pumping well and flow-limiting aquifer boundaries distant from the well. I used simple arithmetic methods (van Tonder and others, 2002; Heilweil and Hsieh, 2006) to confirm that the aquifer is anisotropic and determine that a flow-limiting aquifer boundary may be present approximately 4300 feet (1300 m) from the pumping well. Further experimentation with positions of a no-flow boundary using image well theory and curve matching provided aquifer parameters for a bounded aquifer, which likely are overestimates given that the boundaries likely are semi-permeable.

The fractured-rock aquifer in the Oquirrh Group (Butterfield Peaks Formation and West Canyon Limestone) has moderate to high transmissivity that is two to three times greater in the northwest to southeast direction parallel to the axis of structural folding (11,000–16,000 ft^2/d [1000–1500 m^2/d]) as compared to perpendicular to folding (6,000 ft^2/d [660 m^2/d]). The storativity of the aquifer is between 0.007 and 0.03. Linear flow near the pumping well and a wide drawdown cone characterize the aquifer. The extent of the aquifer is limited by a volcanic rock wedge and/or a normal fault on the western side of the Lake Mountains, which restrict the flow from the Cedar Valley basin-fill aquifer to the west. Basin-fill groundwater must either flow north or south around these obstacles or downward into bedrock underlying the basin fill.

Summary of the Great Blue Limestone Fractured-Rock Aquifer Test

A 46-day multiple-well aquifer drawdown and recovery test on Eagle Mountain Municipal Supply Well 2 produced an aquifer response indicative of an extensive fractured-rock aquifer. The Mississippian Great Blue Limestone aquifer is approximately 525 feet (160 m) thick at this location. No aquifer boundaries were encountered by the cone of depression while pumping for 35 days at 2530 gallons per minute (160 L/s), and maximum drawdown in an observation well 660 feet (201 m) away was 17.45 feet (5.32 m). I estimated aquifer parameters using automatic type-curve matching for a confined, double-porosity solution and compared them to the estimates using a confined isotropic and homogeneous solution. The hydraulic conductivity of the fracture system is approximately 24 feet per day (7.3 m/d), which results in a transmissivity estimate of approximately 12,600 feet squared per day (1170 m^2/d). The storativity of the aquifer is 7 x 10^{-3}. This analysis did not evaluate the possible anisotropy of the aquifer, but did show that the Great Blue Limestone aquifer is moderately transmissive, confined, and at least extensive enough in this area to be unaffected by potential aquifer boundaries for at least one-month pumping periods.

Summary of Estimated Aquifer Parameters

The best estimates of transmissivity, hydraulic conductivity, and storativity available to date in Cedar Valley, listed at the top of table 1, are derived from five UGS aquifer tests and either the original analysis or reanalysis of older aquifer tests on four other wells. While the transmissivity and hydraulic-conductivity values I derived from specific-capacity data (table 1) most likely are less accurate, their much greater number and more widespread distribution allowed me to generate statistics and maps of aquifer parameters for the entire Cedar Valley study area.

Table 7 provides a summary of the hydraulic conductivity, transmissivity, and storativity values derived in this study. The hydraulic conductivity and transmissivity of the basin-fill aquifer range over four and five orders of magnitude, from 2.6 x 10^{-3} to 6.9 x 10^{1} feet per day (7.9 x 10^{-4} to 2.1 x 10^{2} m/d) for K and 2.7 x 10^{-1} to 2.5 x 10^{4} feet squared per day (2.5 x 10^{-2} to 2.3 x 10^{3} m^2/d) for T. The geometric mean of the hydraulic conductivity in basin fill is 2.5 feet per day (0.8 m/d) and of transmissivity is 260 feet squared per day (24 m^2/d). Storativity is approximately 2 x 10^{-2} based on one multiple-well aquifer test. In the bedrock aquifer, the hydraulic conductivity and transmissivity range over three and five orders of magnitude, from 1.0 x 10^{-1} to 5.3 x 10^{2} feet per day (3 x 10^{-2} to 1.6 x 10^{2} m/d) for K and 4.0 to 1.2 x 10^{5} feet squared per day (0.4 to 1.1 x 10^{4} m^2/d) for T. The geometric mean of the hydraulic conductivity in bedrock is 3.4 feet per day (1.0 m/d) and of transmissivity

Table 7. *Aquifer parameter estimates for Cedar Valley aquifers.*

Aquifer		Transmissivity (ft^2/d)	Hydraulic conductivity (ft/d)	Storativity (unitless)
Basin-fill[1]	Range	0.27 to 25,000	2.6×10^{-3} to 69	0.02
	Geometric Mean	260	2.5	
Bedrock[2]	Range	4 to 120,000	0.1 to 530	0.007 to 0.02
	Geometric Mean	360	3.4	

[1] Basin-fill aquifer estimates are compiled from three aquifer tests and 45 specific capacity values. Storativity value is from one multiple-well aquifer test.

[2] Bedrock aquifer estimates are calculated from six aquifer tests and 22 specific capacity values. Storativity values are from two multiple-well aquifer tests.

is 360 feet squared per day (33 m^2/d). The storativity of the bedrock aquifer is likely between 7×10^{-3} and 2×10^{-2} based on two multiple-well aquifer tests. The thickness of the aquifers, which I determined using well logs, averaged 180 feet (55 m) thick, but ranged from 5 to 1044 feet (1.5–318 m) depending on the well construction and location of the well.

Hydraulic conductivity and transmissivity are typically found over such large ranges in any given area that it is most convenient to discuss their variability in terms of the log of individual values, as shown for basin-fill and bedrock wells in Cedar Valley on figures 27 and 28. Although the absolute range of hydraulic conductivity in the basin fill is larger than in the bedrock, examination of the log figures tells us more about the distribution of hydraulic conductivity in the study area. The ranges of all but one extremely small value of hydraulic conductivity and transmissivity calculated from basin-fill wells is narrower than the range of values for the bedrock aquifer, and the values are generally more evenly distributed throughout that range (figures 27 and 28). By comparison, there are more bedrock values on the low end of the range than the high end. The difference in the distribution of values likely is because the basin-fill aquifer sediments fall evenly on the spectrum between fine grained and coarse grained, whereas wells in bedrock will always be completed in consolidated sedimentary rocks, which typically have lower permeability than unconsolidated sediments, but occasionally will be completed in an area having intersecting fractures that boost the permeability by orders of magnitude.

I defined zones of similar hydraulic conductivity (figure 29) by plotting all the hydraulic-conductivity values and computer-generated contours of those values on maps of the basin-fill and bedrock aquifers. I refined the computer-generated contours using knowledge of the local structural geology and aquifer lithology to smooth and simplify the values into three hydraulic-conductivity zones for the basin-fill aquifer and four hydraulic-conductivity zones for the bedrock aquifer. The highest hydraulic-conductivity zone (200–700 ft/d [61–210 m/d]) is in the bedrock aquifer on the eastern side of the valley near two exceptionally productive wells (IDs 1018 and 56, figure 1). High hydraulic-conductivity zones in the basin-fill aquifer (20–50 ft/d [6–15 m/d]) coincide with coarser alluvial fan sediments deposited along the western and eastern margins of the basin-fill aquifer. The lowest hydraulic-conductivity zone in the basin-fill aquifer (1×10^{-6}–4.9 ft/d [3×10^{-7}–1.5 m/d]) is located in the southern arm of the valley and extends north through the center of the valley. I assigned the lowest hydraulic-conductivity zone in the bedrock aquifer (1×10^{-6}–24.9 ft/d [3×10^{-7}–8.6 m/d]) to areas where there was no evidence for higher hydraulic conductivity.

This study provides new field data from the two most important aquifers in the Cedar Valley study area, the principal basin-fill aquifer and the fractured-bedrock aquifer. I used all available historical aquifer test and well log information in combination with the new data to formulate the most comprehensive set of hydraulic properties and aquifer characteristics to date for this area. The most significant new findings of this study are details about the interface between the basin-fill aquifer west of Cedar Pass and the fractured-bedrock aquifer underlying Cedar Pass, an area of interest because of its subsurface groundwater discharge from the Cedar Valley groundwater basin, the increased development of its bedrock groundwater resource over the past decade, and its proximity to a water-right administration boundary. Using well log and potentiometric data, and by analyzing the bedrock aquifer's response to pumping near this interface, I determined that one or more flow-limiting boundaries hinders flow from the basin fill to the bedrock. Volcanic bedrock, a buried fault, a thinning aquifer, or some combination of these geologic elements create these boundaries.

Jordan and Sabbah (2012) used the aquifer properties and our new understanding of groundwater flow out of the Cedar Valley groundwater basin developed in this study in a comprehensive groundwater resources investigation and groundwater modeling study of Cedar Valley. The goal of this aquifer parameter estimation study and the groundwater modeling work (Jordan and Sabbah, 2012) is to provide good scientific information to groundwater policy makers and those seeking to use, develop, and protect groundwater resources in Utah County.

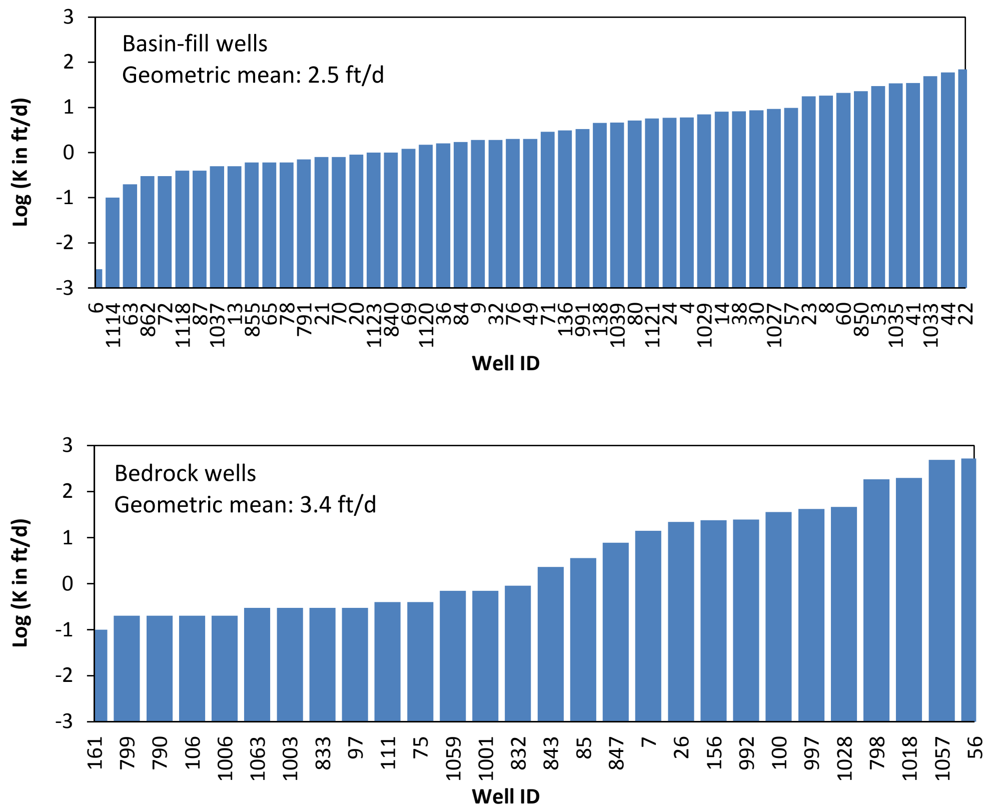


Figure 27. Distribution of the log of hydraulic-conductivity (K) values calculated from aquifer tests and estimated from specific-capacity data for basin-fill and bedrock wells in Cedar Valley. See figure 1 for well locations.

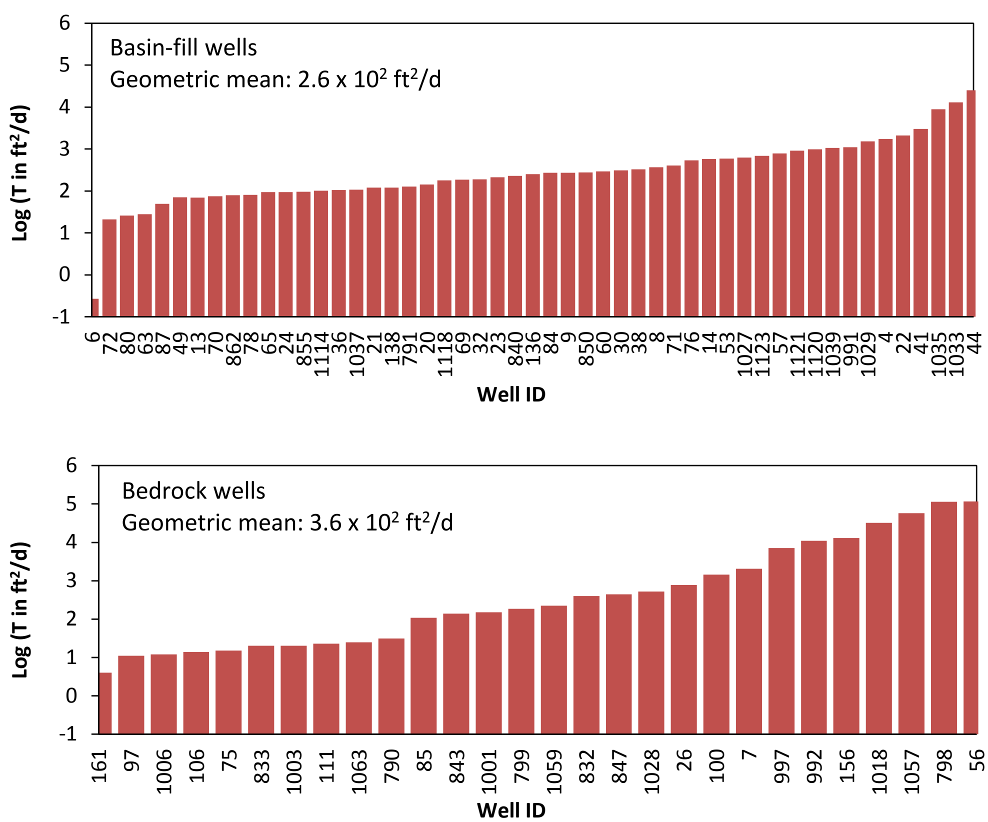


Figure 28. Distribution of the log of transmissivity (T) values calculated from aquifer tests and estimated from specific-capacity data for basin-fill and bedrock wells in Cedar Valley. See figure 1 for well locations.

ACKNOWLEDGMENTS

I thank my fellow geologists in the UGS Groundwater and Paleontology Program for their dedication to collecting good quality data during the aquifer tests. They worked long days, evenings, and weekends to measure water-level data when pressure transducers were unavailable. Special thanks to the well owners and operators, Aaron Allred, Michael Burch, Michael Christensen, Keith Jonsson, Dave Norman, Jim and Grant Smith, Mark Sovine, Christopher Trusty, and David White, for allowing me access to their wells and historical data, and for accepting the terms of each aquifer test despite increased cost and inconvenience in their daily business. Paul Inkenbrandt, UGS, helped me to understand some of the atypical well responses, and his review of the report proved exceptionally helpful. Thomas Lachmar, Utah State University, provided an excellent technical review. Additional reviews by Walid Sabbah, Mike Lowe, Robert Ressetar, Kimm Harty, and Rick Allis, UGS, improved the manuscript.

This work was funded in part by the Utah Division of Water Rights and Eagle Mountain, and I am grateful for their support.

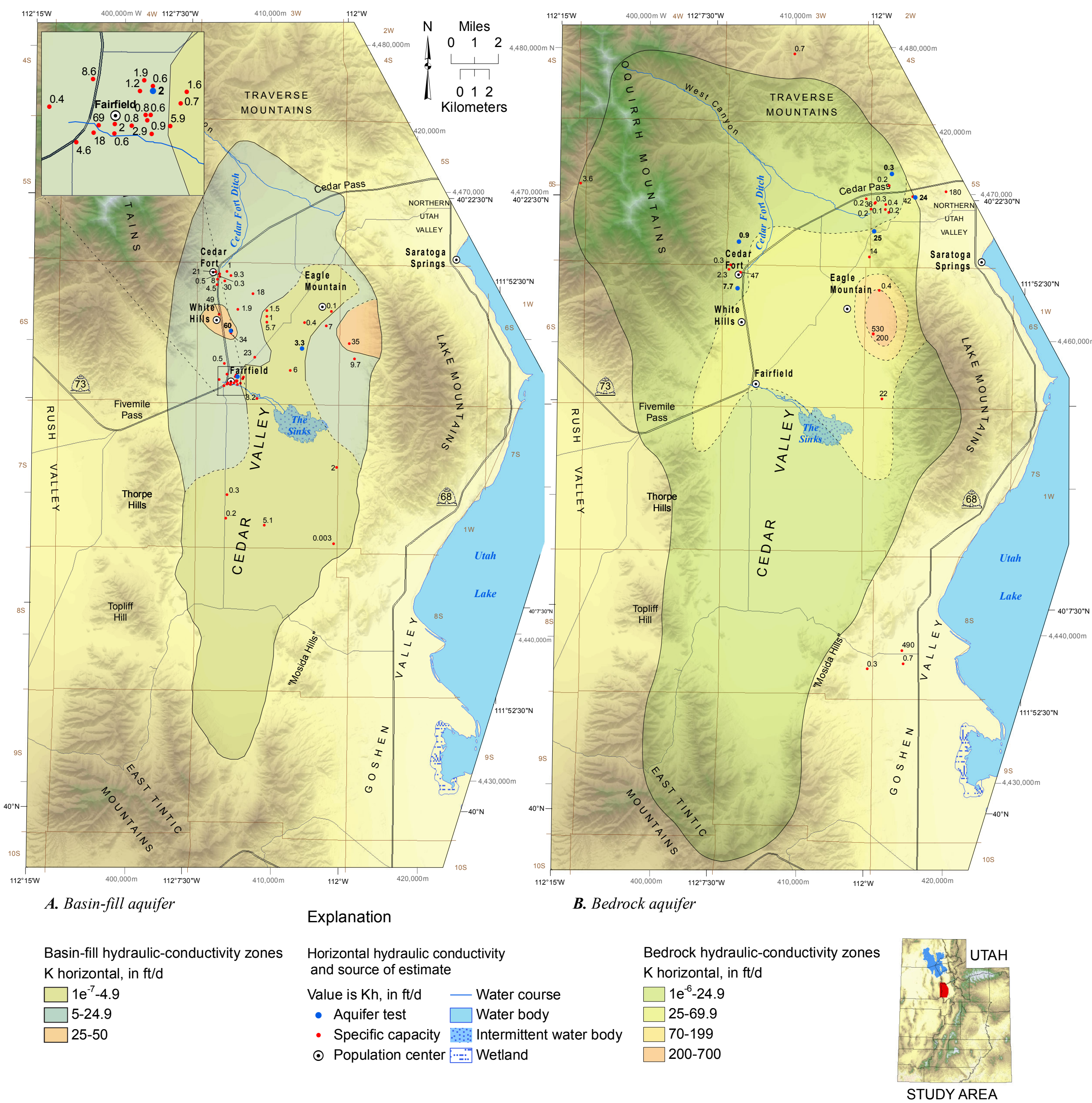


Figure 29. *Location of hydraulic-conductivity estimates derived by aquifer tests and specific-capacity data and hydraulic-conductivity zones for the basin-fill (A) and bedrock aquifers (B).*

REFERENCES

Allen, D.M., and Michel, F.A., 1998, Evaluation of multi-well test data in a faulted aquifer using linear and radial flow models: Ground Water, v. 36, no. 6, p. 938–948 (doi: 10.1111/j.1745-6584.1998.tb02100.x).

Allmendinger, R.W., 1992, Fold and thrust tectonics of the western United States exclusive of accreted terranes, *in* Burchfiel, B.C., Lipman, P.W., and Zoback, M.L., editors, The Cordilleran orogeny—conterminous U.S.: Geological Society of America, The Geology of North America, v. G-3, p. 583–607.

Anderson, M.P., and Woessner, W.W., 1992, Applied groundwater modeling—simulation of flow and advective transport: San Diego, California, Academic Press, 381 p.

Bense, V.F., and Person, M.A., 2006, Faults as conduit-barrier systems to fluid flow in siliciclastic sedimentary aquifers: Water Resources Research, v. 42, W05421 (doi: 10.1029/2005WR004480).

Biek, R.F., 2004, Geologic map of the Saratoga Springs 7.5' quadrangle, Utah County, Utah, and Geologic Map of the Cedar Fort 7.5' quadrangle, Utah County, Utah: Utah Geological Survey Map 201 and 202, 3 plates, scale 1:24,000.

Biek, R.F., 2005, Geologic map of the Jordan Narrows quadrangle, Salt Lake and Utah Counties, Utah: Utah Geological Survey Map 208, 2 plates, scale 1:24,000.

Biek, R.F., Solomon, B.J., Keith, J.D., and Smith, T.W., 2005, Geologic map of the Tickville Spring quadrangle, Salt Lake and Utah Counties, Utah: Utah Geological Survey Map 214, 2 plates, scale 1:24,000.

Birsoy, Y.K., and Summers, W.Y., 1980, Determination of aquifer parameters from step tests and intermittent pumping: Ground Water, v. 18, no. 2, p. 137–146.

Boulton, N.S., 1963, Analysis of data from non-equilibrium pumping tests allowing for delayed yield from storage: Proceedings of the Institute of Civil Engineers, v. 26, p. 469–482.

Bourdet, D., and Gringarten, A.C., 1980, Determination of fissure volume and block size in fractured reservoirs by type-curve analysis: *in* The 1980 Society for Petroleum Engineers Annual Fall Technical Conference and Exhibit, Dallas, Texas, September 21–24, 1980, Paper SPE 9293.

Bradbury, K.B., and Rothschild, E.R., 1985, A computerized technique for estimating the hydraulic conductivity of aquifers from specific capacity data: Ground Water, v. 23, no. 2, p. 240–246.

Bryant, B.B., and Nichols, D.J., 1988, Late Mesozoic and early Tertiary reactivation of an ancient crustal boundary along the Uinta trend and its interaction with the Sevier orogenic belt, *in* Schmidt, C.J., and Perry, W.J., editors, Interaction of the Rocky Mountain foreland and the Cordilleran thrust belt: Geological Society of America Memoir 171, p. 411–430.

Butler, J.J., Jr., Jin, W., Mohammed, G., and Reboulet, E., 2011, New insights from well responses to fluctuations in barometric pressure: Ground Water, v. 49, no. 4, p. 525–533 (doi: 10.1111/j.1745-6584.2010.00768.x).

Cobb, M., 2005, TGuess05 a Microsoft Excel spreadsheet for estimating transmissivity and hydraulic conductivity from specific capacity test data based on Bradbury and Rothschild, 1985: Madison, University of Wisconsin-Madison Department of Geology and Geophysics, version 1.1.

Cook, P.G., 2003, A guide to regional groundwater flow in fractured rock aquifers: Glen Osmond, SA, Australia, CSIRO Land and Water, 108 p.

Cooper, H.H., and Jacob, C.E., 1946, A generalized graphical method for evaluating formation constants and summarizing well field history: American Geophysical Union Transactions, v. 27, p. 526–534.

Domenico, P.A., and Schwartz, W.W., 1990, Physical and chemical hydrogeology: New York, John Wiley & Sons, 824 p.

Dougherty, D., and Babu, D., 1984, Flow to a partially penetrating well in a double-porosity reservoir: Water Resources Research, v. 20, no. 8, p. 1116–1122.

Driscoll, F.G., 1986, Groundwater and wells (2nd ed.): St. Paul, Minnesota, Johnson Filtration Systems, 1108 p.

Duffield, G.M., 2003, AQTESOLV for Windows aquifer test analysis software: Reston, Virginia, HydroSOLVE, Inc., version 3.5.

Duffield, G.M., 2007, AQTESOLV for Windows aquifer test analysis software, professional edition: Reston, Virginia, HydroSOLVE, Inc., version 4.5.

ESI Engineering, Inc., 2003, Drinking water source protection plan, Community Center Well: Salt Lake City, unpublished consultant's report submitted to Utah Department of Environmental Quality, Division of Drinking Water for the town of Cedar Fort, 13 p., 5 appendices.

Feltis, R., 1967, Ground-water conditions in Cedar Valley, Utah County, Utah: Utah State Engineer Technical Publication 16, 31 p.

Ferre, T.P.A., and Thomasson, M.J., 2010, Understanding the impacts of anisotropy on the extent of drawdown: Ground Water, v. 48, no. 4, p. 478–479 (doi: 10.1111/j.1745-6584.2009.00656.x).

Ferris, J.G., Knowles, D.B., Brown, R.H., and Stallman, R.W., 1962, Theory of aquifer tests: U.S. Geological Survey Water-Supply Paper 1536-E, 174 p.

Fetter, C.W., 1988, Applied Hydrogeology: New York, Macmillan, 592 p.

Freeze, A.R., and Cherry, J.A., 1979, Groundwater: Englewood Cliffs, New Jersey, Prentice Hall, 604 p.

Gernand, J.D., and Heidtman, J., 1997, Detailed pumping test to characterize a fractured bedrock aquifer: Ground Water, v. 35, no. 4, p. 632–637.

Gringarten, A.C., Ramey, H.J., Jr., and Raghavan, R., 1974, Unsteady-state pressure distributions created by a well with single infinite-conductivity vertical fracture: Society of Petroleum Engineers Journal, v. 14, no. 4, p. 347–360 (doi: 10.2118/4051-PA).

Gringarten, A.C., and Witherspoon, P., 1972, A method of analyzing pump test data from fractured rock aquifers, *in* Proceedings of the Symposium on Percolation Through Fissured Rock, Stuttgart, Germany, International Association of Rock Mechanics, p. T3-B1-T3-B8.

Halford, K.J., 2006, Documentation of a spreadsheet for time-series analysis and drawdown estimation: U.S. Geological Survey Scientific Investigations Report 2006-5024, 38 p.

Hantush, M.S., 1961, Aquifer test on partially penetrating wells: Proceedings of the American Society of Civil Engineers, v. 87, p. 171–195.

Hantush, M.S., 1964, Hydraulics of wells, *in* Chow, V.T., editor, Advances in hydroscience: New York Academic Press, p. 281–442.

Hantush, M.S., and Jacob, C.E., 1955, Non-steady radial flow in an infinite leaky aquifer: American Geophysical Union Transactions, v. 36, p. 95–100.

Hare, P.W., and Morse, R.E., 1997, Water-level fluctuations due to barometric pressure changes in an isolated portion of an unconfined aquifer: Ground Water, v. 35, no. 4, p. 667–671 (doi: 10.1111/j.1745-6584.1997.tb00132.x).

Heilweil, V.M., and Hsieh, P.A., 2006, Determining anisotropic transmissivity using a simplified Papadopulos method: Ground Water, v. 44, no. 5, p. 749–753 (doi: 10.1111/j.1745-6584.2006.00210.x).

Hintze, L.F., and Kowallis, B.J., 2009, Geologic history of Utah: Brigham Young University geology studies special publication 9: Provo, Utah, Department of Geological Sciences, Brigham Young University, 225 p.

Huntley, D., Nommensen, R., and Steffey, D., 1992, The use of specific capacity to assess transmissivity in fractured-rock aquifers: Ground Water, v. 30, no. 3, p. 396–402 (doi: 10.1111/j.1745-6584.1992.tb02008.x).

Hurlow, H.A., 2004, The geology of Cedar Valley, Utah County, Utah, and its relation to ground-water conditions: Utah Geological Survey Special Study 109, 36 p., 4 plates.

Jacob, C.E., 1940, On the flow of water in an elastic artesian aquifer: Transactions of the American Geophysical Union, v. 21, p. 574–586.

Jenkins, D.N., and Prentice, J.K., 1982, Theory for aquifer test analysis in fractured rocks under linear (nonradial) flow conditions: Ground Water, v. 20, no. 1, p. 12–22 (doi: 10.1111/j.1745-6584.1982.tb01325.x).

Jordan, J.L., 2008, Report of 2007 monitoring well drilling for Eagle Mountain city: Salt Lake City, contract deliverable report for Eagle Mountain, Utah Geological Survey, 72 p.

Jordan, J.L., 2009, 2007 Well 3 aquifer testing, Eagle Mountain city, Utah: Salt Lake City, contract deliverable report for Eagle Mountain, Utah Geological Survey, 55 p., appendices variously paginated.

Jordan, J.L., and Sabbah, W.W., 2012, Hydrogeology and simulation of groundwater flow in Cedar Valley, Utah County, Utah: Utah Geological Survey Special Study 145, 221 p., 2 plates.

Jordan, J.L., and Sabbah, W., 2007, Ground-water flow, water-level trends, and the connection between Fairfield Spring and the basin-fill aquifer in Cedar Valley, Utah County, north-central Utah, *in* Willis, G.C., Hylland, M.D., Clark, D.L., and Chidsey, T.C., Jr., editors, Central Utah—diverse geology of a dynamic landscape: Utah Geological Association Publication 36, p. 345–359.

Kasenow, M., 2006, Aquifer test data—analysis and evaluation: Highlands Ranch, Colorado, Water Resources Publications, LLC, 348 p.

Kruseman, G.P., and de Ridder, N.A., 2000, Analysis and evaluation of pumping test data (2nd ed.): Wageningen, The Netherlands, International Institute for Land Reclamation and Improvement, 377 p.

Lachmar, T.D., Bradbury, K.B., and Evans, J.P., 2002, Structure and hydrogeology of deformed sedimentary bedrock aquifers, western Summit County, Utah: Environmental & Engineering Geoscience, v. 8, no. 3, p. 219–236.

Mace, R.E., 2001, Estimating transmissivity using specific-capacity data: Texas Bureau of Economic Geology Geological Circular 01-2, 44 p.

Moench, A.F., 1984, Double-porosity models for a fissured groundwater reservoir with fracture skin: Water Resources Research, v. 20, no. 7, p. 831–846.

Moench, A.F., 1985, Transient flow to a large-diameter well in an aquifer with storative semiconfining layers: Water Resources Research, v. 21, no. 8, p. 1121–1131.

Moench, A.F., 1988, The response of partially penetrating wells to pumpage from double-porosity aquifers, *in* Proceedings of the International Conference on Fluid Flow in Fractured Rocks, Atlanta, Georgia, May 16–18.

Montgomery, S.B., 1995, Hydrogeologic study, John Walden well, water source protection report with delineation protection zones and management areas: Bountiful, Utah, unpublished consultant's report submitted to Sowby & Berg Consultants, American Fork, Utah, and Utah Department of Environmental Quality, Division of Drinking Water, 15 p., 1 appendix.

Montgomery Watson, Inc., 2000, Well report for Eagle Mountain city exploration well no. 3: Salt Lake City, unpublished consultant's report for Eagle Mountain, variously paginated.

Montgomery Watson Harza, 2001, Town of Eagle Mountain exploration well no. 3 drinking water source protection plan: Salt Lake City, unpublished consultant's report for Eagle Mountain, variously paginated.

Moore, W.J., 1973, Preliminary geologic map of western Traverse Mountains and northern Lake Mountains, Salt Lake and Utah Counties, Utah: U.S. Geological Survey Miscellaneous Field Studies Map MF-490.

Mower, R.W., and Allred, C., 1997, Drinking water source protection plan for a new well system no. 25148, source no. 01: Fairview, Utah, unpublished consultant's report submitted to Utah Department of Environmental Quality, Division of Drinking Water for Meadow Ranch, LC., 28 p.

Neuman, S.P., 1972, Theory of flow in unconfined aquifers considering delayed response of the water table: Water Resources Research, v. 8, no. 4, p. 1031–1043.

Neuman, S.P., 1974, Effect of partial penetration on flow in unconfined aquifers considering delayed gravity response: Water Resources Research, v. 10, no. 2, p. 303–312.

Papadopulos, I.S., 1965, Nonsteady radial flow to a well in an infinite anisotropic aquifer, *in* Proceedings of the Dubrovnik Symposium on the Hydrology of Fractured Rocks: Dubrovnik, Yugoslavia, International Association of Scientific Hydrology, p. 21–31.

Papadopulos, I.S., and Cooper, H.H., 1967, Drawdown in a well of large diameter: Water Resources Research, v. 3, no. 1, p. 241–244.

Rasmussen, T.C., and Crawford, L.A., 1997, Identifying and removing barometric pressure effects in confined and unconfined aquifers: Ground Water, v. 35, no. 3, p. 502–511 (doi: 10.1111/j.1745-6584.1997.tb00111.x).

Renard, P., 2005, Hydraulics of wells and well testing, *in* Anderson, M.G., editor, Encyclopedia of hydrological sciences: Hoboken, New Jersey, John Wiley & Sons, v. 4, p. 2323–2340.

Renard, P., Glenz, D., and Mejias, M., 2009, Understanding diagnostic plots for well-test interpretation: Hydrogeology Journal, v. 17, no. 3, p. 589–600 (doi: 10.1007/s10040-008-0392-0).

Rorabaugh, M.J., 1953, Graphical and theoretical analysis of step-drawdown test of artesian well: Proceedings of the American Society of Civil Engineers, v. 79, no. 362-23.

Singhal, B.B.S., and Gupta, R.P., 2010, Applied hydrogeology of fractured rocks (2 ed.): Dordrecht, The Netherlands, Springer, 408 p.

Smith, E.D., and Vaughan, N.D., 1985, Aquifer test analysis in nonradial flow regimes—a case study: Ground Water, v. 23, no. 2, p. 167–175 (doi: 10.1111/j.1745-6584.1985.tb02789.x).

Spane, F.A., 2002, Considering barometric pressure in groundwater flow investigations: Water Resources Research, v. 38, no. 6, 1078 (doi: 10.1029/2001WR000701).

Streltsova, T.D., 1988, Well testing in heterogeneous formations: New York, John Wiley and Sons, Inc., 413 p.

Theis, C.V., 1935, The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using groundwater storage: American Geophysical Union Transactions, v. 16, p. 519–524.

Theis, C.V., Brown, R.H., and Meyer, R.R., 1963, Estimating the transmissibility of aquifers from the specific capacity of wells: U.S. Geological Survey Water-Supply Paper 1536-I, 10 p.

Toll, N.J., and Rasmussen, T.C., 2007, Removal of barometric pressure effects and earth tides from observed water levels: Ground Water, v. 45, no. 1, p. 101–105.

University of Utah Department of Atmospheric Sciences, 2008, MesoWest weather observations data: Online, http://mesowest.utah.edu/cgi-bin/droman/mesomap.cgi?state=UT&rawsflag=3, accessed 4/05/2008.

Utah Division of Water Resources, 2010, 2009 Residential Water Use: Utah Department of Natural Resources, 37 p.

Utah Division of Water Rights, 2007, Water right information for water right 0754006M00: Online, http://waterrights.utah.gov/cgi-bin/docview.exe?Folder=welllog430695, accessed August 19, 2009.

Utah Division of Water Rights, 2009a, Utah water use program records: Online, http://waterrights.utah.gov/distinfo/wuse.asp, accessed March 3, 2010.

Utah Division of Water Rights, 2009b, Public water supplier information for White Hills Subdivision Water Company in the Utah water use program records: Online, http://www.waterrights.utah.gov/cgi-bin/wuseview.exe?Modinfo=Pwsview&SYSTEM_ID=1446, accessed September 24, 2011.

van Tonder, G., Bardenhagen, I., Riemann, K., van Bosch, J., Dzango, P., and Xu, Y., 2002, Manual on pumping test analysis in fractured-rock aquifers: Bloemfontein, South Africa, Institute for Groundwater Studies, University of the Free State, 228 p.

APPENDICES

APPENDIX A

WATER-LEVEL AND DISCHARGE DATA COLLECTED FOR AQUIFER TESTS

on CD: Appendix A Tables A-1 - A-7 wls and Q.xlsx

Table A-1. *Water-level data collected for the White wells aquifer test.*

Table A-2. *Water-level data collected for the medium-diameter basin-fill aquifer test.*

Table A-3. *Water-level data collected for the small-diameter basin-fill aquifer test.*

Table A-4. *Water-level data collected for the Well 3 fractured rock aquifer test.*

Table A-5. *Average discharge of Eagle Mountain Well 3 during 2007 aquifer test.*

Table A-6. *Average discharge of Eagle Mountain Well 2 during 2007 aquifer test.*

Table A-7. *Water-level data collected for the Well 2 fractured rock aquifer test.*

APPENDIX B

WELL DRILLERS' LOGS, GEOPHYSICAL LOGS, AND LITHOLOG LOGS

Figure B-1. *Cadastral well numbering system.*

on CD: B-1 NumberingSystem.pdf

Figure B-2. *Driller's log for well UGS ID 44.*

on CD: B-2 pg 1 welllog44pg1.tif (page 1)

on CD: B-2 pg 2 welllog44pg2.tif (page 2)

Figure B-3. *Page 1 of driller's log for well UGS ID 1035.*

on CD: B-3 welllog1035.pdf

Figure B-4. *Driller's and geophysical logs for well UGS ID 991.*

on CD: B-4 pg 1 welllog991pg1.tif (page 1)

on CD: B-4 pg 2 welllog991pg2.tif (page 2)

on CD: B-4 pg 3 991win29853.pdf (page 3)

Figure B-5. *Driller's log for well UGS ID 49.*

on CD: B-5 welllog49.tif

Figure B-6. *Driller's and lithologic logs for Eagle Mountain Well 3 (UGS ID 992).*

on CD: B-6 well3.pdf

Figure B-7. *Driller's and lithologic logs for Eagle Mountain Well 2 (UGS ID 156).*

on CD: B-7 well2.pdf

APPENDIX C

TRANSMISSIVITY ANISOTROPY ANALYSIS OF WELL 3 AQUIFER-TEST DATA

Appendix C. Transmissivity anisotropy analysis of Well 3 aquifer-test data using a simplified version of the Papodopulos (1965) method by Heilweil and Hsieh (2006).

Pumping well	Well 3
Observation well A (in x direction)	MW2b
Observation well B (in y direction)	807

Input from test data is shown in yellow shading.

Step*	Parameter	Value	Unit	Formula	Note
1	Q	1930	gpm		Well discharge
	Δs	7.60	feet		Change in drawdown over one log cycle time, read from graph
	$\sqrt{T_{xx}T_{yy}}$	8963	ft^2/d	$\frac{(264)(Q)}{(7.48)(\Delta s)}$	Substitution of $T_{xx}T_{yy}$ into Cooper-Jacob straight line equation ($\sqrt{T}=2.3Q/(4\pi\Delta s)$) including conversion factors for English units
	$T_{xx}T_{yy}$	8.03E+07			Square $\sqrt{T_{xx}T_{yy}}$ to obtain $T_{xx}T_{yy}$
2	r_a	4940	feet		Distance from pumping well to obs well A
	r_b	4350	feet		Distance from pumping well to obs well B
	t_{0a}	16,000	minutes		Time at x-intercept for obs well A
	t_{0b}	32,000	minutes		Time at x-intercept for obs well B
	S/T_{xx}	1.02E-06		$\frac{(2.25)(t_{0a})}{(r_a)^2}$	
	S/T_{yy}	2.64E-06		$\frac{(2.25)(t_{0b})}{(r_b)^2}$	
	$S^2/(T_{xx}T_{yy})$	2.71E-12		$(S/T_{xx})\cdot(S/T_{yy})$	
3	S	***1.47E-02***	none	$\sqrt{(T_{xx}T_{yy}\cdot(S^2/T_{xx}T_{yy}))}$	Storativity
4	T_{xx}	***14,395***	ft^2/d	$S/(S/T_{xx})$	Transmissivity in the x direction
5	T_{yy}	***5581***	ft^2/d	$S/(S/T_{yy})$	Transmissivity in the y direction

Results:

Transmissivity in the NW-SE direction aligned with axis of structural deformation: **14,000 ft^2/d**

Transmissivity in the NE-SW direction perpendicular to axis of structural deformation: **6000 ft^2/d**

Hydraulic conductivity in the NW-SE direction: **26 ft/d**

Hydraulic conductvity in the NE-SW direction: **11 ft/d**

Specific storage of aquifer (S_s=S/b) where b = 530 ft: **2.8E-05 ft^{-1}**

Ratio of anisotropy: **2.3:1**

* Steps correspond to steps described in Heilweil and Hsieh, 2006.

Heilweil, V.M., and Hsieh, P.A., 2006, Determining anisotropic transmissivity using a simplified Papadopulos method: Ground Water, v. 44, no. 5, p. 749-753 (doi: 10.1111/j.1745-6584.2006.00210.x).

Papadopulos, I.S., 1965, Nonsteady radial flow to a well in an infinite anisotropic aquifer, *in* Proceedings of the Dubrovnik Symposium on the Hydrology of Fractured Rocks: Dubrovnik, Yugoslavia, International Association of Scientific Hydrology, p. 21-31.