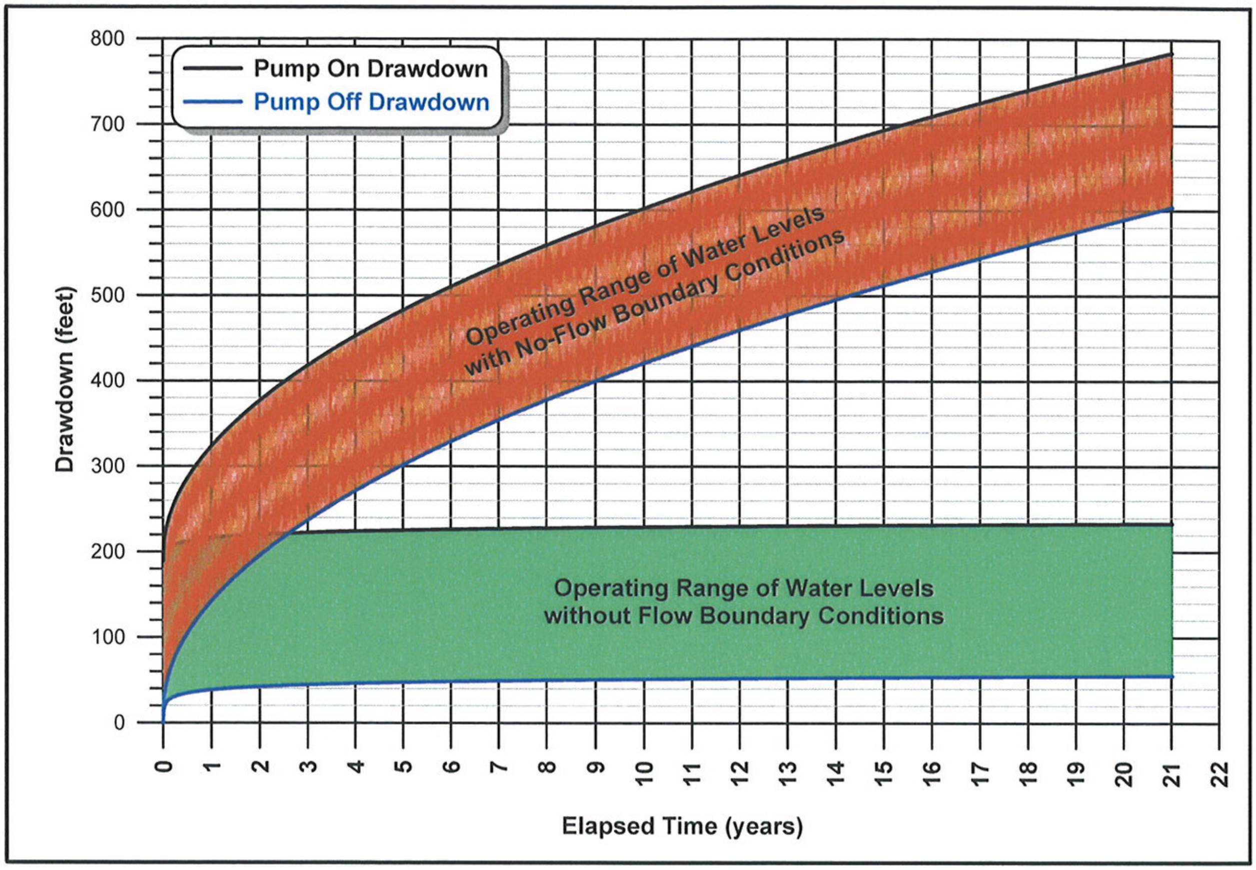
Figure 5-25
20-Year Projected Drawdown at ASR Test Well With and Without Boundaries and 260-Gpm Cyclical Pumping Rate



The difference between simulated drawdown with no boundary conditions and simulated drawdown with boundary conditions is remarkable, as demonstrated on Figure 5-25. The projected drawdown with no negative boundary conditions indicates groundwater development can be sustained for as much as 100 years, including interference between properly spaced wells (not shown), as previously demonstrated by the well field model with 15 wells and similar hydraulic parameters shown on Figure 5-15. This calculation is the type of projection made for the Sleepy Hollow No. 6 well wherein yield is projected for a 100-year period without taking into account the uncertainty of the hydraulic parameter determination or the limits imposed by negative boundary effects evident in the test response.

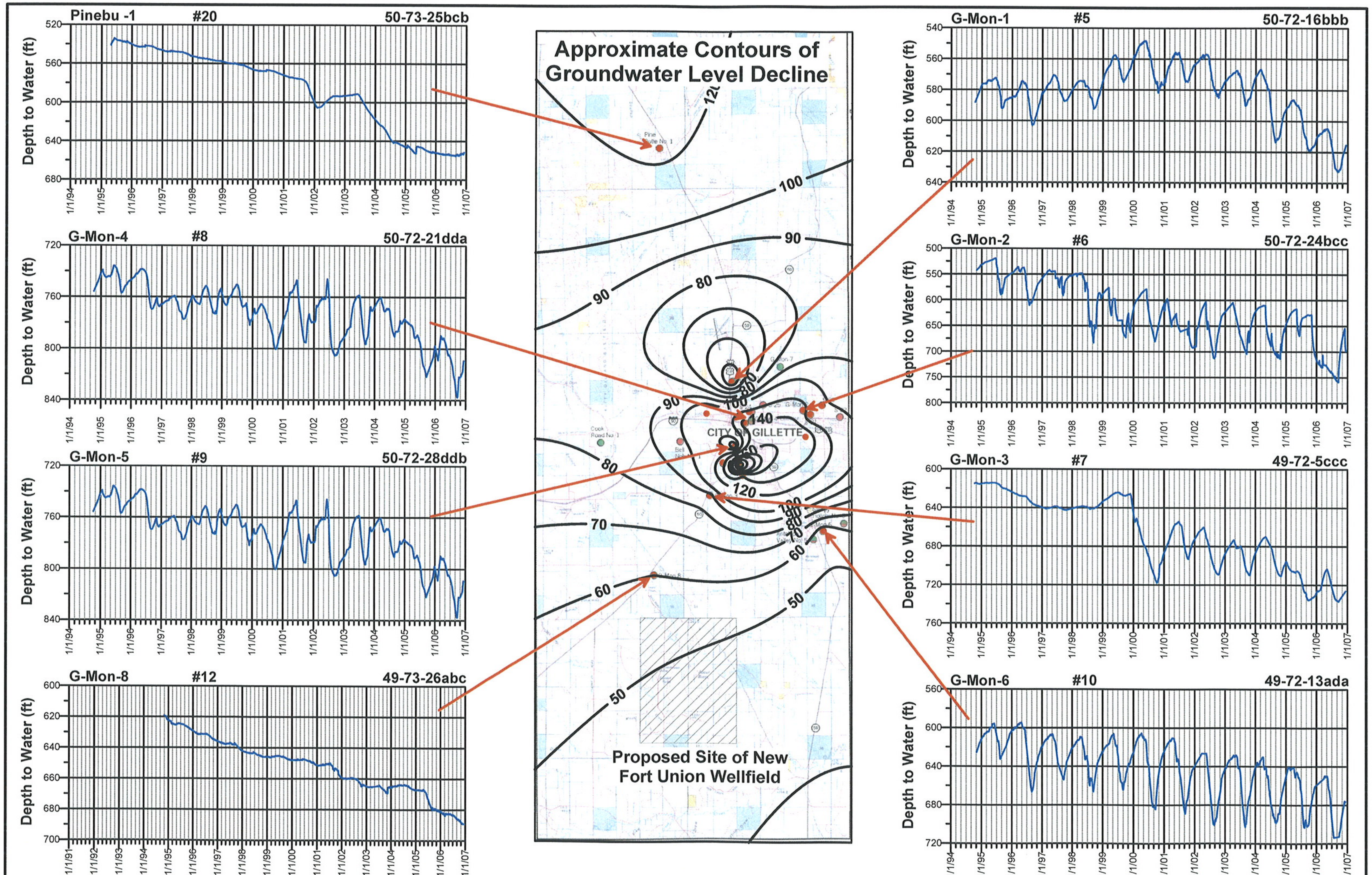
The projection of long-term sustainable groundwater development disappears when the effects of negative boundary conditions are taken into account at a single well, not considering interference from other wells. The projected drawdown at the ASR site after 20 years of cyclical pumping 260 gpm is approximately 770 feet (Figure 5-25) as compared to the available water column of 850 feet. A dedicated monitoring well located one mile north of the proposed Southern Well Field area exhibits a relatively consistent rate of groundwater decline of 5 feet per year in the 12 years ending in 2005. If a 5-ft/yr decline is projected over the 20-year model simulation, an additional 100 feet of groundwater level decline can be added to the 770 feet shown on Figure 5-14 for a total of 870 feet, 20 feet more than the available water column, not taking into account interference from other production wells in the well field.

Therefore, the projection of drawdown in a single well, pumped at 260 gpm for 12 hours each day for 20 years and subject to the negative boundary effects revealed by the ASR site aquifer test, indicates that a 20-year sustainable groundwater development using wells at any reasonable spacing is not possible at 260-gpm rates, 130-gpm average rate, for a 20 year period if the present rate of groundwater level decline in the area continues. This conclusion does not take into consideration an additional uncertainty which is the possibility that over a 20-year well field life, additional negative boundary conditions may ensue due to the geometry of the discontinuous lenses of sandstone in the Fort Union Formation and/or additional interference drawdown due to coal bed methane (CBM) pumping.

5.6.1.3 Uncertainty Due to On-Going Groundwater Level Declines

Groundwater levels, both static and pumping, have been measured intermittently in the City of Gillette production wells completed in the Fort Union Formation since the early 1980's in some wells and since the late 1990's in other more recently completed wells. In addition, the Wyoming State Engineer's Office has operated a number of dedicated monitoring wells completed in the Fort Union Formation since late 1994 and early 1995. Figure 5-26 shows generalized long-term groundwater level declines in the Fort Union

Figure 5-26
Generalized Groundwater Level Declines in the Gillette Area



aquifer in the vicinity of the City of Gillette, based on the foregoing records. The calculated groundwater declines are not normalized to any particular starting date and therefore are generalized with respect to time. However, they serve to indicate the total magnitude of groundwater level declines in the area to the extent defined by the historic observations. The groundwater level declines on Figure 5-26 also assume that the groundwater level declines observed at each location are extensive enough to be coextensive with those at other locations and are not limited to relatively small local areas around each well. The latter assumption is reasonable for the area immediately around Gillette and extending to the south, where there are quite a few wells observed; however, the Pine Buttes No. 1 observations to the north are likely more localized than suggested by the groundwater level decline contours on Figure 5-26.

The historic and on-going decline of groundwater levels in the area around Gillette began considerably before any coal bed methane pumping of groundwater and appears to be driven by use of the City of Gillette wells and other public water supply wells in the surrounding communities. Therefore, the ongoing decline of the Fort Union aquifer water levels must be projected into the future as another factor that will affect the sustainability of groundwater development from this aquifer system.

It is evident from the individual hydrographs provided on Figure 5-26 that changes in pumping schedules at some of the City wells, in order to improve water quality and meet drinking water standards, has had significant effect on the local groundwater levels. When some City wells have been put out of service to avoid production of water requiring treatment for fluoride or other constituents, local groundwater levels have recovered for a few years. Therefore, projection of the historic and current rates of groundwater decline into the future involves considerable uncertainty associated with potential changes in operation of the various public water supply wells in the area as well as the number of such wells in the future as the local population continues to grow.

It has already been demonstrated that the current rate of groundwater level decline in the proposed Southern Well Field area will adversely impact sustainable development of groundwater. The uncertainty associated with the potential for more public water supply wells to produce from the Fort Union aquifer must somehow be addressed in assessing sustainable development from this groundwater source in the future.

5.6.1.4 Uncertainty Summary

The existing level of uncertainty about the Fort Union aquifer is why this investigation was initiated. The foregoing analyses do nothing to dispel the uncertainty. In the absence of collecting more detailed information about the aquifer, a time consuming and expensive process, a way to reduce uncertainty may be to consider obtaining the future water supply from a different groundwater source that potentially offers less uncertainty.

5.6.1.5 Madison versus Fort Union Uncertainties

The previous section of this report presents some hypothetical models for the purpose of demonstrating how uncertainties about aquifer hydraulic parameters in the Fort Union aquifer can influence the outcome of modeling predictions from favorable to sustainable groundwater development to unfavorable to groundwater development. The latter analysis then goes on to show that determination of the hydraulic parameters includes a number of problems that increase the uncertainty. Ultimately, the predictions, based on a reasonable range of hydraulic parameters, straddle the divide between favorable and unfavorable conclusions about the Fort Union aquifer and tilt toward the unfavorable side. In order to analyze the performance of the Madison aquifer, the same model used for the hypothetical Fort Union aquifer well field is operated for a hypothetical Madison well field of the same configuration, but using hydraulic parameters and allowable drawdown limits for the Madison aquifer in the vicinity of the existing City of Gillette Madison well field.

In order to compare "apples to apples", the foregoing model was operated using hydraulic parameters appropriate for the Madison aquifer while retaining the original well field geometry and pumping rate used in the evaluation of the Fort Union. In an August 1981 report to the City of Gillette regarding testing of wells in the existing Madison well field, Anderson and Kelly summarized calculated values of aquifer transmissivity ranging from 15,450 gpd/ft to 5,000 gpd/ft (ignoring one value of 300,000 gpd/ft) and associated aquifer storativity values ranging from 5.4×10^{-5} to 3.5×10^{-4}. The smallest value of transmissivity, 5,000 gpd/ft obtained at two wells, and the smaller value of storativity associated with those same wells, 1.4×10^{-4}, were used in the model. The average depth to the top of the Madison aquifer is approximately 2,350 feet in the well field area and the static water level approximately 400 feet, providing an available water column of 1,950 feet which drawdown during sustained development should not exceed. Aquifer thickness was set to 650 feet in the model.

As shown on Figure 5.27, the average drawdown projected by the simulation for 20 years of pumping is between 400 and 425 feet for most of the wells in the well field. This projection is significantly less than the allowable drawdown of 1950 feet. On Figure 5.28, the average drawdown for 100 years of pumping is only slightly increased from that in the 20-year simulation, ranging from 475 to 500 feet for most of the wells. In both instances, the hydraulic properties of the aquifer provide a different type of drawdown distribution than occurred in the Fort Union simulations. These projections are highly conservative, being based on the smallest values of transmissivity and aquifer storativity obtained from tests of Madison wells in the real-world well field. Accordingly, the projections probably overstate future drawdown in the well field.

Figure 5.27
20-Year Average Drawdown for 15 Madison Aquifer Wells

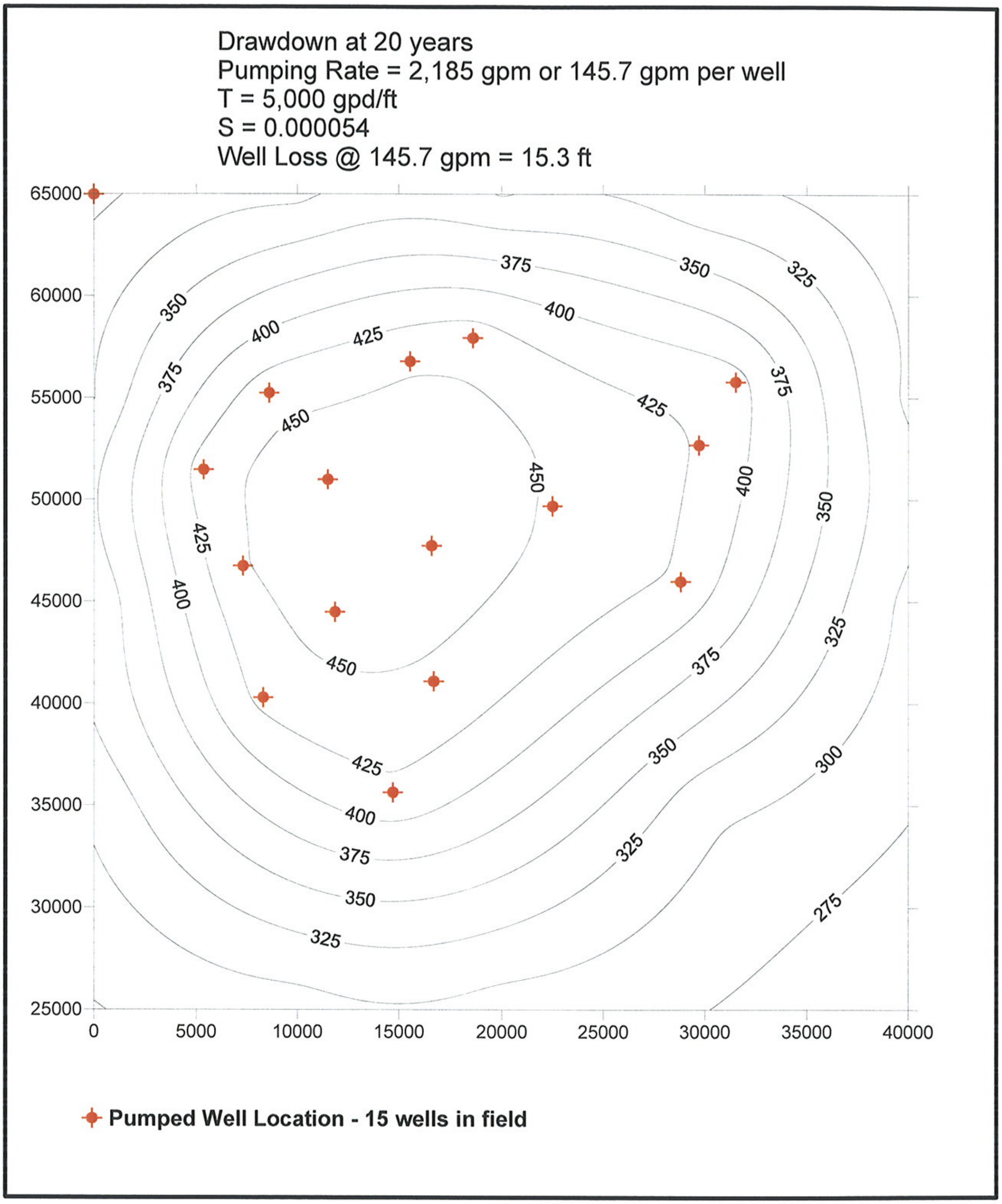
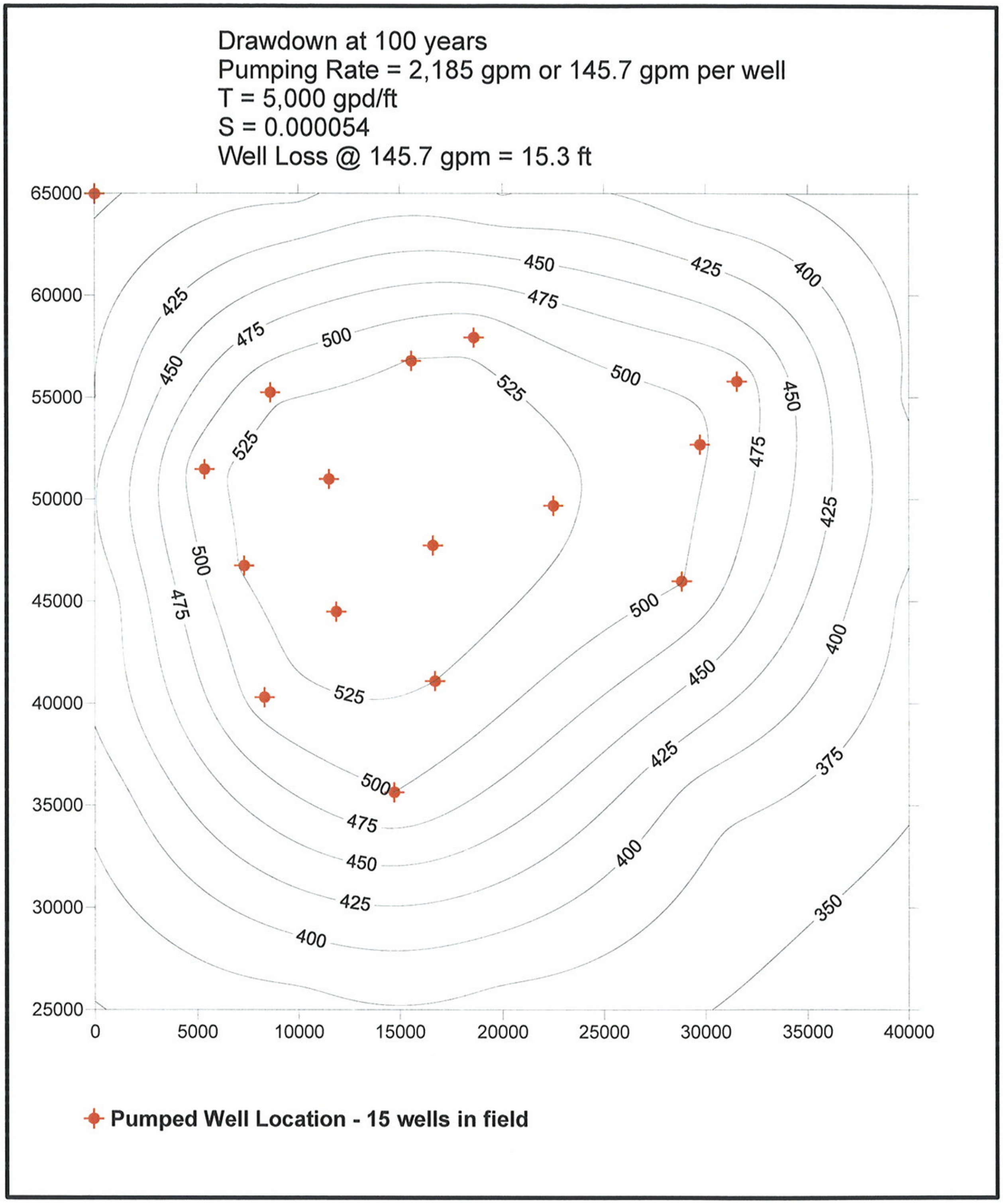


Figure 5.28
100-Year Average Drawdown in 15 Madison Aquifer Wells



However, the importance of the projected Madison aquifer drawdown on Figures 5.27 and 5.28 is not the fact that the projected drawdowns are smaller than those for the Fort Union aquifer simulations. The significance is in the fact that the drawdown projected by the simulations predicts that only 25 to 26 percent of the available water column will be used in 100 years of sustained development of 2,185 gpm of water supply. Accordingly, the aquifer will support a larger amount of sustained groundwater development than assumed in the simulations.

More importantly, the huge difference between the projected drawdown and the available water column in the wells means that the analysis is not as sensitive to aquifer variability and other sources of uncertainty as the analysis of the Fort Union aquifer. The Madison aquifer analysis shows that there is considerable margin for error and uncertainty in projecting sustainable groundwater development from the Madison aquifer of the magnitude needed in the future by the City of Gillette whereas there is essentially no margin for error when sustained development of the same magnitude is projected for the Fort Union aquifer. Accordingly, it is suggested that the Madison aquifer is potentially a more attractive source because it offers less uncertainty.

5.7 MADISON AQUIFER TEST RESULTS

The foregoing comparison of Madison aquifer response to Fort Union aquifer response is based on substitution of Madison aquifer hydraulic parameters into the model for the Wester-Wetstein (2004) recommended well field array for the proposed Southern Well Field. The Madison aquifer hydraulic values substituted into the Southern Well Field model were obtained from Anderson & Kelly Consultant's reports presenting their findings at the end of constructing and testing the first eight Madison wells for Gillette in 1980. Although the foregoing comparison serves to show the considerable contrast between the predicted responses of the two aquifers, it also raises a question.

The question raised by the results of the comparative models is why the drawdown predicted in the Madison aquifer model is so much greater than the observed changes in groundwater levels in the Madison well field after 12 years of well field operation. The actual separation between the existing Madison wells is much less than that used in the model, a factor that should cause real drawdown to be greater than that predicted by the model, not less as observed in actual operations. The actual pumping rates used in the Madison well field are considerably greater than those used in the model simulations, again suggesting actual drawdown should be more than that predicted in the model, not less as actually observed.

It might be argued that the pumping duration in actual operation of the Madison well field has historically been less than that used in the model. That may have been true in the early years of the well field when it was used as a seasonal water supply; however, in recent years the Madison well field has been used throughout most of the year at relatively high pumping rates. Comparison of all of these factors to the model simulations indicates that the model used to contrast the Madison aquifer to the Fort Union aquifer in response to the modeled pumping scheme greatly overestimates Madison aquifer drawdown, compared to the drawdown observed in actual operations. The question then becomes that of why the model over-predicts drawdown. What factors are present in the hydraulic properties of the Madison aquifer and its hydraulic response to pumping stress that differ from the assumptions used in the model?

In an attempt to answer the foregoing questions, the original pumping test data from the first eight Madison wells completed in 1980 are analyzed herein. The test data are for single-well tests, in other words, data are from the pumped wells, not from observation wells. When each production well was tested in 1980, the aquifer response was measured manually in the pumped well with an electrical water level indicator tape or airline and monitored in the other production wells with float-equipped Steven's recorders. Only those production wells closest to the one being pumped showed any response. If the monitoring well records still exist as tabulations or recorder charts, they were not found for use in this study. Only the data provided in the reports prepared by Anderson & Kelley consultants for each pumped well are used herein.

In their reports about the drilling and testing of the individual wells, Anderson & Kelley divided the first eight production wells into "cavity" and "non-cavity" wells, stating that the response of the cavity wells to pumping was markedly different and involved less drawdown than the response of the non-cavity wells. Cavity wells are those wherein the drilling tools penetrated voids in the Madison Limestone that resulted in noticeable "rod drops", i.e., the drilling tools dropped through each void until they encountered hard rock at the bottom of the void. The cavity wells were M-2 and M-4. Wells M-1, M-3, M-5, M-6, M-7, and M-8 were non-cavity wells where voids in the limestone were not penetrated.

Preliminary analysis of the pumping tests for the eight Madison wells determined that most of the test results are greatly influenced by the pumping test procedures. This included initially high discharge rates at the start or a test, irregular changes in the discharge rates, and other adjustments or equipment problems that deviate from standard test requirements to maintain a constant discharge rate within 95 percent of the nominal rate. Accordingly, most of the pumping test data are of limited use in calculating aquifer hydraulic parameters; however, they are sufficient to support conclusions about the types of aquifer responses obtained during the tests. They are also sufficient to show that the aquifer response in wells penetrating voids in the limestone is completely different from that in wells that do not penetrate voids.

The time-drawdown data from two of the non-cavity wells, M-5 and M-6, are adequate to support calculation of the aquifer transmissivity in the non-cavity groundwater flow. Data from the M-2 and M-4 wells are adequate to support a conclusion about the type of hydraulic flow occurring in the cavities penetrated by the wells, but are not adequate to support calculation of aquifer transmissivity. None of the data from the pumped wells support calculation of aquifer storativity because storativity cannot be calculated from single-well tests.

5.7.1 Wells M-5 and M-6

The best test results from a non-cavity well are the data from well M-6. Therefore, the first aquifer test evaluated herein is that conducted in well M-6 so that the results can be used for comparison to the other tests, each of which involves a departure from the conventional testing procedures requiring constant discharge rates.

The first step in any type of aquifer test interpretation is to determine the type of groundwater flow to the pumped well and the appropriate analytical solution to apply to that type of groundwater flow. In general, the flow of groundwater to wells falls into two categories – radial flow and linear flow. The diagnostic time-drawdown plot for any type of well test is logarithmic drawdown versus logarithmic time, i.e., the so-called double log or log-log plot. Radial flow on a log-log plot is characterized by the Theis equation (Theis, 1935) for confined radial flow and the Neuman equations (Neuman, 1974; 1979) for unconfined radial flow. Linear flow is characterized by a log-log straight line. These types of plots are referred to as "type curves".

Radial and linear flow may be further modified by various aquifer boundary conditions, recharge, interference drawdown, double porosity, non-instantaneous release of storage, capillarity, and other factors. Various analytical solutions have been developed to numerically evaluate aquifer test results under various combinations of the foregoing factors. In addition to the diagnostic log-log plots, the analytical solutions dealing with one or another modification of radial or linear flow utilize specialized plots which also may be diagnostic of certain types of flow and easier to interpret than the log-log plot. The well-known Cooper-Jacob (1946) semi-log, straight-line solution for radial flow is one such example and is simply a special case of the Theis (1936) non-equilibrium equation that describes the Theis type curve.

Figure 5-29 shows the diagnostic log-log, time-drawdown plot of the constant rate test of well M-6. The data on Figure 5-28 conform to the typical shape of a log-log plot for radial flow to the pumped well (type curve not shown) until about 3,000 minutes of elapsed pumping time. After approximately 50 hours of elapsed pumping time, drawdown stabilized. The stabilized drawdown is easier to perceive on an arithmetic (Cartesian) plot of the data as shown on Figure 5-30. Verification of radial flow response in the data prior to 3,000 minutes is provided on Figure 5-31 with a semilogarithmic Cooper-Jacob plot. The data plot in a semilogarithmic straight line on Figure 5-30 which is diagnostic of radial flow to the pumped well. The data on Figure 5-31 also support a solution for the transmissivity of the aquifer, yielding a value of 2,726 gpd/ft for non-cavity flow in the limestone aquifer.

Figure 5-29
Log-Log Time-Drawdown Plot of Well M-6 Constant Rate Test

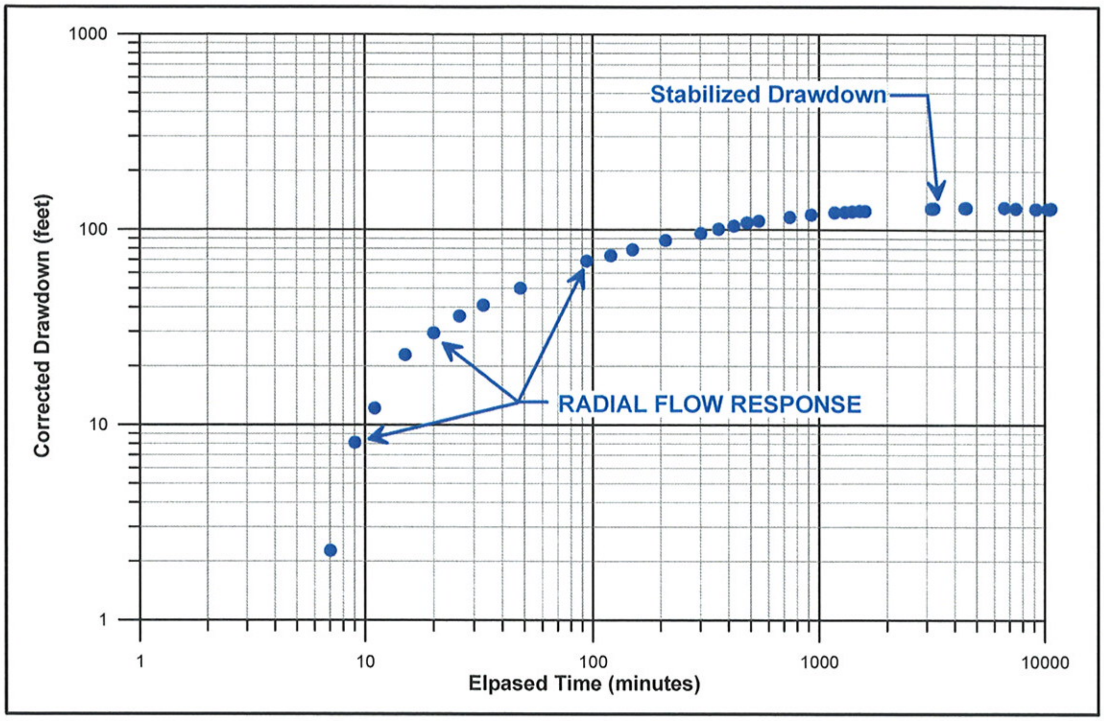


Figure 5-30
Arithmetic Plot of Well M-6 Data Showing Stabilized Drawdown

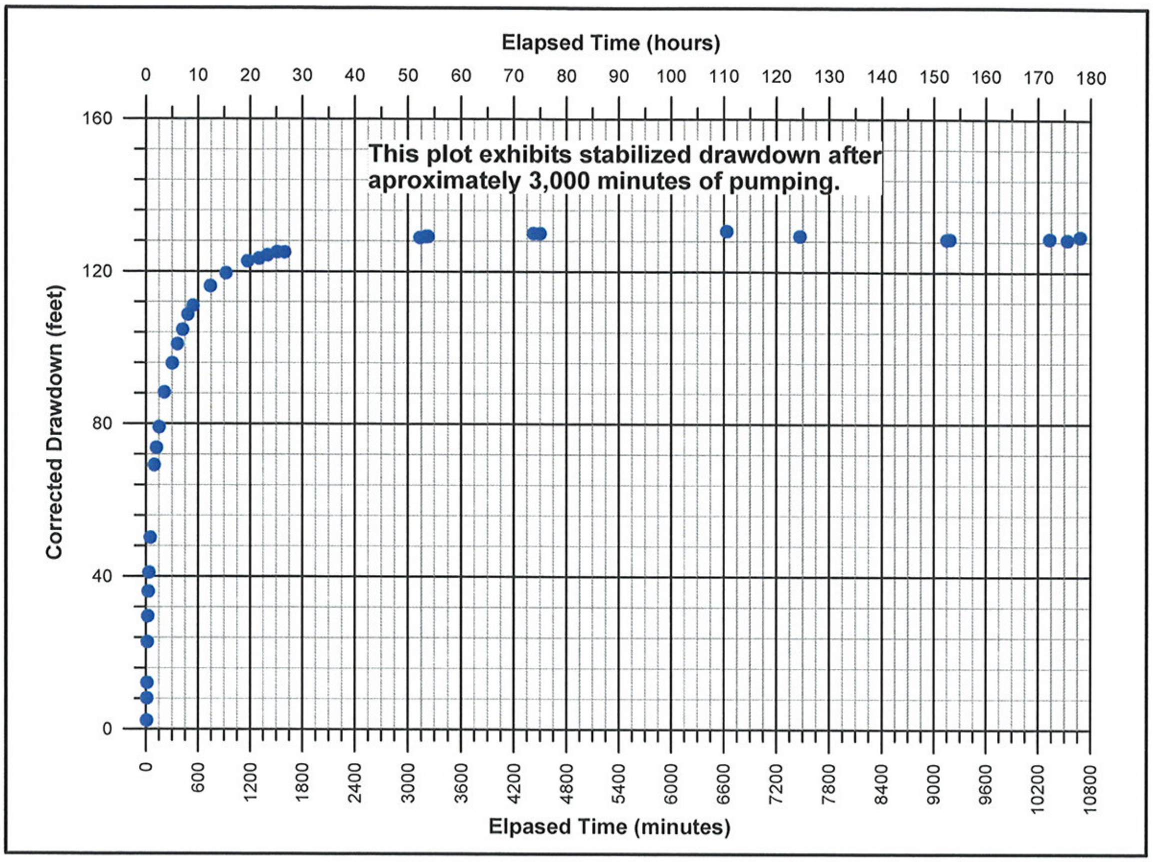
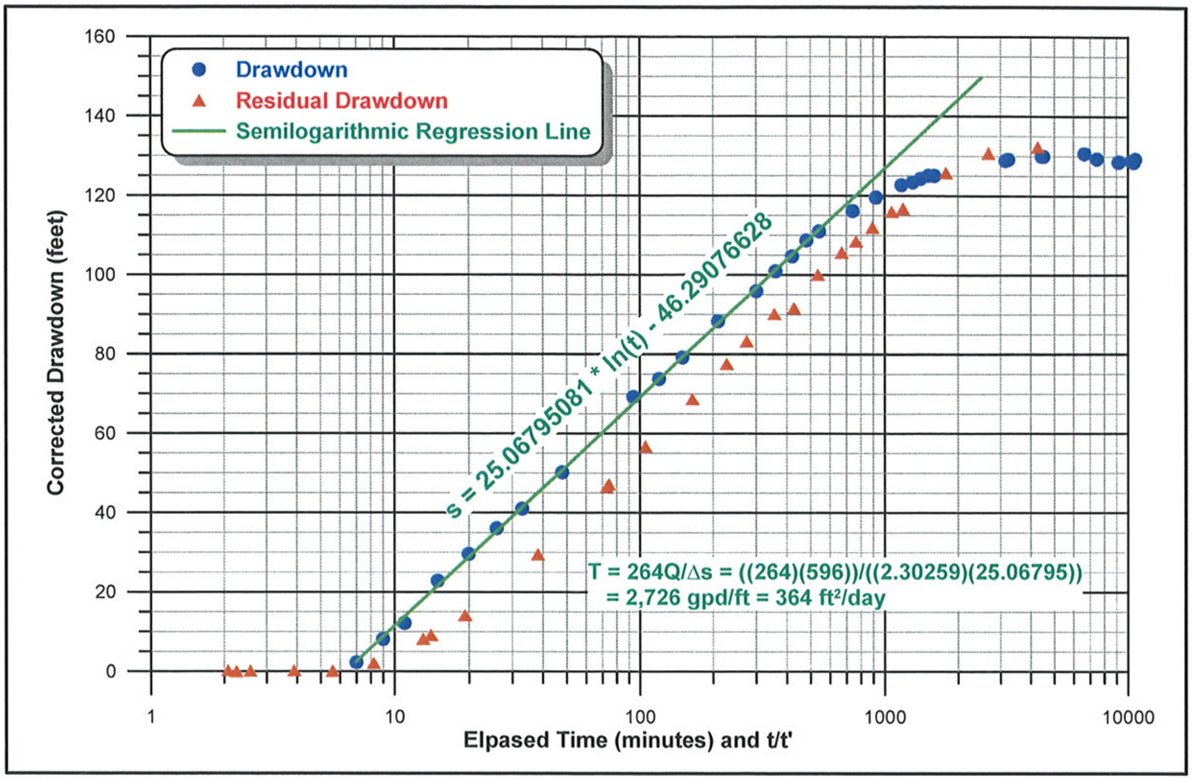


Figure 5-31
Cooper-Jacob Solution for Well M-6 Constant Rate Test

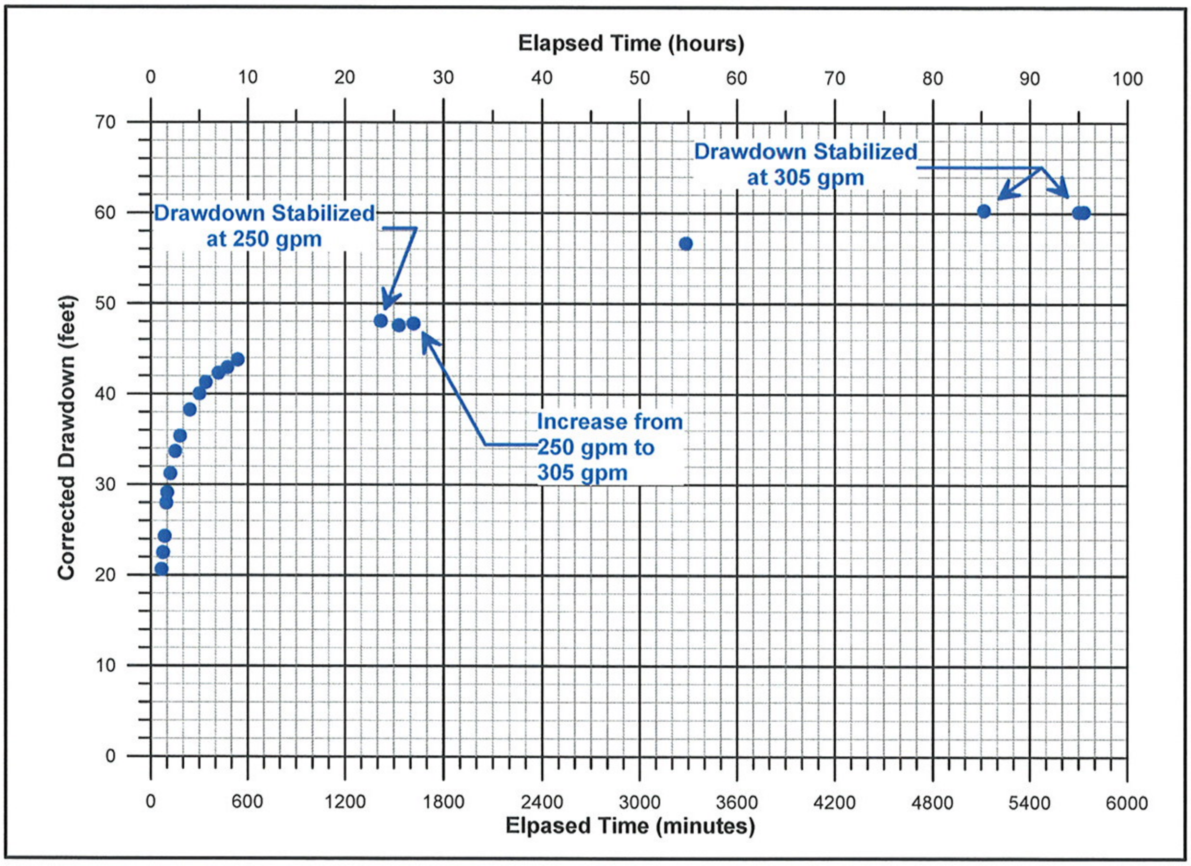


The vertical separation between the residual drawdown (plotted versus t/t' where t is time since pumping started and t' is time since pumping stopped) indicates 10 to 11 feet of well loss drawdown. The full recovery of drawdown well before t/t' = 2 indicates the cone of depression around the pumped well was receiving recharge from an external source. The fact the drawdown and recovery curves are parallel indicates the source of recharge acted constantly throughout the test and, therefore, was proportional to the footprint of the cone of depression (depressurized area) around the pumped well. The latter observations are consistent with stabilized drawdown after 50 hours of pumping. The stabilized drawdown is the result of vertical leakage into the Madison aquifer from an adjacent aquifer separated from the Madison by a poorly confining layer. In this case, the geologic relationships indicate the Minnelusa Formation is the most likely source of the vertical leakage. Accordingly, the aquifer response shown on Figure 5-29 and Figure 5-31 is the so-called "leaky confined aquifer" response.

The response of the aquifer at well M-5 to constant rate pumping is similar to that at well M-6, but with two complications. One complication is a significant amount of drawdown in the pumped well due to well loss, i.e., the differential head required to overcome resistance to radial convergence of groundwater flow in the area near the well. A second complication is that the pumping rate was increased from 250 gpm to 305 gpm at approximately 1,420 minutes, thus shifting the time-drawdown curve.

Figure 5-32 is a Cartesian plot of the time-drawdown data providing the clearest presentation of the fact that drawdown stabilized at the end of both the 250-gpm and 305-gpm pumping rates, again demonstrating vertical leakage into the confined aquifer.

Figure 5-32
Cartesian Plot of Well M-5 Time-Drawdown Response



The data shown on Figure 5-32 are corrected for well loss drawdown. Well loss drawdown in well M-5 is significant and affects the shape of the diagnostic log-log plot. The estimation of well loss drawdown and drawdown corrected for well loss are shown on Figure 5-33. The well loss is estimated because data collection did not start until 65 minutes of pumping time and the recovery was not recorded through t/t' = 2. Corrected drawdown is used on the remainder of the plots for well M-5.

Figure 5-33
Well Loss And Drawdown Corrected For Well Loss In Well M-5

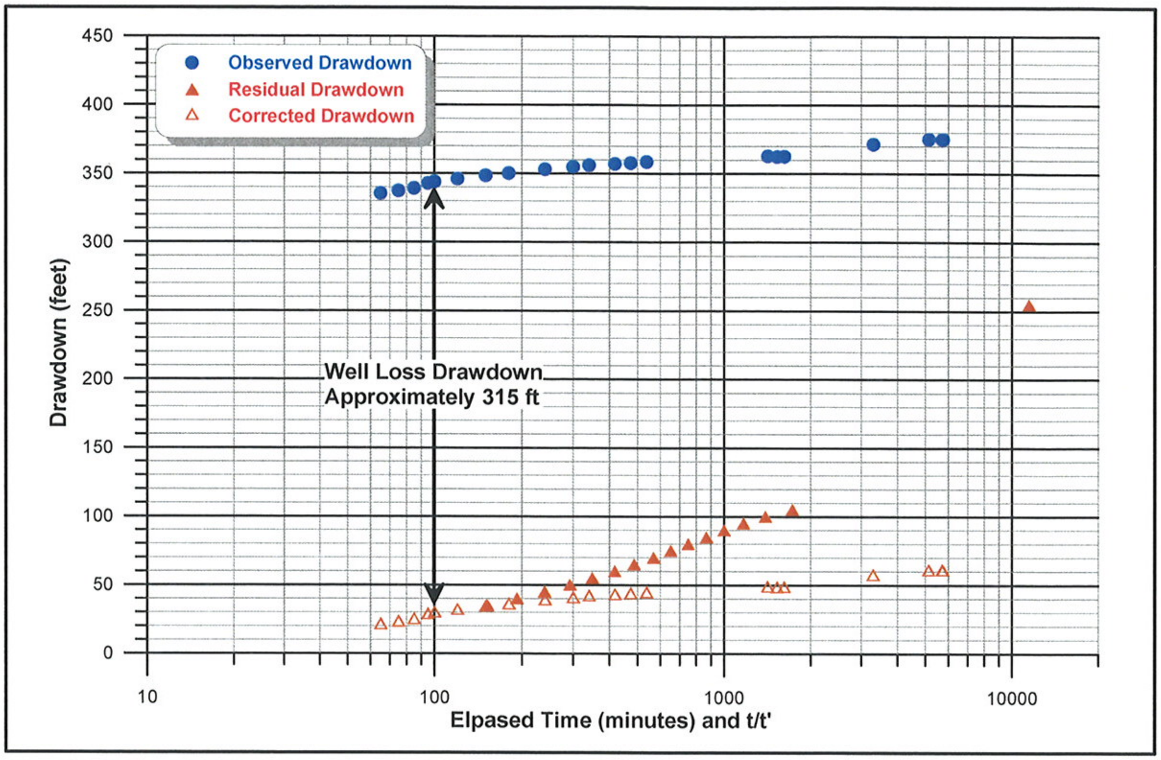
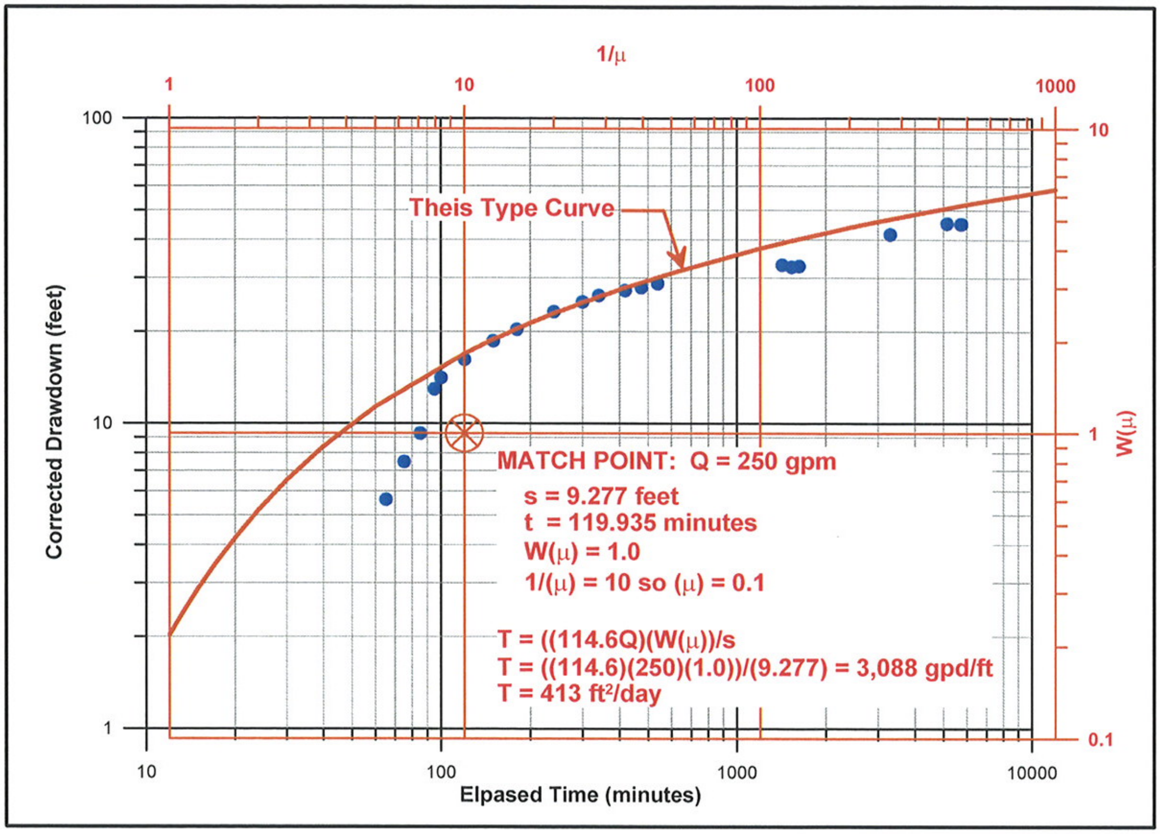


Figure 5-34
Theis Solution for Aquifer Transmissivity at Well M-5



As shown on Figure 5-34, the Theis solution for non-cavity limestone aquifer flow at well M-5 is 3,088 gpd/ft. This compares favorably to the value of 2,726 gpd/ft calculated from the test of well M-6 and gives some idea of the potential variability of transmissivity for non-cavity flow in the limestone. The foregoing transmissivity values calculated for non-cavity flow in the Madison Limestone aquifer are very similar to the value of 2,709 gpd/ft determined for multiple sandstone layers at the ASR test well in the Fort Union Formation. This leads to the question of why the Madison wells are significantly more productive than those in the Fort Union Formation if the aquifer transmissivities are similar.

Part of the answer to this question is that vertical leakage into the Madison Limestone when it is depressurized, presumably from the overlying Minnelusa Formation, ultimately replaces removal of water from storage in the Madison Limestone as the source of pumped water. When groundwater is no longer removed from storage in the Madison aquifer, because vertical leakage equals the pumping rate, drawdown in the Madison aquifer wells stabilizes, allowing the wells to be pumped at a much higher rate than wells in the Fort Union aquifer with comparable aquifer transmissivity, but no stabilization of pumping water levels. The second part of the answer to the foregoing question is that the Madison aquifer's penetrating voids in the limestone, i.e., the so-called cavity wells, are much more hydraulically efficient and produce greater yields than the non-cavity wells in the Madison aquifer or the wells in the Fort Union aquifer.

5.7.2 Wells M-2 and M-4

The data for an aquifer test of a Madison aquifer well penetrating a void and which are the least affected by irregularities during the test are those from well M-4, although only one residual drawdown measurement was recorded. The M-4 test data also contain the most useful set of step test data for determination of well loss drawdown. Figures 5-35 and 5-36 show the step test plots and resultant specific capacity curve. The recovery curve for the step test shows the water level recovered to 8.85 feet above the initial static level during the first minute, presumably a surge due to drainage of the pump column. The recovery water levels then decline to approximately one foot above static level and slowly decline to 0.3 feet above static level over the next two hours. The latter differences between the original static level and the recovered water levels probably reflect barometric pressure effects on the water level in the confined Madison aquifer.

Figure 5-36 is a plot of specific capacity (pumping rate divided by drawdown) versus pumping rate. The type of specific capacity curve shown on Figure 5-36 is normally a straight line. The shape of the curve for well M-4 on Figure 5-36 reflects the influence of a so-called "negative" or "no flow" boundary in the aquifer system. In other words, the cone of depression around the pumped well expanded against an area of the aquifer that was not releasing groundwater from storage at as large a rate as the rest of the depressurized part of the aquifer, perhaps an impervious no-flow boundary. The specific capacity curve on Figure 5-36 is represented by a third order polynomial, the equation for which is shown on the figure. It can be seen that the polynomial projection does not fit the trend of the data very well between the last two data points for the highest pumping rates. This is because of a very small amount of data scatter at these points; however, it has a significant effect on use of the specific capacity curve in the analysis.

Namely, the well loss drawdown must be calculated from the specific capacity curve. If the polynomial equation is used to project the specific capacity for the constant rate test discharge of 656 gpm, well loss is over-predicted and corrected drawdown values for the beginning of the constant rate test become negative. Visual inspection of the specific capacity curve data points indicates the specific capacity should be about 10.5 gpm/ft-dd for 656 gpm. This results in a well loss of 62.48 feet. For reasons that will be explained below, this value is likely to more properly be about 62.25 feet of well loss drawdown, resulting in a correction to the diagnostic log-log curve for the 656-gpm constant rate test that is most consistent with this type of aquifer response. Figure 5-37 shows the log-log plot for drawdown corrected for well loss of 62.25 feet.

Figure 5-35
Step Test Time-Drawdown Plots for Well M-4

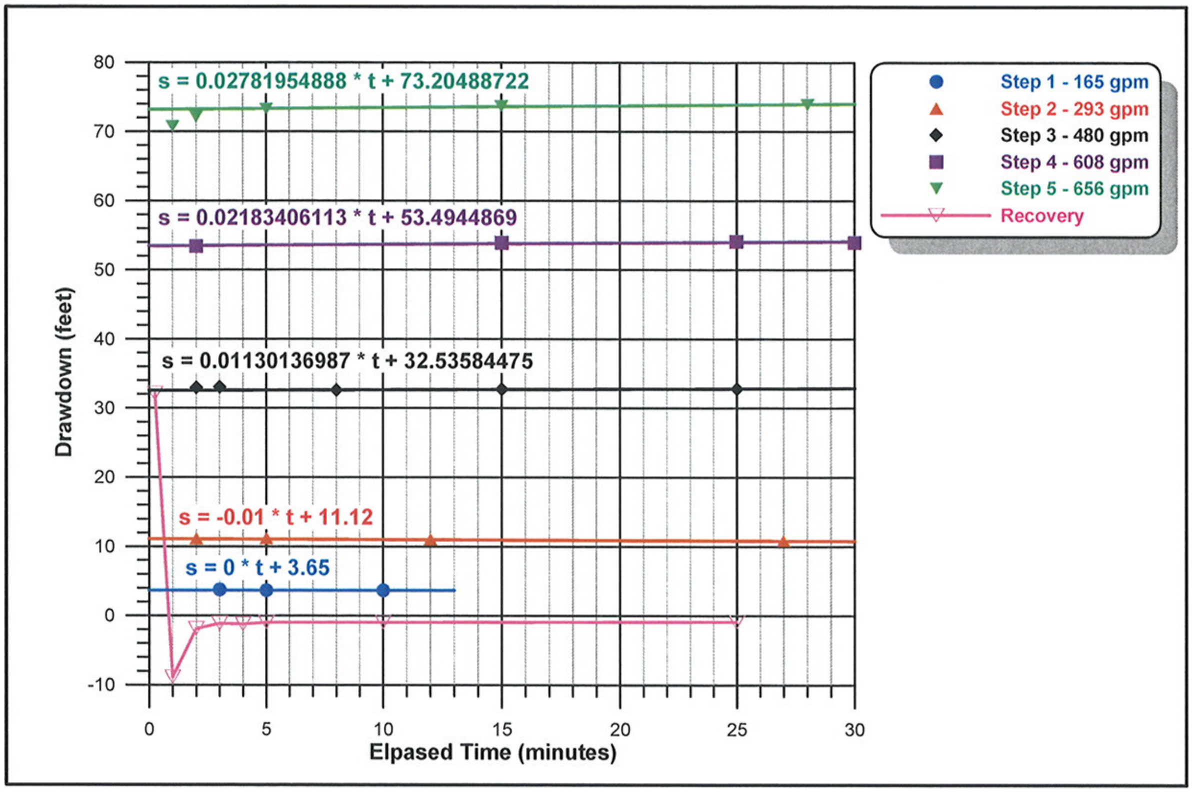


Figure 5-36
Specific Capacity Curve for Well M-4

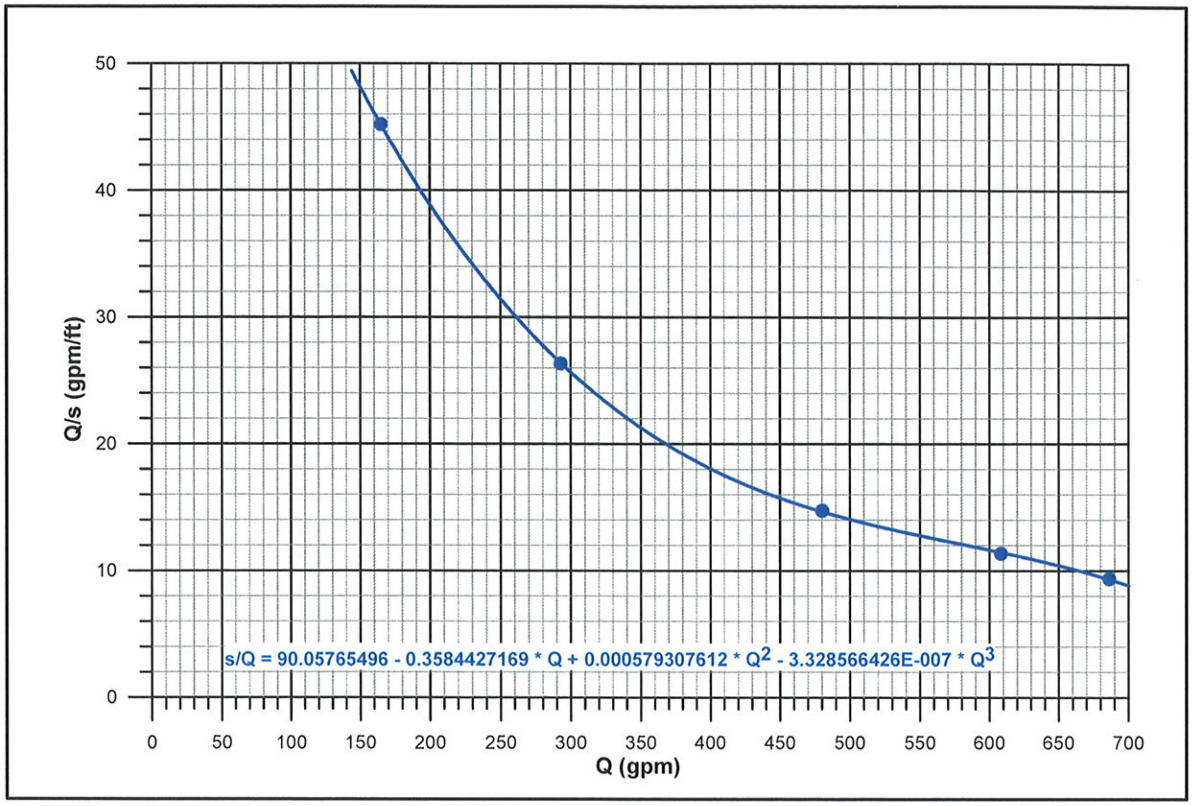


Figure 5-37
Diagnostic Log-Log Plot For Well M-4 At 656 gpm

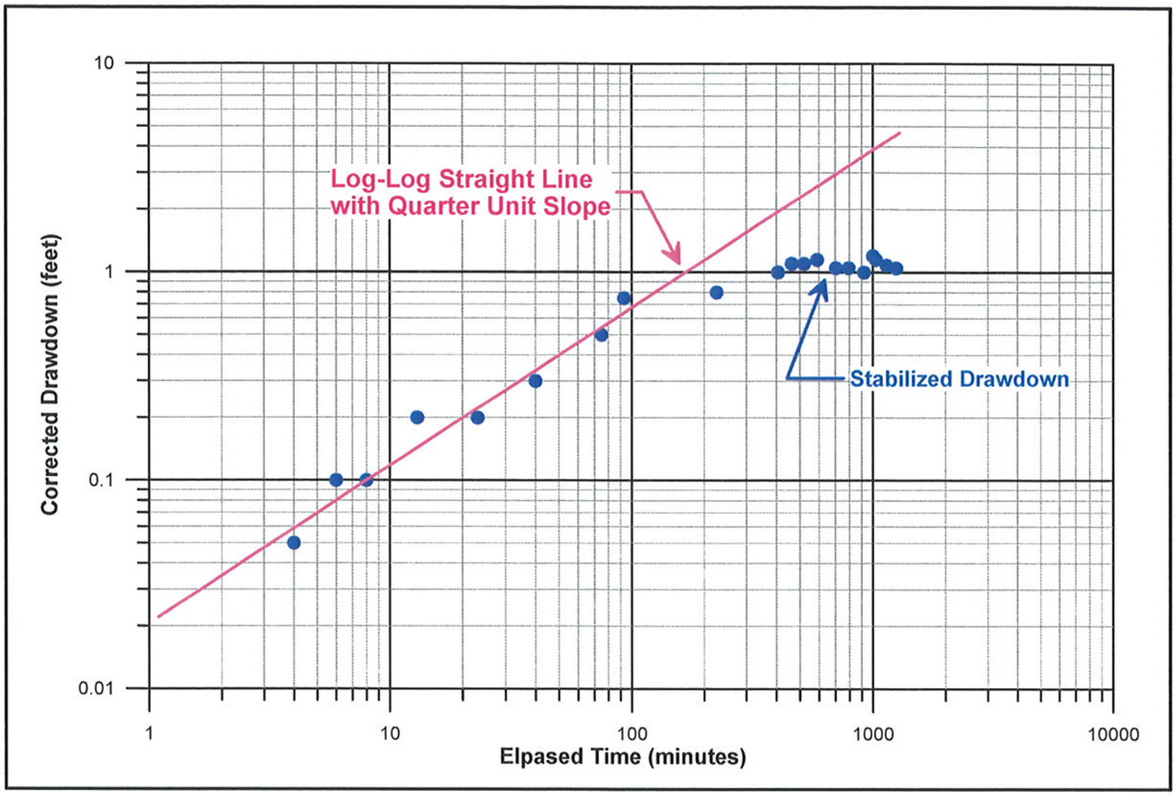
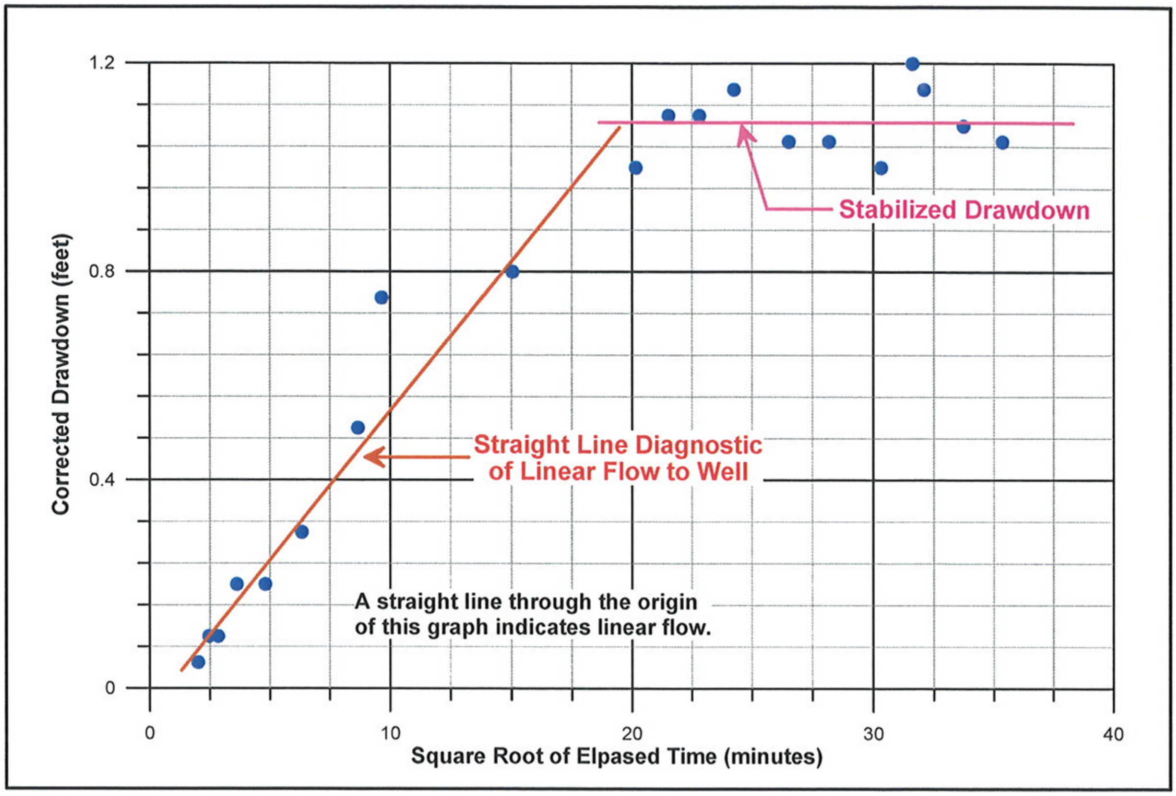


Figure 5-38
Plot Of Drawdown Versus the Square Root of Time for Well M-4



The log-log time-drawdown plot on Figure 5-37 produces a log-log straight line for the first 90 minutes of pumping. A log-log straight line is diagnostic of linear flow to the pumped well. Linear flow occurs in many forms in nature including strip aquifers, plane fractures with no significant storage, relatively high transmissivity dikes intruded into relatively low transmissivity aquifers, and linear shear zones that contain significant groundwater storage and may or may not receive additional inflow from the rock mass containing the shear zone, to name a few. The most likely cause of linear flow through the aquifer to well M-4 is one or more interconnected voids, i.e., subaqueous caverns, as indicated by penetration of voids by the well and by the well-known cavernous nature of the upper part of the Madison Limestone.

It should be noted that the author has conducted pumping tests of wells penetrating abandoned underground coal mines in the Fort Union Formation and found that those voids provide linear flow response with log-log straight lines exhibiting half unit and quarter unit slopes, so this type of response from a void in an aquifer is not an unknown phenomena.

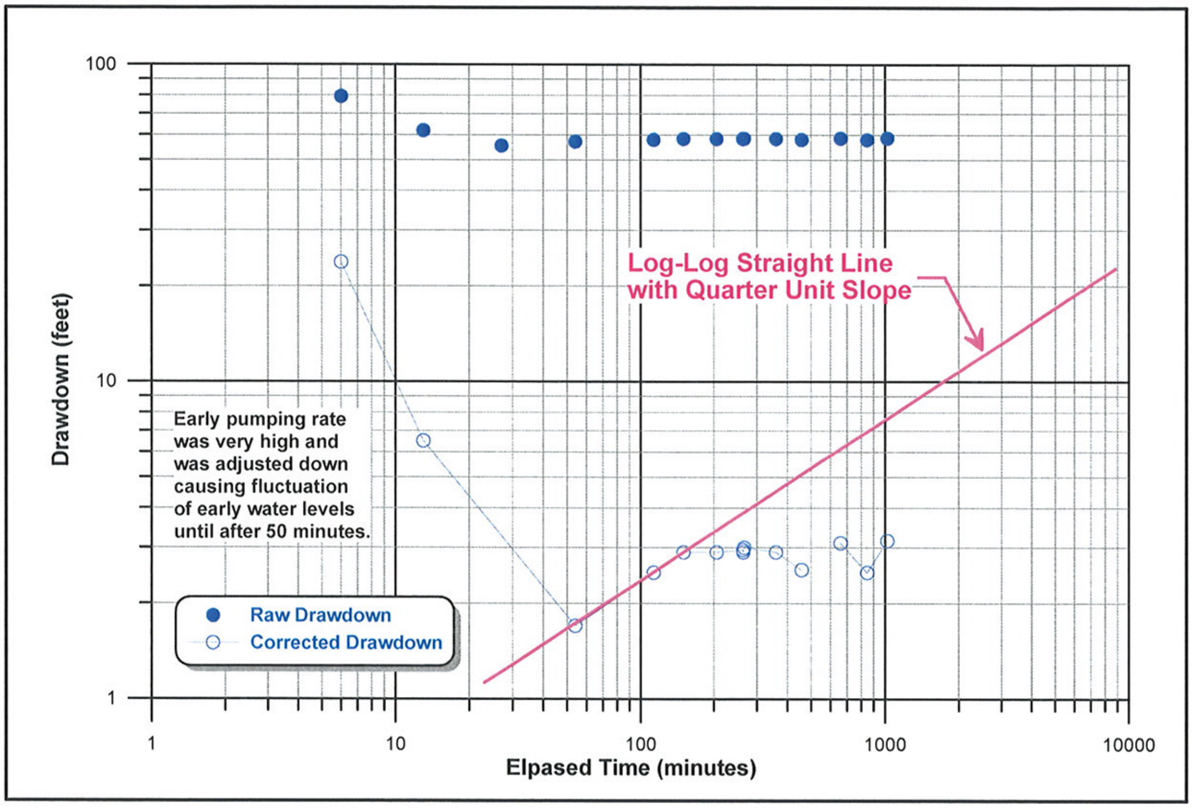
The log-log straight line on Figure 5-37 exhibits a quarter unit slope. A quarter unit slope is universally indicative of a linear shear zone, strip aquifer, or solution enlarged fracture (cavern) which has ceased releasing groundwater storage to the pumped well and become a conveyance system for groundwater released from storage in the rock containing the linear feature. In the case of well M-4, the quarter unit slope response suggests a cavern, probably developed along a relatively linear fracture, which has been penetrated by the well and is conveying to the pumped well water released from storage in the surrounding limestone.

The specialized plot used to verify linear flow to a pumped well is a plot of arithmetic drawdown versus the square root of elapsed pumping time, as shown on Figure 5-38. Linear flow response on the latter plot is a straight line through the origin, as is shown on Figure 5-38. The quarter unit slope of the corrected drawdown data on Figure 5-36 and the alignment of the plot through the origin of Figure 5-38 both require the specific capacity of the well (Q/s) to be 10.538 gpm/ft-dd and the well loss to be 62.25 feet. The fact that a well loss correction of 62.25 feet results in both a quarter unit slope of the log-log straight line and a straight line through the origin of the specialized plot on Figure 3-38 confirms that this is the correct value of well loss for the 656-gpm pumping rate in this well.

Figure 3-38 shows that after 400 minutes of pumping time (square root time equal to 20), drawdown stabilizes similar to the aquifer responses in the non-cavity wells. The stabilized drawdown indicates that vertical leakage into the depressurized zone along the linear cavern or solution-enlarged fracture and the aquifer rock bounding the fracture became equal to the pumping rate after 400 minutes. It should be noted that during the log-log straight line response with a quarter unit slope, the flow of water out of the aquifer material and into the linear flow feature such as a fracture or cavern is essentially perpendicular to the linear feature and has not expanded out far enough away from the linear feature far enough to result in pseudo-radial flow in an elongate cone of depression centered on an axis along the linear cavern or enlarged fracture. Accordingly, the width of the zone receiving vertical leakage is relatively narrow, as constrained by the hydraulic properties of the aquifer rock forming the walls of the linear flow feature.

The aquifer response observed in well M-4 is easy to recognize as a linear flow response. The aquifer response observed in well M-2, the other well that penetrated a void in the Madison Limestone, is not as easy to recognize as a linear flow response because the initial pumping rate was very high and an adjustment of that pumping rate to a lower pumping rate resulted in a fluctuation in the observed drawdown during the early part of the test that somewhat obscures the type of aquifer response obtained. The time-drawdown data from well M-2 are shown on Figure 5-39 in a log-log plot. As shown on Figure 5-39, a log-log straight line of quarter unit slope fits through the last three data points before the drawdown stabilized, suggesting a linear flow response is likely, despite the paucity of data.

Figure 5-39
Log-Log Plot of 600-Gpm Test Data from Well M-2



Analytical solutions for linear flow response, as obtained from the Madison aquifer at well M-4 and probably at M-2, are published in the literature; however, they are mathematically rigorous and must be solved by methods that are beyond the scope of this analysis. Therefore, solutions for Madison aquifer transmissivity are not provided herein for wells exhibiting linear flow response. Likewise, transmissivity values provided in the original well completion reports in the 1980's for the Madison wells exhibiting linear or non-radial flow response cannot possibly be valid. This fact was recognized by Anderson & Kelly consultants who stated in their reports that some of the aquifer responses to testing were not amenable to conventional solutions. Put in the context of the time, "conventional" solutions were limited to the Theis (1935) and Cooper-

Jacob (1946) solutions, i.e., radial flow solutions that are inappropriate for linear flow responses. Accordingly, one of the potential problems with the 1980 interpretations was failure to distinguish between radial flow and linear flow responses. The results of tests conducted at well M-3 are a good example of this problem.

5.7.3 Well M-3

The failure to recognize a linear flow response is characterized by the tests of well M-3 in 1991, prior to and following hydraulic fracture stimulation of the well. In a May 1992 report by James M. Montgomery, Consulting Engineers, titled, "Madison Well Field, Well M-3 Enhancement," the results of a pre-fracturing constant rate test of well M-3 at 642 gpm are described as follows:

> "***The drawdown data were analyzed using the straight-line method developed by Cooper and Jacob (1946) for tests of well in confined aquifers, and are presented graphically on Figure 4. The initially high pumping rate during the first few minutes of the test caused the early-time drawdown to deviate from that predicted for the final long-term pumping rate, as shown by the data for the first three minutes of the test. Adjustment of the pumping rate to approximately the long-term average rate allowed the drawdown to approach the predicted trend between three and seven minutes into the test. From seven minutes until 710 minutes into the test, the drawdown data fit the predicted trend fairly well. The drawdown stabilized at 218 feet at 710 minutes into the test and remained at that point until the end of the test. Analysis of the drawdown data indicated a transmissivity of 4,400 gallons per day per foot (gpd/ft), and the stabilized water level near the end of the test allowed calculation of a specific capacity of 2.9 gallons per minute per foot of drawdown (gpm/ft) at the 642 gpm average pumping rate.***" (Emphsis added)

Figure 5-40 is a reproduction of the Cooper-Jacob straight-line plot from the J.M. Montgomery (1992) report. To all outward appearances, the semilogarithmic straight-line on Figure 5-40 (Figure 4 from the J.M. Montgomery (1992) report) is a good fit to the data. However, this approach overlooks the basic principle of correcting the data for well loss and examining the diagnostic log-log plot of the corrected data. Figure 5-41 shows a semilogarithmic plot of the same time-drawdown data plotted on Figure 5-40, used to determine the correction for well loss. Again the drawdown data appear to define a semilogarithmic straight line; however, the residual drawdown data do not define a parallel straight line during recovery from pumping. Figure 5-42 shows the drawdown values corrected for well loss plotted on the diagnostic log-log plot. The data plot as log-log straight lines, indicating flow to well M-3 was linear flow.

Figure 5-40
Reproduction of J.M. Montgomery Plot of Pre-Fracture Test of Well M-3

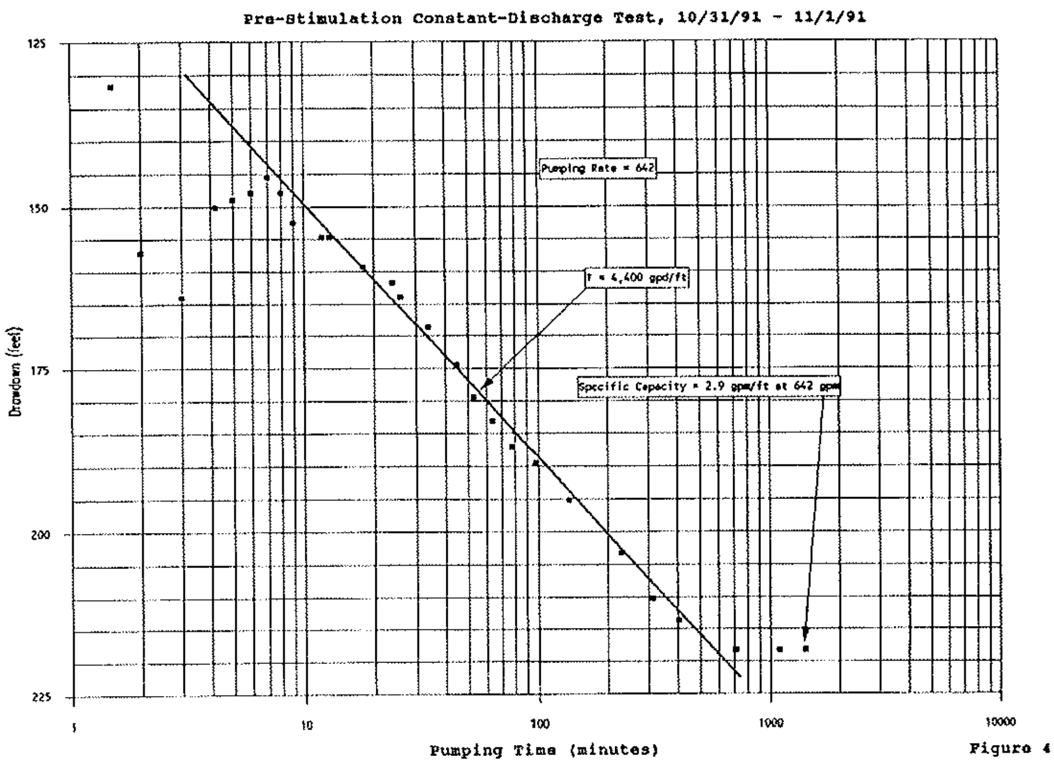


Figure 5-41
Pre-Frac Test of Well M-3 With Residual Drawdown and Well Loss

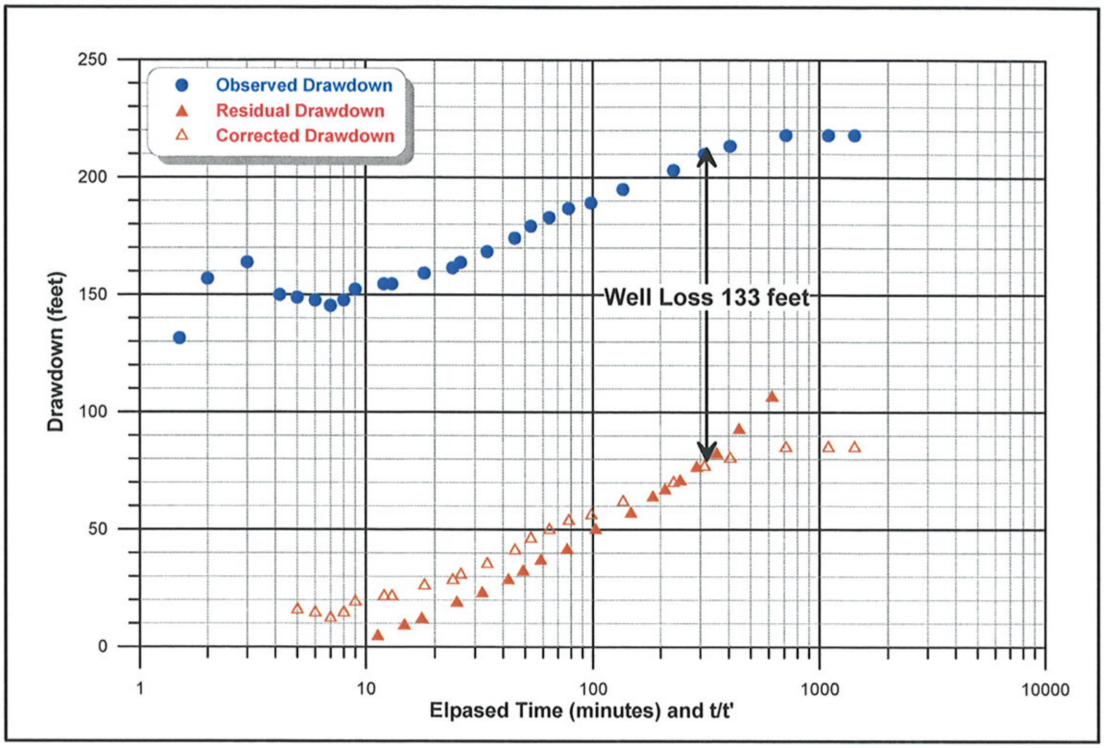
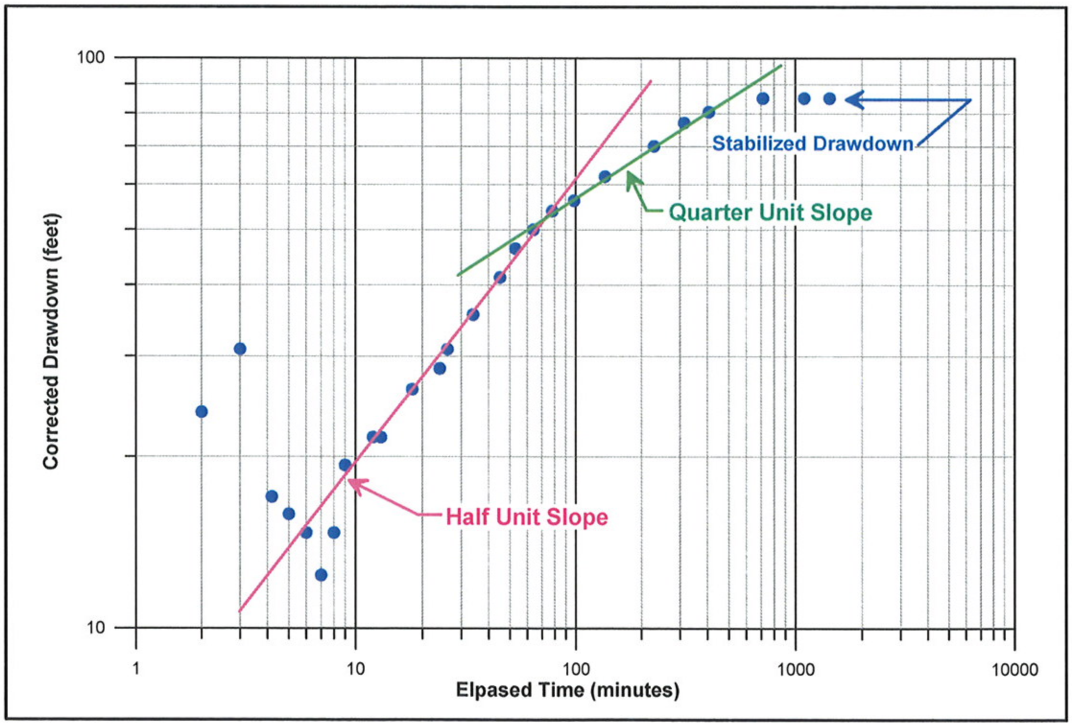


Figure 5-42
Diagnostic Log-Log Plot of Corrected Pre-Frac Drawdown at Well M-3



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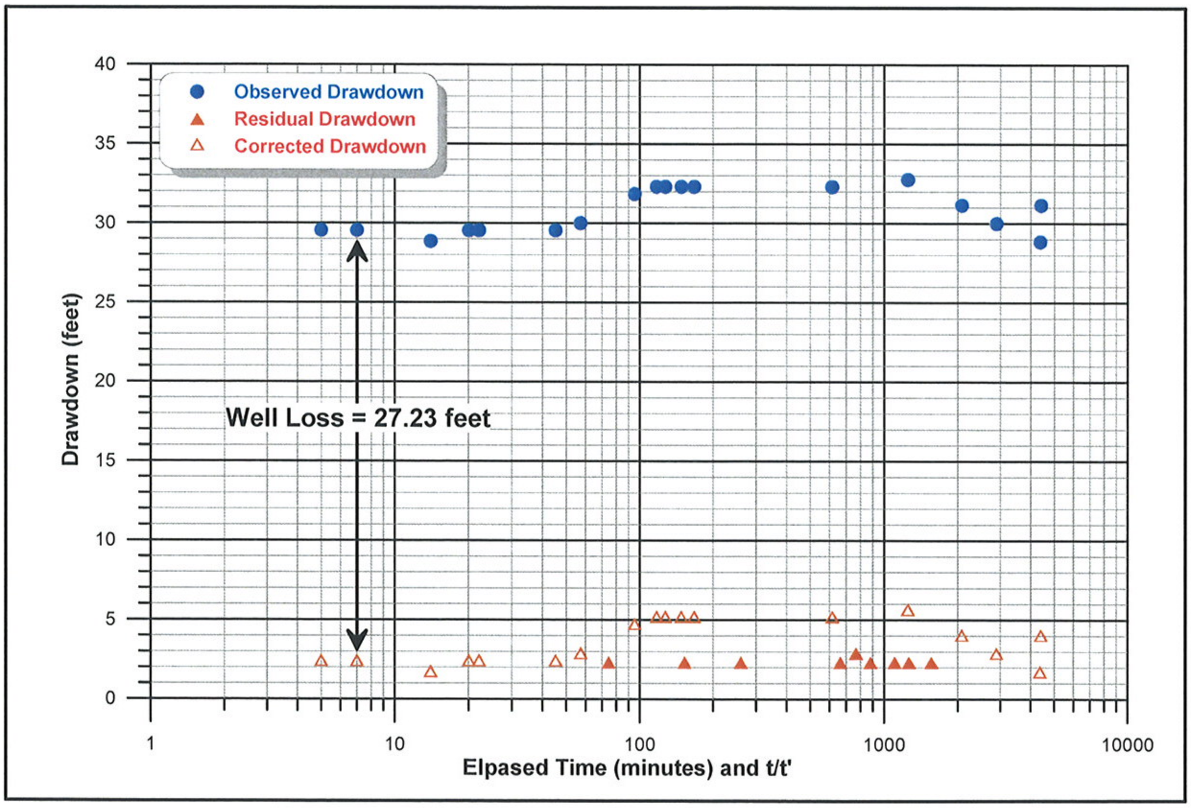
The diagnostic plot of the pre-frac test data from well M-3 not only demonstrate that flow to this well was along a strong linear feature such as a solution enlarged joint or fracture in the limestone, but reveals additional information about the hydraulic performance of the aquifer. The log-log line of half unit slope in the first 80 minutes of the pumping test is diagnostic of release of groundwater storage from the fracture or joint-controlled cavern. The subsequent log-log line of quarter unit slope is indicative of a change from release of groundwater storage from the linear feature during the half unit slope response to release of groundwater storage from the surrounding rock during the quarter unit slope response. In the quarter unit slope response, the fracture or cavern has been relegated to the role of a conveyance conduit, but is not providing the water pumped from the well. The water pumped from the well during the quarter unit slope response is provided by release of storage in the limestone around the linear feature, with that release constrained to the rock near the fracture or cavern wall and flowing essentially perpendicular to the fracture or cavern wall. After the quarter unit slope response, the drawdown stabilizes, indicating that release of groundwater from storage in the Madison aquifer has stopped and the pumped water is provided by vertical leakage into the Madison aquifer across a confining bed, as previously discussed.

The linear flow response in well M-3 means that the application of the Cooper-Jacob (1946) straight-line solution to the test data in 1992 was not appropriate and the transmissivity value of 4,400 gpd/ft is not defensible, although it is very similar to the values obtained in this report from the two wells with demonstrable radial flow response. It is also important to compare the results of the foregoing analysis demonstrating linear flow response to the Cooper-Jacob (1946) solution provided by Anderson & Kelly in 1980, based on the response of well M-1 to pumping at M-3, indicating a transmissivity value of 20,000 gpd/ft. Again, the latter value of aquifer transmissivity may not be defensible, considering the linear flow response at pumped well M-3. However, it must be recognized that observation wells penetrating the porous aquifer rock near a linear flow feature may give a radial flow response, so the Anderson & Kelly application of the Cooper-Jacob (1946) solution to the M-1 observation well response may have been valid. Unfortunately, the observation well data has not been recovered for this investigation and the large difference between the 1980 value of 20,000 gpd/ft and the values obtained herein of 2,726 and 3,088 gpd/ft at wells M-5 and M-6 cast considerable doubt on the validity of the 1980 interpretation resulting in transmissivity equal to 20,000 gpd/ft.

This does not mean that a transmissivity of 20,000 gpd/ft cannot be present in the Madison aquifer. The voids penetrated by wells M-2 and M-4 most offer extremely high transmissivity compared to the surrounding rock, even if they make up only a small percentage of the overall aquifer cross-sectional area. Although it is not known what percentage of the Madison aquifer consists of cavernous rock, the amount of groundwater that can be transmitted through such conduits is tremendous, as demonstrated by the 1,400-gpm yields of wells M-9 and M-10. Likewise, the hydraulic fracturing of well M-3 in 1992 evidently connected the well more directly to the cavernous openings in the limestone, resulting in a tremendous increase in specific capacity. J.M. Montgomery (1992) reported a 10-fold increase in the wells specific capacity. Figure 5-43 shows the post-fracturing constant rate test data.

Figure 5-43
Post-Hydrofracture Constant Rate Test of Well M-3 at 750 gpm



As shown on Figure 5-43, residual drawdown immediately recovered to 2.31 feet of drawdown in two minutes or less after 73 hours of continuous pumping of the well at 750 gpm. This indicates a well loss of 27.23 feet as compared to the pre-fracture well loss of 133 feet. The residual drawdown of 2.31 feet persisted for the first hour after pumping stopped and then measurements were discontinued. The residual drawdown of 2.31 feet is equivalent to 1.0 psi on the airline used to measure water levels in the well during the test. Since the finest subdivisions on precision pressure gages are generally not smaller than 1.0 psi, the residual drawdown of 2.31 feet is the best resolution the typical airline pressure gage can provide, a fact that suggests the residual drawdown may be the result of measurement limitations, not of actual drawdown.

If the residual drawdown is real or only an artifact of the measurement methods, the fact is that after hydraulic fracturing, well M-5 achieved stabilized drawdown in less than five minutes of pumping and exhibited 2.31 feet or less of actual aquifer water level drawdown (when corrected for well loss) thereafter. That drawdown recovered almost instantaneously when pumping ceased. Not only does this performance indicate an extremely high aquifer transmissivity in the cavernous part of the aquifer, but the high transmissivity allowed the "cone" of depression, i.e., the depressurized area, to expand nearly instantaneously throughout the cavernous system such that vertical leakage into the cavernous system equaled the pumping rate in a matter of minutes. Accordingly, drawdown of the water level in the aquifer outside the well was stabilized in a matter of minutes at or less than the resolution of measurement of 2.31 feet provided by an airline. Well loss increased the drawdown observed in the well to 29.54 to 32.77 feet during the 3-day test, with drawdown in the aquifer outside the well stabilized at 2.31 feet or less.

5.7.4 Madison Aquifer Summary

The methods of aquifer test interpretation presented above were not generally available or known by hydrogeologists in the United States at the time of the 1980 tests. The few analytical solutions published in the literature prior to 1980 (Gringarten and Ramey, 1975; Gringarten and others, 1975; and Gringarten and Witherspoon, 1972) were published in journals dealing with petroleum reservoir hydraulics or rock mechanics and were limited to plane fractures. Plane fractures do not offer significant storage and offer different hydraulic properties than the cavities in the Madison Limestone. Subsequent articles presented these concepts in hydrogeologic literature (Gringarten, 1982; Boehmer and Boonstra, 1986; for example); however, they have not enjoyed widespread use in the United States, primarily because of the practical difficulties in applying the math required for the solutions.

This has started to change with the advent of commercial software programs which allow hydrogeologists at a wide range of mathematical capabilities to apply the solutions efficiently and correctly with reasonable expenditure of time and money. However, the linear flow solutions currently available in the most widespread commercial software packages remain limited to solutions for plane fractures. Plane fractures are fractures with no storage (zero width), uniform drawdown throughout the length of the fracture (no hydraulic gradient in the linear flow path), and infinite or nearly infinite hydraulic conductivity (and therefore infinite transmissivity). Plane fracture conditions may occur in high-pressure petroleum reservoirs and perhaps in some aquifer systems; however, they are not appropriate for the type of linear flow observed in the Madison aquifer, i.e., linear flow offering substantial groundwater storage and hydraulic gradients. Therefore, solutions to the type of linear flow observed in the tests of the City of Gillette's so-called cavity wells remain academic unless a great deal of effort is expended to apply the solutions. This summary history of the technology should make it abundantly clear that the previous aquifer test interpretations were limited to the available technology and the reports about those interpretations state clearly that there were substantial problems recognized in applying that technology.

Processing the previous aquifer test data with the diagnostic and specialized plots for linear flow provides a new perspective on the hydraulics of the Madison aquifer in the vicinity of the City of Gillette's Madison well field. Radial flow response in two of the pumped wells provides values of transmissivity for the Madison aquifer rock mass, independent of open cavities or flow conduits. These values, 2,726 gpd/ft and 3,088 gpd/ft, indicate at least the local hydraulic properties of the limestone aquifer where interconnected porosity is so widespread as to support radial flow to the pumped wells. The radial flow responses show that 1,000 to 3,000 minutes of continuous pumping is required to expand the footprint of the depressurized area (cone of depression) in the Madison aquifer enough that vertical leakage into the depressurized area equals the pumping rate and drawdown in the Madison aquifer stabilizes.

The linear flow response obtained in the other aquifer tests in the City of Gillette's Madison well field indicates that the voids penetrated by some of the wells are not isolated voids in the limestone, but are instead well connected groundwater flow conduits through the aquifer. The upper part of the Madison limestone (Mission Canyon Formation) is well-known for its fossil cavern system (fossilized karst). Less well recognized are other internal caverns and autobreccia zones related to solution and collapse. Many caverns in the limestone may be joint-controlled, producing an interconnected network of solution cavities or long, linear conduits aligned with the discontinuities in the limestone rock. Enlargement of joints by solution of the limestone is often the first step in development of a cavern system that later progresses to more complex interconnections of multiple solution conduits.

When a well penetrating one of these conduits is pumped, the flow of groundwater to the well is through the path of least resistance, i.e., the solution cavity. This flow is hydraulically "linear" as compared to radial flow to the well through a homogeneously porous media. Accordingly, the aquifer test data exhibit classical linear flow responses on the diagnostic time-drawdown plots. However, it must be recognized that other types of groundwater flow occur simultaneously with linear flow to the pumped well.

The Madison aquifer at the City of Gillette's Madison well field is a confined aquifer. Confined groundwater storage exists because the groundwater in the cavities supports part of the lithostatic load of the aquifer rock and overlying overburden rock. Removal of groundwater through a pumped well reduces the confined water pressure in the cavities, thus transferring load from the groundwater to the rock matrix of the aquifer. The transfer of load to the rock causes the rock to consolidate or settle, thus reducing the volume of the pores and cavities in the rock. The water produced from the reduction in void volume as the rock compresses is the largest component of confined groundwater storage released to the pumped well. In addition, there is an infinitesimal expansion of the compressed water, offering additional small, but measurable release storage from the confined cavities in the limestone that is the second and smaller component of confined storage released to the pumped well.

In the early part of the pumping withdrawals from the confined Madison aquifer, all of the pumped water may be provided by release of confined aquifer storage in the cavities in the limestone. This aquifer response is represented on the diagnostic plot as the log-log straight line with a half unit slope. During this phase of the confined aquifer response, all of the groundwater flow takes place in the cavity penetrated by the well and all the release of groundwater storage is also from the cavity. This phase of flow is referred to as "linear" flow. This phase of the pumping may appear in a test, as in the 1992 test of well M-3, or it may be so fleeting in duration that it is not detected in the test measurements.

The next phase of aquifer response, which is often the first response detected during a test, as in wells M-2 and M-4, is one in which the flow of water to the pumped well is conveyed through the cavity, but the release of groundwater storage is from the pores in the rock containing the cavity. In this phase of the aquifer response, represented by a log-log straight line of quarter unit slope on the diagnostic plot, the cavity is no longer releasing confined groundwater storage. All of the pumped water is provided by storage released out of the pores in the surrounding rock. The storage is released into the cavity and then flows through the cavity to the pumped well. Thus, in addition to the flow through the cavity to the well, there is flow to the cavity through the rock containing the cavity. The latter flow into the cavity is generally perpendicular to the cavity walls, limited to a relatively narrow zone along the cavity, and therefore parallel flow into the cavity, not radial or pseudo-radial into the cavity. Accordingly, the linear flow through the rock into the cavity followed by linear flow through the cavity to the pumped well is referred to as "bilinear" flow. In bilinear flow, the rate of drawdown is controlled by the release of storage from the aquifer rock and the cavity is merely a conveyance conduit to the pumped well.

If bilinear flow can occur (due to the hydraulic properties of the rock hosting the cavity penetrated by the well), it follows that the expansion of a "cone of depression" or depressurized area away from the cavity with increased pumping duration will ultimately result in the depressurized area becoming large with respect to the length of the depressurized "linear" cavity. When this occurs, over a period of transition, the depressurized area becomes essentially a somewhat elongated or oval cone of depression around the depressurized part of the linear cavity with curved groundwater flow lines converging on the cavity. This phase of aquifer response is referred to as "pseudo-radial" flow and can be analyzed with the Cooper-Jacob (1946) semilogarithmic straight-line solution. This phase of aquifer response was not observed in the City of Gillette Madison wells because vertical leakage into the depressurized areas around the pumped wells stabilized drawdown before pseudo-radial flow could commence.

The stabilization of drawdown points out an extremely significant function of the open cavities in the limestone regarding the Madison aquifer's response to pumping. The caverns or solution enlarged joints in the limestone provide high-capacity flow paths directly to wells that penetrate the cavities. The high formation pressure (water pressure) in the cavities allows a pumped well to generate tremendous differential pressure to drive groundwater through the cavities. For example, if the top of a cavity is at a depth of 2,400 feet and the static water level in the well is 500 feet, the formation pressure in the cavity is approximately 823 psi, assuming the specific gravity of the groundwater is 1.0. Because the cavities are open conduits, only a fraction of the available formation pressure must be used as drawdown in the pumped wells to obtain large flows of groundwater through the cavities to the pumped wells.

More importantly, the more open the cavity and the better the interconnection between the cavity and cross-cutting solution cavities, the less resistance to flow through the cavity. In well-connected cavities with little resistance to flow, not only is the drawdown in the pumped well small, the decrease in pressure along the cavity when the well is pumped expands along the cavity nearly instantaneously as a hydraulic response. The faster the decrease in pressure expands out away from the well, the sooner the vertical leakage increases to equal the pumping rate and stabilize drawdown in the cavity. This effect is evident in the tests of the City of Gillette's Madison wells.

In the wells that did not penetrate cavities and which exhibited radial flow to the wells, 1,000 to 3,000 minutes of pumping time was required before drawdown stabilized (Figures 5-30 and 5-32). At well M-2 which did penetrate a cavity, stabilization occurred after approximately 150 minutes of pumping (Figure 5-39) Similarly, well M-4 exhibited stabilized drawdown after approximately 450 minutes of pumping (Figure 5-37) from a cavity in the limestone. Well M-3, which as not recorded as directly penetrating a cavity, exhibited strong linear flow response including linear and bilinear flow, exhibited stabilized drawdown after 700 minutes of pumping. After well M-3 was hydraulically fractured to interconnect it more effectively with the open cavities in the limestone, stabilized drawdown occurred essentially instantaneously with 2.31 feet of drawdown on less. This indicates that at the 750-gpm pumping rate, propagation of the depressurized area throughout the cavity system to the extent required to generate enough vertical leakage to stabilize drawdown was essentially instantaneous at a very small amount of drawdown.

The open cavities in the Madison limestone also have an influence on water quality, particularly on the fluoride concentrations that are a regulated constituent in public drinking water supplies. In general, fluoride and several other dissolved constituents in the groundwater are at smaller concentrations in the groundwater from the cavities and at higher concentrations in groundwater from the porous, non-cavity portions of the aquifer. This is thought to be a simple matter of groundwater flow rates and time for dissolution of minerals in the aquifer mineral matrix to take place with longer contact times for dissolution to occur existing in the groundwater flow through the porous rock than in the open cavities. These conditions were documented by J.M.

Montgomery (1992) in their report about pre-frac and post-frac conditions in the Madison aquifer during the hydraulic fracturing stimulation of well M-3. Their summary conclusions about the water quality changes are as follows:

> "***As shown in Table 1 in the Introduction section of this report, there were some key differences in water quality in the cavity versus the non-cavity wells. The major differences occurred in the concentrations of sodium, magnesium, chloride, sulfate, fluoride and hardness. The pre-stimulation water quality analyses indicated that the concentration of all of these constituents had dropped since the well as initially drilled. The sodium, magnesium, and fluoride levels had dropped to concentrations found in water from the cavity wells. Whereas, the chloride concentration remained relatively unchanged. The evolution of the quality of water from Well M-3 as compared to the water quality from cavity wells is illustrated on Figure 9, using fluoride, sodium and chloride for typical examples. Differences also exist in sulfate, magnesium and hardness. Concentrations of all these parameters, as shown in Table 4, were initially significantly greater in the non-cavity wells.***
>
> ***The change in water quality toward that of the cavity wells, may be due to Well M-3 deriving a larger proportion of its water from the cavity system in the area with time. The changes in concentrations of chloride, sulfate, fluoride and hardness indicate that the well stimulation increased connection of the well to the cavity system."*** (J.M. Montgomery, 1992).

Figure 5-44 is a reproduction of Table 3 from J.M. Montgomery (1992) showing laboratory analytical reports for water quality from well M-3 prior to and after hydraulic fracturing. Figure 5-45 is a reproduction of Table 4 from J.M. Montgomery (1992) summarizing selected water quality constituent concentrations in cavity and non-cavity wells.

Figure 5-43
Table 3 from J.M. Montgomery (1992)

Table 3 Summary of Water Quality Analyses of Groundwater Samples from Well M-3

	Pre-Stimulation 10/31/91	Post-Stimulation 12/9/91	
Microbiological			**Units**
Total Coliform	5 neg.	5 neg.	
Turbidity	0.21	0.58	N.T.U.
Inorganic Chemical			
Arsenic	<0.005	<0.005	mg/L
Baruim	<0.01	<0.01	mg/L
Cadmium	0.002	<0.001	mg/L
Chromium	<0.02	<0.02	mg/L
Lead	<0.01	0.02	mg/L
Mercury	<0.001	<0.001	mg/L
Nitrate	0.32	0.26	mg/L
Selenium	<0.005	<0.005	mg/L
Silver	<0.005	<0.005	mg/L
Fluoride	0.97	0.75	mg/L
Radiological			
Gross Alpha	<1	1.5 +/- 1.3	pCi/L
Gross Beta	<1	<1	pCi/L
Raduim 226	0.3 +/- 0.2	<0.2	pCi/L
Radium 228	<1	3.5 +/- 0.6	pCi/L
Natural Uranium	0.0017	0.009	mg/L
Secondary Standards			
pH	6.8	6.9	Std. Units
Chloride	9	4	mg/L
Copper	<0.01	0.03	mg/L
Foaming Agents	<1	<1	mg/L
Sulfate	292	264	mg/L
Total Dissolved Solids	668	590	mg/L
Specific Conductance	824	864	umhos/cm @ 25 degrees C
Zinc	<0.01	<0.01	mg/L
Color	0	0	color units
Corrosivity (Langlier Ind	(-0.36)	(-0.24)	
Iron	0.17	0.25	mg/L
Manganese	<0.02	<0.02	mg/L
Odor	none detected	none detected	
Sodium	7	7	mg/L
Calcium	130	124	mg/L
Magnesium	41	40	mg/L
Potassium	3	2	mg/L
Carbonate	0	0	mg/L
Bicarbonate	250	268	mg/L
Boron	<0.1	<0.1	mg/L
Silica	16.2	14.1	mg/L
Hardness as CaCO3	493	474	mg/L
Total Alkalinity	205	220	mg/L
Total Acidity	0	0	mg/L
Volatile Organic Compounds	none detected	none detected	mg/L
Pesticides/Herbicides	none detected	none detected	mg/L

Figure 5-44
Table 4 from J.M. Montgomery (1992)

Table 4 Gillette Madison Well Field Water Quality

Parameter	Non-Cavity Well Average	M-3 Initially (1980)	M-3 Pre-Stimulation (10/91)	M-3 Post-Stimulation (12/91)	Cavity Well Average
Sodium (mg/L)	11.73	11.4	7	7	5.35
Magnesium (mg/L)	52.53	51.1	41	40	41.5
Chloride (mg/L)	11.23	10	9	4	2
Sulfate (mg/L)	333.75	333	292	264	241.5
Fluoride (mg/L)	2.25	1.91	0.97	0.75	0.76
Hardness (mg/L as CaCO3)	521	532	493	474	475

Non-Cavity Well Average from Wells M-5, M-6, M-7 and M-8.
Cavity Well Average from Wells M-2 and M-4.

5.8 Madison-Fort Union Aquifer Comparison

Table 5-1 provides a comparison of the various characteristics of the Madison and Fort Union aquifers that influence the long-term sustainable development potential of both aquifers.

This review the Fort Union aquifer characteristics indicates that there is a significant groundwater resource available for development in the Fort Union aquifer. However, the discontinuous nature of the multiple sandstone lenses in the Fort Union aquifer, the fact they are embedded in much finer grained sediments that are generally considered barriers to groundwater flow, and the resultant hydraulic boundaries imposed on groundwater flow to wells penetrating those lenses all impose significant limitations on sustainable groundwater development from this aquifer. This does not mean the aquifer should not be used, but when compared to the sustainable development potential of the Madison aquifer, the Fort Union aquifer offers considerable uncertainty about the life of sustainable development whereas comparatively little uncertainty exists about the life of sustainable development in the Madison aquifer.

In contrast to the Fort Union aquifer which exhibits internal boundaries to groundwater flow, the Madison aquifer offers high-pressure conduits of tremendous flow potential for individual wells and receives vertical leakage from contiguous strata. As demonstrated by the aquifer test data reviewed in this report, withdrawal of groundwater from storage in the Fort Union aquifer followed by an equal duration of recovery does not result in full recovery of the groundwater levels at the pumped well, due to the limited extent of the individual sandstone lenses providing the pumped water. Residual drawdown remaining after recovery from pumping becomes cumulative with repeated pump/no-pump cycles, due to the boundaries of the sandstone lenses, causing long-term decline in groundwater levels at a rate greater than predicted from the hydraulic properties of the aquifer without such limiting boundaries.

Vertical leakage into the Madison aquifer provides essentially all of the groundwater pumped from wells connected to the open cavities in the limestone. This means no significant long-term release of groundwater from storage in the Madison aquifer is required to supply the water pumped from the well field, therefore, use of Madison wells will not result in residual drawdown or induce a long-term decline in groundwater levels in the Madison aquifer. The Madison aquifer, if properly developed and managed, will be a reliable source of groundwater for the City of Gillette well beyond the foreseeable future; whereas the life of sustainable development of groundwater from the Fort Union aquifer is predictably finite and will likely experience limitations within the foreseeable future with respect to satisfying the projected demands for municipal water supply for the City of Gillette.

TABLE 5-1

Comparison of Madison and Fort Union Aquifer Characteristics

	Fort Union Aquifer	Madison Aquifer
Potential Well Yields	75-140 gpm in core[1/] wells with traditional well construction. 250 gpm ± in new wells with modern water-well construction.	550-600 gpm [2/] in porous limestone with radial flow and no stimulation. 1,400 gpm or more in open cavities with or without stimulation by hydraulic fracturing (hydro-frac).
Aquifer Transmissivity		
Porous Rock	2,709 gpd/ft based on reinterpretation of test of ASR test well.	2,726-3,088 gpd/ft based on wells M-6 and M-5, respectively without hydraulic fracturing and with radial flow response.
Open Cavities	No open cavity flow recognized in Fort Union aquifer.	Nearly infinite in open caverns with linear flow, with or without stimulation by hydraulic fracturing.
Boundary Conditions	No-flow or negative boundary imposed by limited extent and lack of interconnection between sandstone lenses producing water to wells. Negative boundary conditions cause increased rate of groundwater level decline in response to long-term pumping.	Positive boundary provided by vertical leakage into cone of depression replaces groundwater pumped from storage in aquifer. Vertical leakage from adjacent formation greatly reduces groundwater level decline in response to long-term pumping.
Long-Term Drawdown	Significant long-term drawdown in response to pumping at rates required to supply future demand for City of Gillette will result in declining well yields within foreseeable future.	Small amounts of drawdown will occur in response to pumping at rates and annual volumes required to supply future demand for City of Gillette far beyond foreseeable future; therefore long-term well yields are anticipated to be reliable.
Water Quality[3/]	Core[1/] wells: Average TDS = 341 mg/L Average Fluoride = 2.12 mg/L Non-Core wells: Average TDS = 586 mg/L Average Fluoride = 2.70 mg/L	Cavity and fractured wells: Average TDS = 601 mg/L Average Fluoride = 0.65 mg/L Non-cavity wells: Average TDS = 687 mg/L Average Fluoride = 1.52 mg/L

[1/] **Wester-Wetstein (2004; Table 4-1)**
[2/] **Wester-Wetstein (2004; Table 4-3)**
[3/] **Wester-Wetstein (2004; Table 4-4)**

Water quality differences between the Fort Union and Madison aquifers include hardness, total dissolved solids (total dissolved mineral content), and fluoride concentrations. The Fort Union groundwater is very soft whereas the Madison groundwater is very hard. Although this is not a public health and welfare issue, it is an aesthetic issue which has caused some complaints from the public about hard water from the Madison aquifer. The latter complaints were not necessarily about the hardness per se, but more about change from soft water when only the Fort Union wells were in use to hard water when Madison wells were put on line during high demands. A related issue was the contrast in hardness between parts of the Gillette distribution system receiving mostly Fort Union water and those parts receiving primarily Madison water. Some public concern about this issue will probably continue into the future; however, year-round use of Madison water in much of the system has reduced public complaints about hardness to a large extent, indicating it was the seasonal change in the hardness that was an issue more than the hardness itself.

Table 5-2 shows total dissolved solids (TDS) and fluoride concentrations for Fort Union and Madison aquifer wells. The data for the Fort Union aquifer are divided into two categories of wells – those originally referred to as "core wells" (Wester-Wetstein, 2004) that were in production year round and other wells used to supplement production during periods of high demand. In the Wester-Wetstein (2004) report, the core wells are S-9, S-18, S-19, S-22, S-24 and S-26 whereas wells S-17, S-21, S-23, S-25 and S-27 were used during periods of high demand. As shown on Table 2, the core wells offer much smaller fluoride concentrations than the other Fort Union wells. Accordingly, the non-core Fort Union Wells used for high demand periods generally are not operated unless their production can be blended with water from the Madison aquifer (or from the core Fort Union wells) to achieve an end water quality that does not exceed the MCL for fluoride.

The Madison aquifer wells are also divided into two categories – cavity wells and non-cavity wells. Fluoride concentrations in Madison groundwater from the cavity wells are less than half of those in water from the non-cavity wells and TDS concentrations are likewise somewhat less in water from cavity wells than in water from non-cavity wells. The data in Table 5-2 are summarized from Wester-Wetstein (2004; Table 4-4) and were current as of November 8, 2004, purportedly representing the chemical properties of the groundwater under historical operation of the sampled wells.

TABLE 5-2
Summary of TDS and Fluoride Concentrations in Fort Union and Madison Groundwater

Fort Union Core Wells			Madison Cavity Wells – High Capacity		
Well No.	TDS (mg/L)	Fluoride (mg/L)	Well No.	TDS (mg/L)	Fluoride (mg/L)
S-9	346	1.10	M-2	624	0.66
S-18	322	1.30	M-3	590	0.68
S-19	262	1.10	M-4	608	0.64
S-22	312	5.40	M-9	581	0.63
S-24	528	2.40	M-10	601	0.63
S-26	274	1.40			
Averages:	**341**	**2.12**	**Averages:**	**601**	**0.65**
Fort Union Non-Core Wells			**Madison Non-Cavity Wells**		
Well No.	TDS (mg/L)	Fluoride (mg/L)	Well No.	TDS (mg/L)	Fluoride (mg/L)
S-12	359	1.37	M-1	672	1.10
S-17	875	3.90	M-5	678	2.03
S-20	422	1.40	M-6	714	1.69
S-21	655	3.00	M-7	645	1.57
S-23	490	2.40	M-8	724	1.22
S-25	746	4.00			
S-27	558	2.80			
Averages:	**586**	**2.70**	**Averages:**	**687**	**1.52**

The data on Table 5-2 show that Madison aquifer wells that are well-connected to cavity flow, either by fortuitous penetration of large, highly transmissive caverns, or by hydraulic fracturing; offer significantly lower fluoride concentrations than any of the other wells, including non-cavity Madison wells. The foregoing conclusion remains true even if the two highest concentrations of fluoride in the core Fort Union wells are disregarded.

5.9 Groundwater Hydrographs

Hydrographs of groundwater level fluctuations reveal significant differences between trends in the Fort Union aquifer and those in the Madison aquifer. This is a factor to take into account when comparing the Fort Union aquifer to the Madison aquifer as a potential source of water supply for the future. Undoubtedly part of the differences in the trends is due to the large difference between ongoing groundwater abstractions out of the Fort Union aquifer compared to those from the Madison aquifer. The Fort Union aquifer has been subjected to growing use for public water supply as well as coal mining and CBM pumping for a number of years whereas the Madison aquifer between the Black Hills and the City of Gillette has been subjected to only limited municipal water supply use. However, the differences in the aquifer fluctuations are rooted in deeper causes than differences in historic use of the groundwater.

The hydrographs show long-term fluctuations in the Madison aquifer in response to recharge. As discussed previously in this report, the Fort Union aquifer does not appear to exhibit response to on-going natural recharge and discharge, except in localized subareas. Investigations conducted since the start of coal mining in the basin in the late 1960's and early 1970's have failed to detect measurable natural recharge of discharge below the uppermost part of the Tongue River Member, based on surface water flow statistics, groundwater hydrographs, and groundwater radioisotope data. Investigators of the Fort Union aquifer in the Powder River Basin have therefore concluded that very little, if any, recharge and natural discharge occurs in the Tongue River Member of the Fort Union Formation in the Powder River Basin, including the area around Gillette. This being the case, it is not expected that the hydrographs of the Fort Union aquifer would exhibit seasonal or long-term response to recharge as apparent in the hydrographs of the Madison aquifer groundwater levels.

However, many of the groundwater level hydrographs from the Fort Union aquifer in the Gillette area exhibit significant downward trends in groundwater levels. The decline of groundwater levels often starts well before the advent of CBM pumping and occurs in parts of the Tongue River Member and deeper parts of the Fort Union well below the depth of CBM pumping. The timing of such declines indicates they are related to municipal and community use of groundwater from the Fort Union aquifer, rather than from CBM pumping. Therefore, the groundwater hydrographs from the Tongue River Member and deeper parts of the Fort Union aquifer near Gillette that exhibit downward trends provide evidence of the removal of groundwater from storage in the aquifer, i.e., groundwater mining, largely by use of the groundwater for domestic drinking water and related residential and municipal landscape irrigation. Thus, the hydrographs document discharge of groundwater from the Fort Union aquifer, albeit through wells rather than natural discharge.

Some of the Fort Union aquifer hydrographs show recovering groundwater levels related to taking nearby wells out of service, but they largely resume declining after a few years of recovery. This should not be surprising in view of the documentation of little or no recharge to this part of the regional aquifer system in the Fort Union Formation. However, the hydrographs are significant in that they show the aquifer is responding to pumping in the manner that is predicted by the absence of significant natural recharge.

Likewise, the groundwater hydrographs from the Madison aquifer are significant in that they document recharge to the regional Madison aquifer system. Not only do they indicate recharge takes place, they show relatively synchronous fluctuations in regional groundwater levels in response to fluctuations in recharge. The difficulties and potential error associated with estimates of recharge to the Madison aquifer were discussed previously. Whatever the actual amount of short-term and long-term recharge to the Madison aquifer might be, the size of the area between the recharge areas is large and the groundwater fluctuations throughout that area are significant. Therefore, the relatively synchronous groundwater level fluctuations in the Madison aquifer throughout the region suggests a highly transmissive aquifer system through which the hydraulic response to changes in groundwater levels in the recharge areas is communicated rapidly throughout the system.

Groundwater level changes of 25 to 100 feet in the Madison aquifer throughout the same region likewise argue for a significant amount of groundwater volume related to the changes in storage reflected on the hydrographs. Therefore, the amount of natural recharge (or natural discharge when groundwater levels decline) must likewise be relatively large, although difficult to measure. These concepts are consistent with the aquifer test information from the City of Gillette's Madison aquifer well field that reveal large, open conduits transmitting groundwater through the limestone aquifer under confined conditions with relatively high formation pressure.

Therefore, the groundwater hydrographs for the Madison aquifer reflect a relatively high transmissivity aquifer through which changes in groundwater storage are rapidly communicated and which receives significant recharge. Groundwater hydrographs for the Fort Union aquifer reflect a relatively low transmissivity aquifer in which changes in groundwater storage are localized and are essentially all downward except for transient recovery of groundwater levels when localized pumping ceases. Such recovery is provided by redistribution of groundwater storage at the expense of nearby groundwater levels which must decline to provide recovery in the areas of deepest drawdown.

The latter responses fit the classic definition of groundwater mining to support sustainable development, subject to the local hydraulic limitations of the aquifer. Although the same definition may apply equally to development of groundwater from the Madison aquifer, the Madison aquifer offers two advantages that are absent in the Fort Union aquifer. One advantage is that the Madison aquifer receives vertical leakage, probably from the overlying Minnelusa Formation, when its confined aquifer portions are depressurized by pumping wells. Given sufficient pumping duration, the vertical leakage equals the pumping discharge rate from the Madison aquifer and mining of groundwater and associated decline of groundwater levels ceases, even though pumping continues. This offsets much of the effects of groundwater mining in the Madison aquifer. The second advantage of the Madison aquifer over the Fort Union aquifer is its highly transmissive nature which allows groundwater to be released from storage over a wide area (as compared to the Fort Union aquifer) with relatively small impact on groundwater levels compared to the much greater impact of the same amount of pumping in the Fort Union aquifer, even if vertical leakage did not offset pumping withdrawals in the Madison aquifer.

The hydrograph data were generously provided by the Wyoming State Engineer's Office (SEO) and consists of provisional and unpublished data collected at a number of dedicated groundwater level monitoring wells. These data are not available through the SEO's website and were abstracted from the SEO's databases by Jeremy Manley, Groundwater Management Specialist, Groundwater Division, and converted from the SEO's proprietary database format to Microsoft Access files so that they could be used for this study. Figure 5-45 shows the locations of the monitoring wells by aquifer.

Figure 5-45
Locations Of Groundwater Monitoring Wells

Figure 5-46
Monitoring well locations and hydrograph lines of section near Gillette

5.9.1 Wasatch Aquifer Hydrographs

Figure 5-47 shows two hydrographs for the Wasatch aquifer. Hydrograph RCH-2B is located near Pine Buttes and reflects seasonal pumping and/or recharge cycles superimposed over an approximately 14-year groundwater fluctuation of zero net change in storage in a shallow aquifer. The net change in groundwater storage indicated on a hydrograph is equal to zero between points of identical groundwater levels on the hydrograph.

Hydrograph H13 is located inside the Gillette city area and exhibits recovery of the groundwater level from the beginning of the record in 1991 to the beginning of 2003 after which the trend is essentially no groundwater level change. It can be speculated that the observed recovery of 12 feet over about 12 years is related to the City ceasing use of its Wasatch aquifer wells in 1985.

Figure 5-48 shows hydrographs from wells completed in the Wasatch aquifer and/or the uppermost Fort Union aquifer coal beds where CBM pumping is taking place. The result of depressurizing the aquifer to stimulate methane production is obvious.

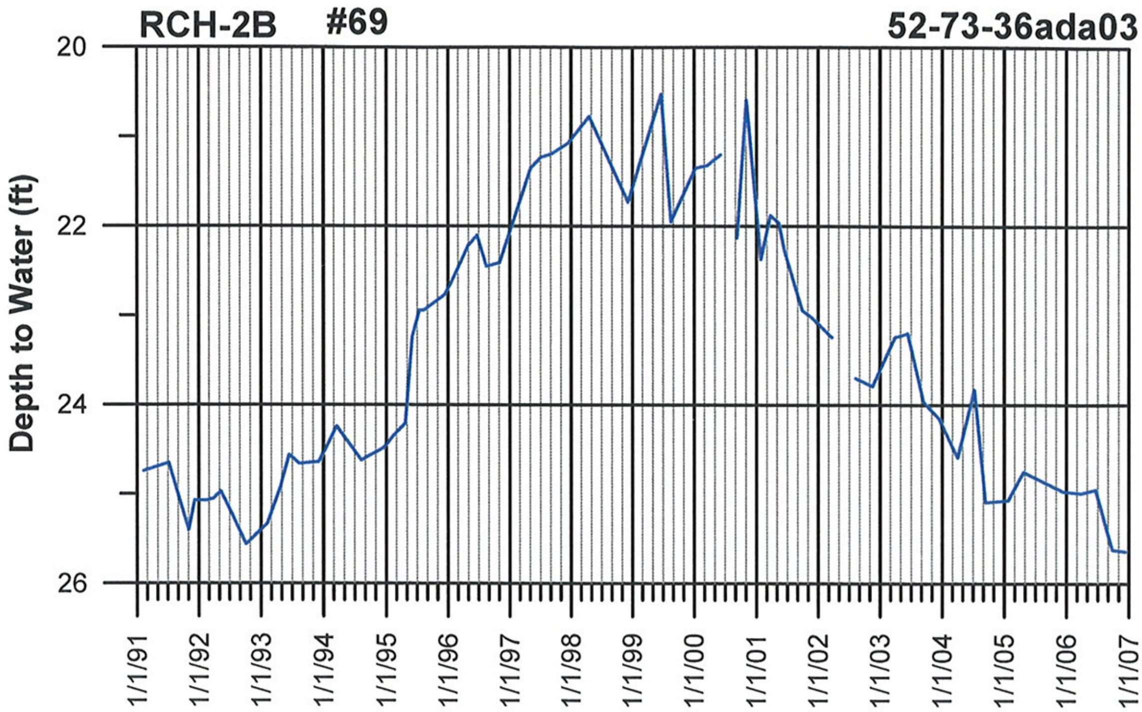
5.9.2 Fort Union Aquifer Hydrographs

The Fort Union Aquifer hydrographs shown on Figures 5-49a through 5-49d generally depict downward groundwater level trends with seasonal pumping superimposed over the trends. Each page of hydrographs relates to one of the lines of section shown on Figure 5-46. Each hydrograph is obviously sensitive to the pumping operations of nearby wells and some hydrographs appear to reflect only seasonal pumping cycle fluctuations compared to the general downward trend in other wells.

5.9.3 Madison Aquifer Hydrographs

Madison aquifer hydrographs are shown on Figure 5-50. The locations for the Madison wells are shown on Figure 5-45. The hydrograph located nearest the Black Hills, Stateln-1, exhibits a gradual 10-foot decline in groundwater levels in 1991 through 1998

Figure 5-47
Wasatch Aquifer Hydrographs



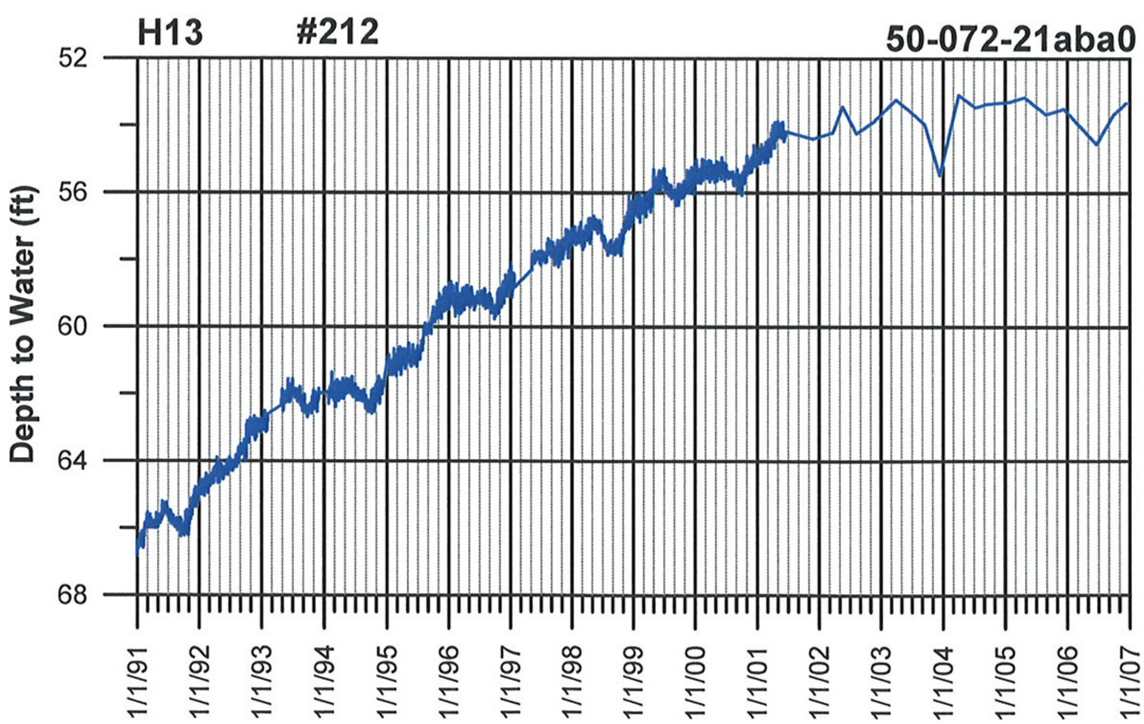


Figure 5-48
Wasatch aquifer and/or Fort Union coal bed hydrographs

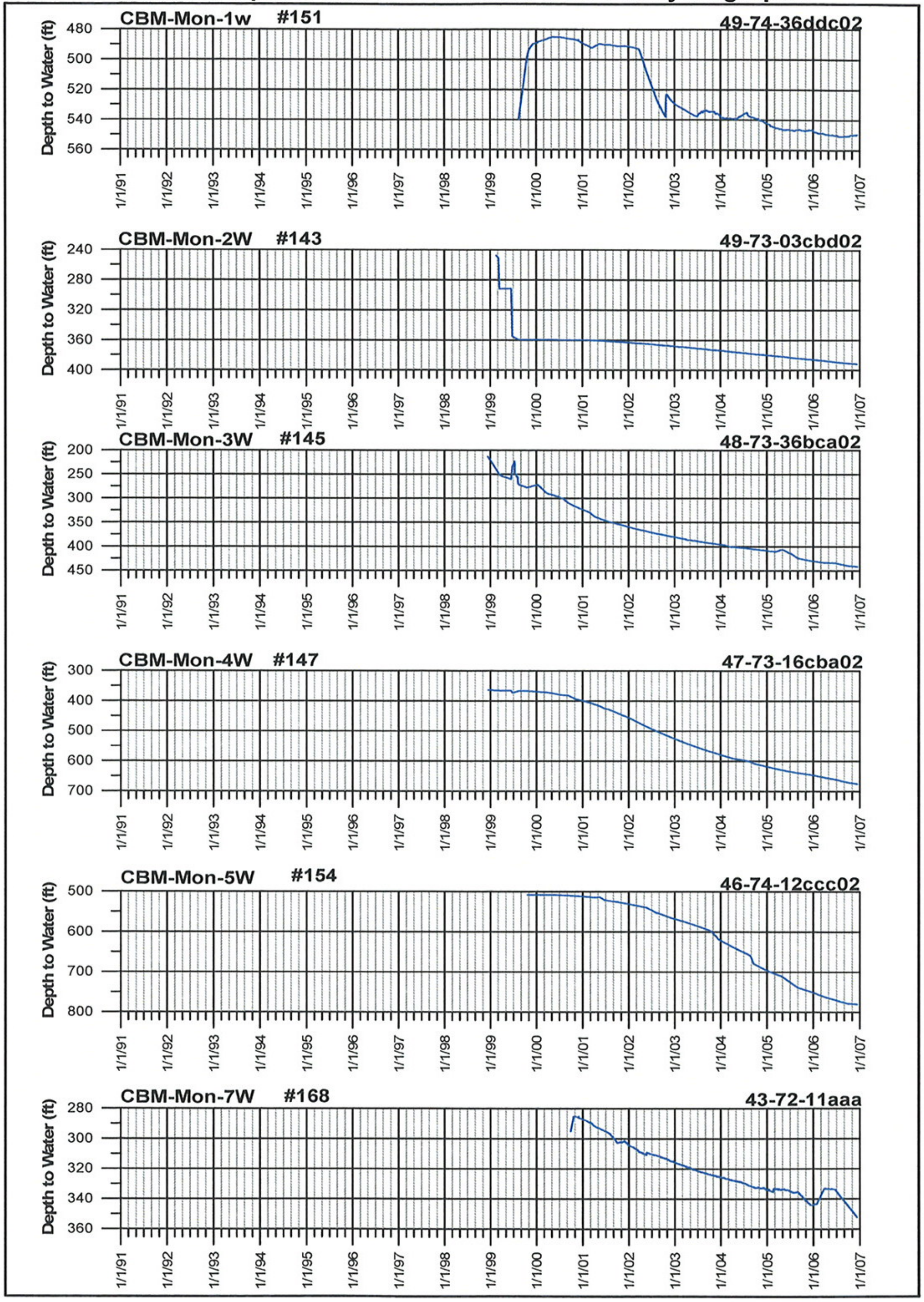


Figure 5-49a
Fort Union aquifer hydrographs from line of section A-A

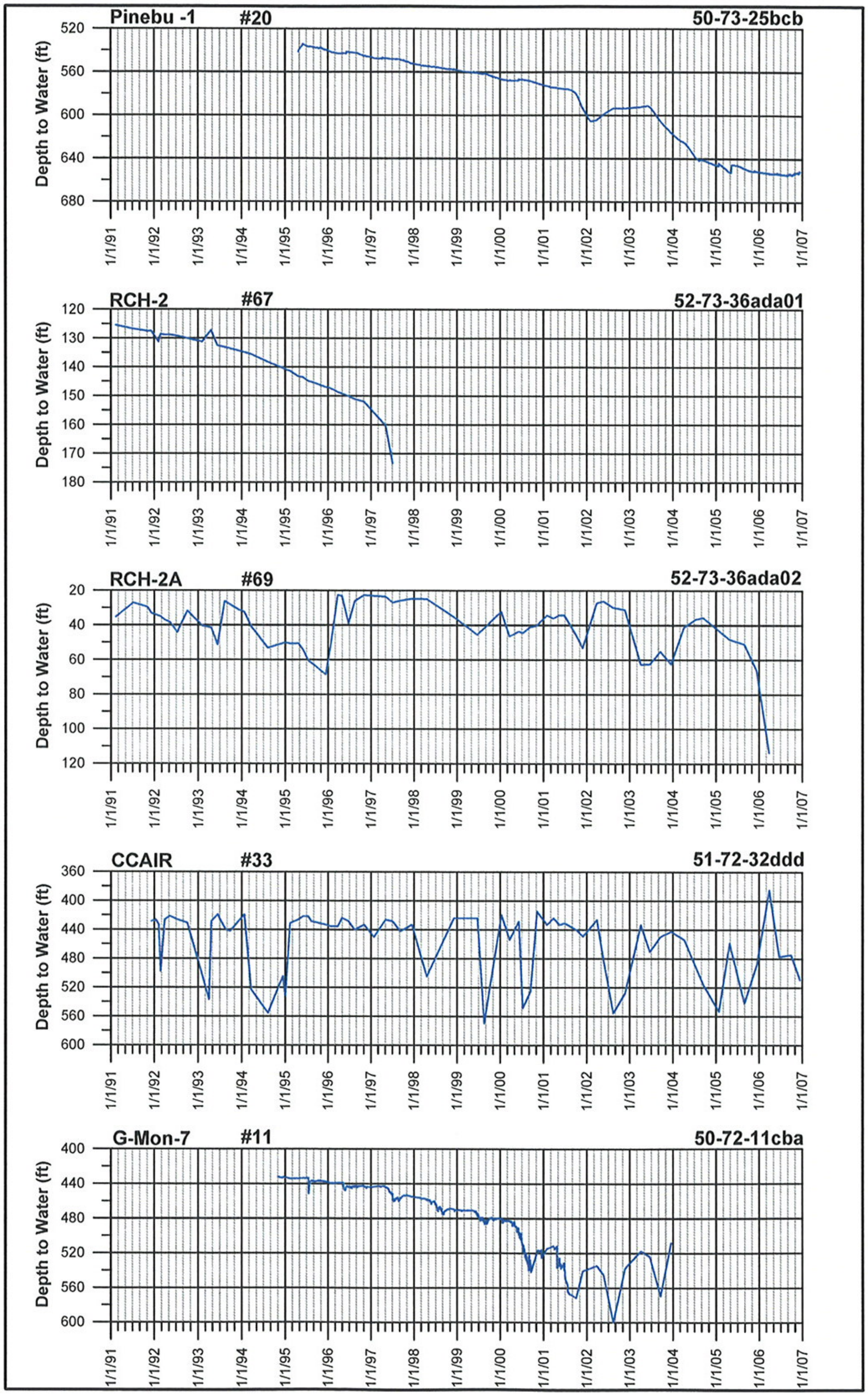


Figure 5-49b

Fort Union aquifer hydrographs from line of section B-B

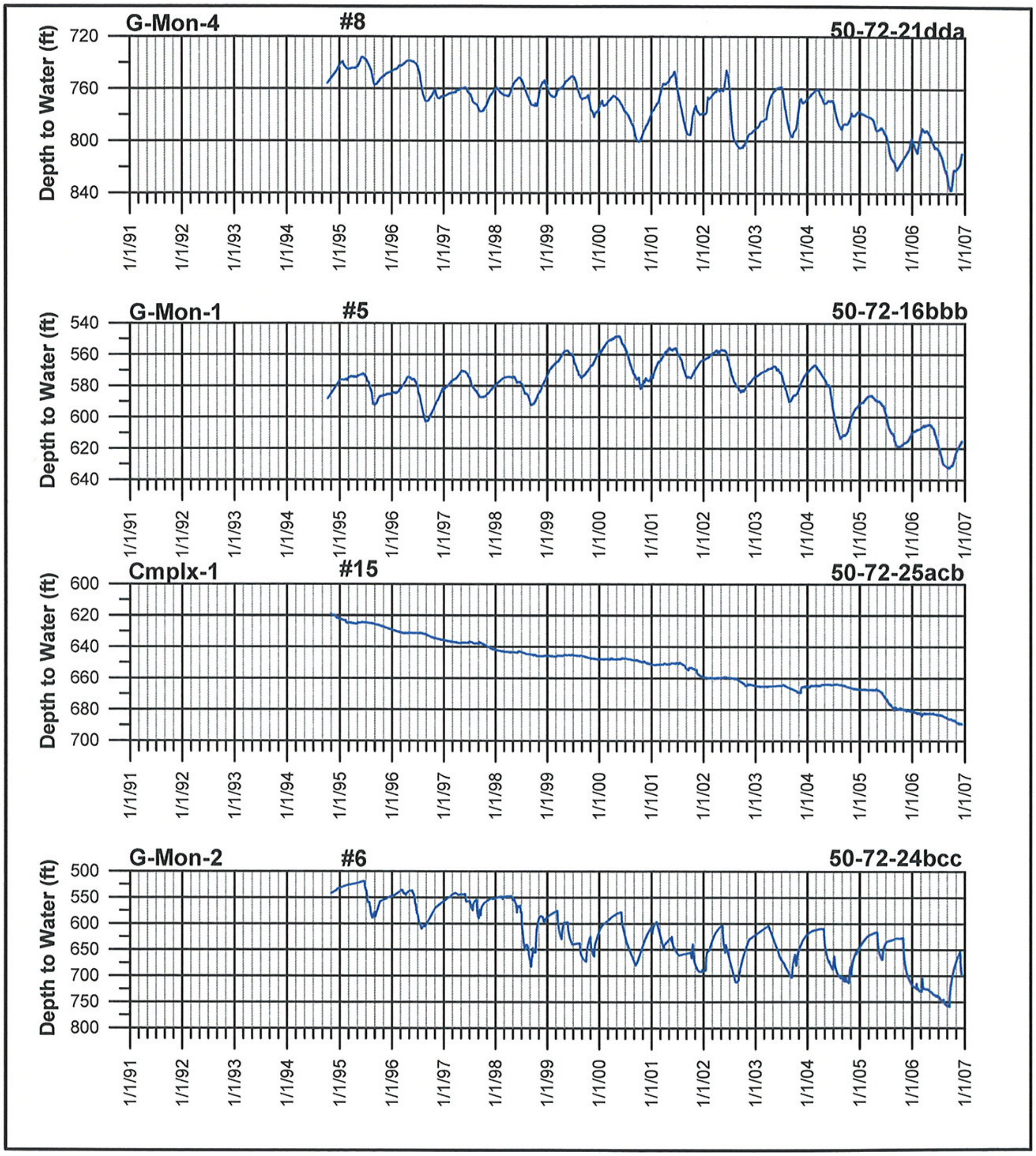


Figure 5-49c
Fort Union aquifer hydrographs from line of section C-C

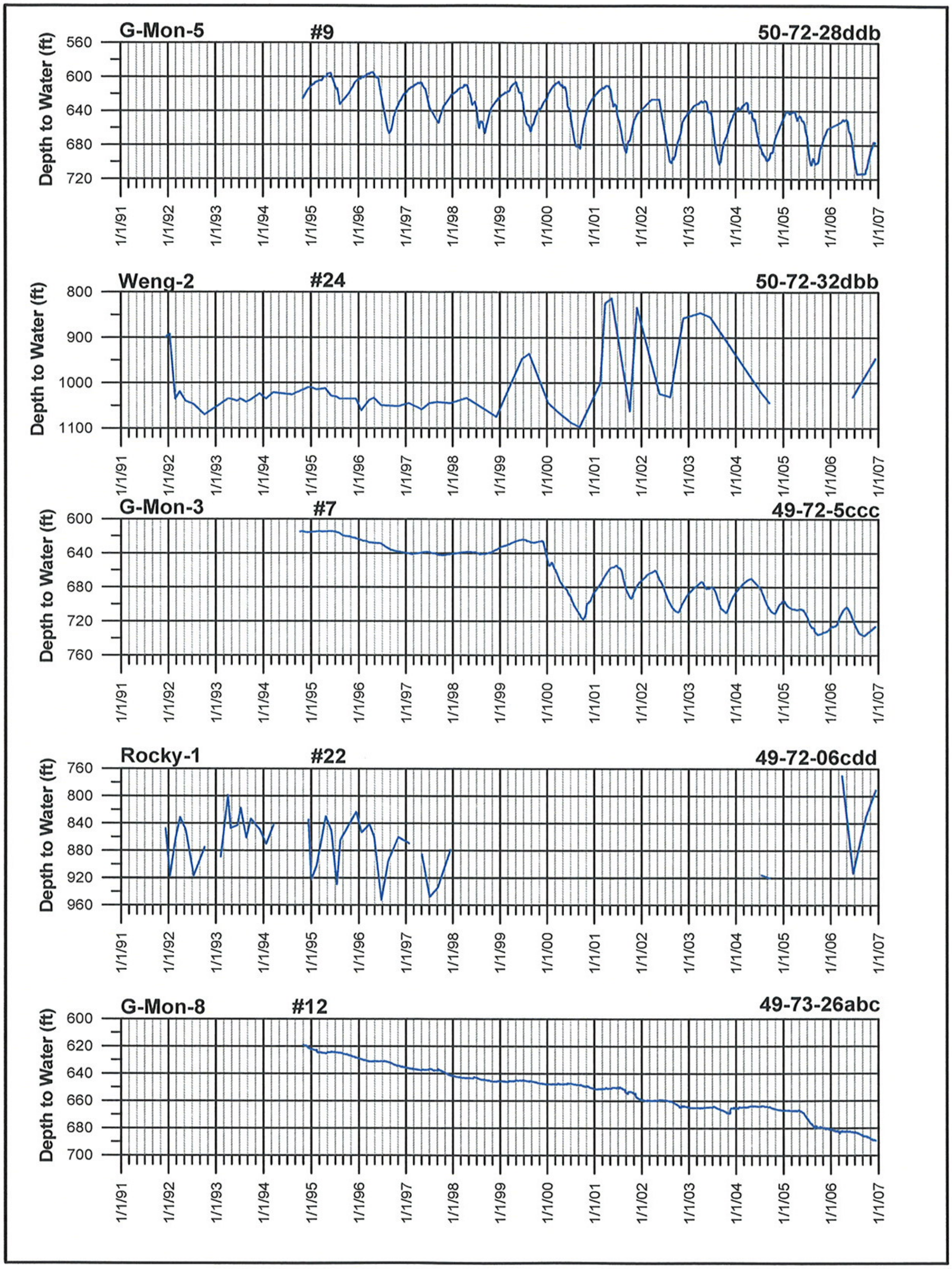


Figure 5-49d
Fort Union aquifer hydrographs from line of section D-D

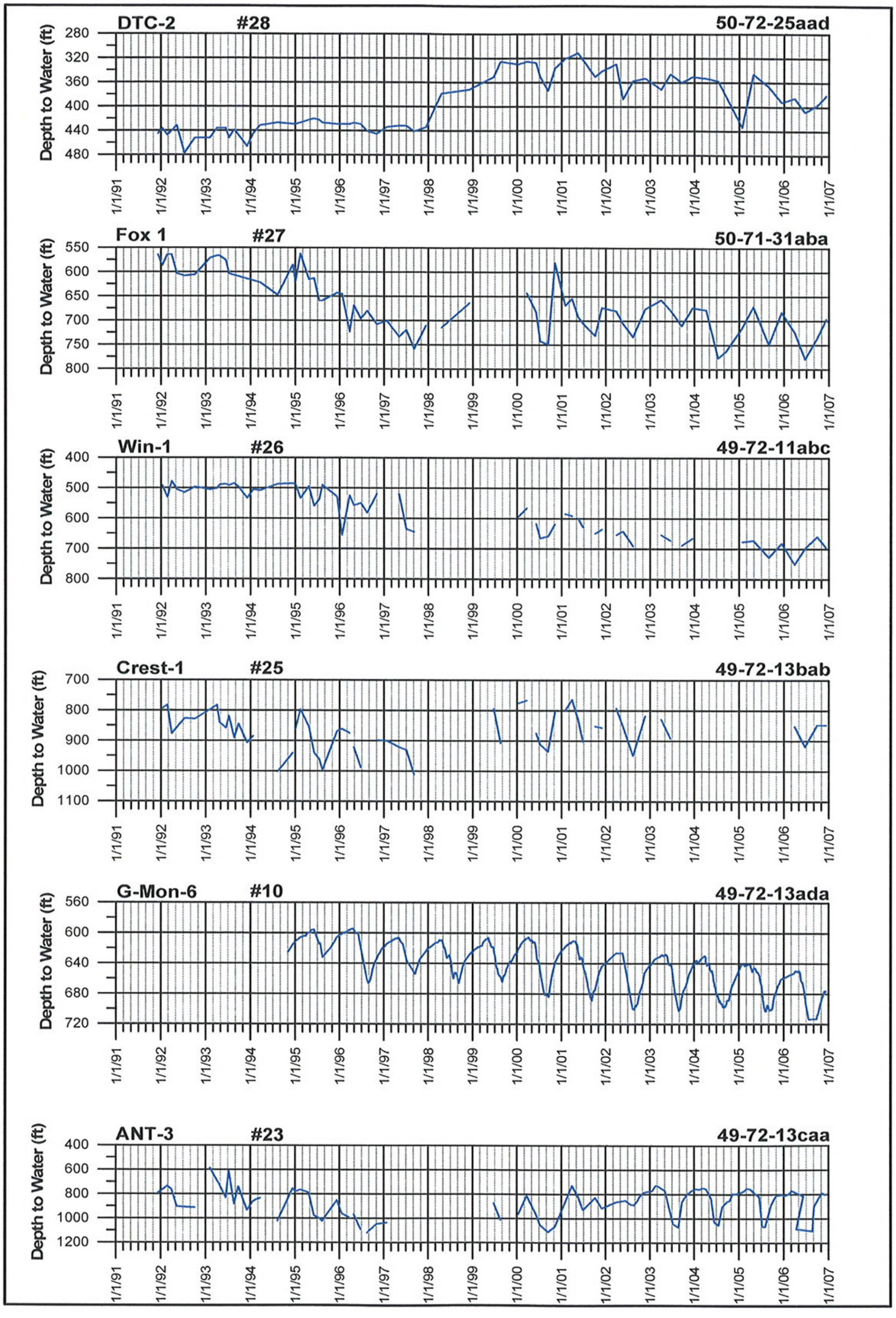
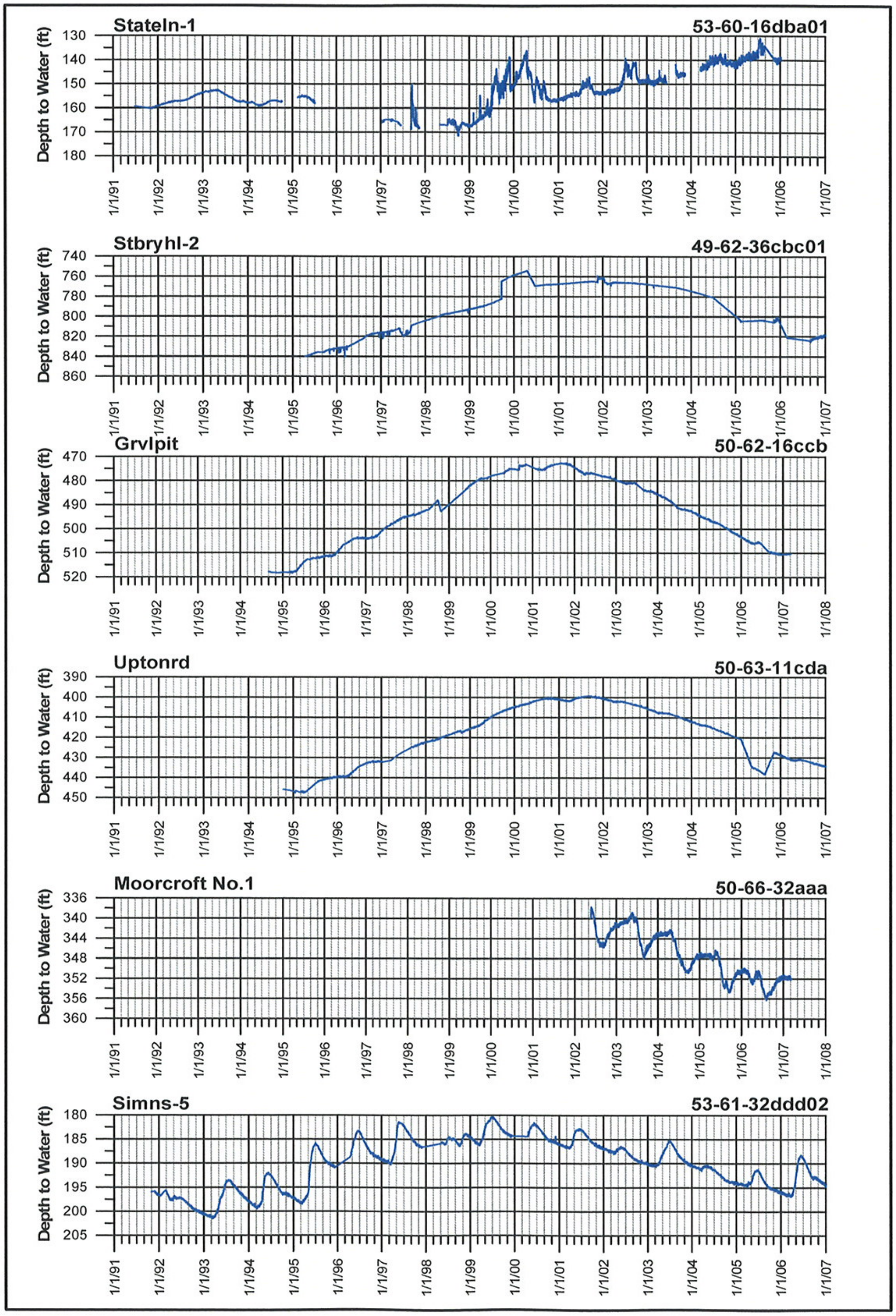


Figure 5-50
Madison aquifer hydrographs



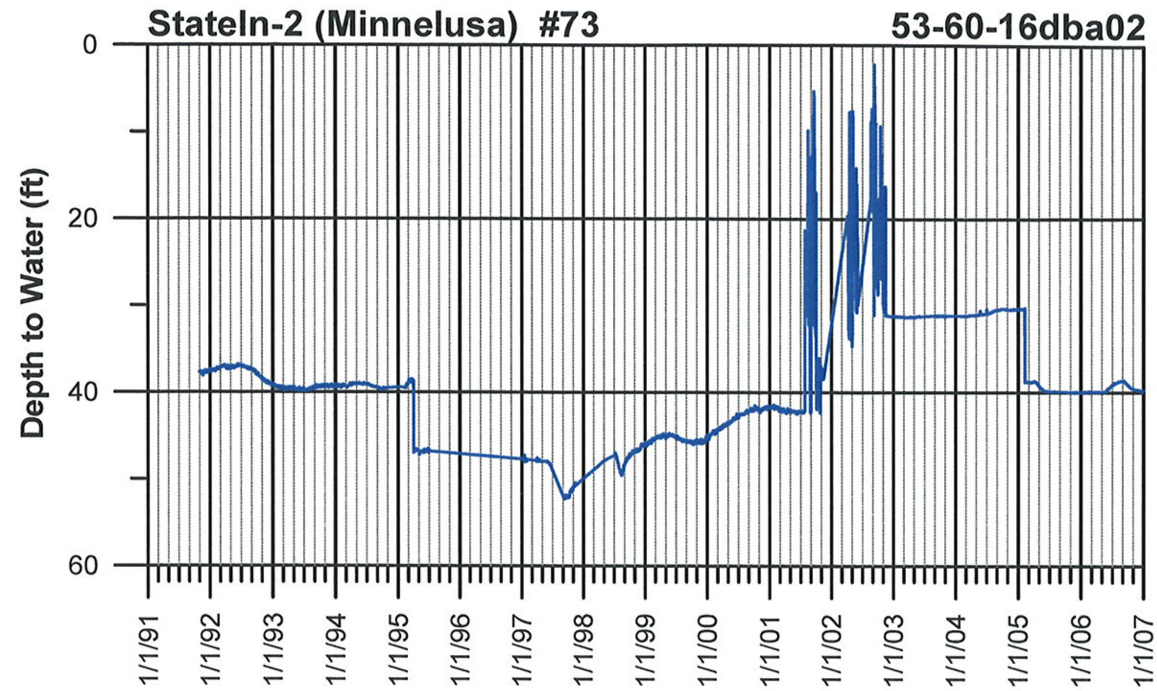
followed by nearly 40 feet of groundwater level rise through the end of 2005. The remaining hydrographs, including a short period of record for the new municipal well for Moorcroft, exhibit approximately 20 to 80 feet of groundwater level rise from about 1991 through the early part of 2000 followed by a similar decline in groundwater levels through the remainder of the record. This long-term fluctuation results in 10 to 20 feet of storage above zero net storage change across the monitored region at the end of the period of record in all of the hydrographs. Seasonal pumping cycles are superimposed over the long-term fluctuation at the Moorcroft and Simns-5 locations. These hydrographs show that the current level of pumping abstractions from this part of the Madison aquifer system do not equal and are much less than the long-term natural recharge and discharge as well as less than the long-term fluctuations in natural recharge and discharge.

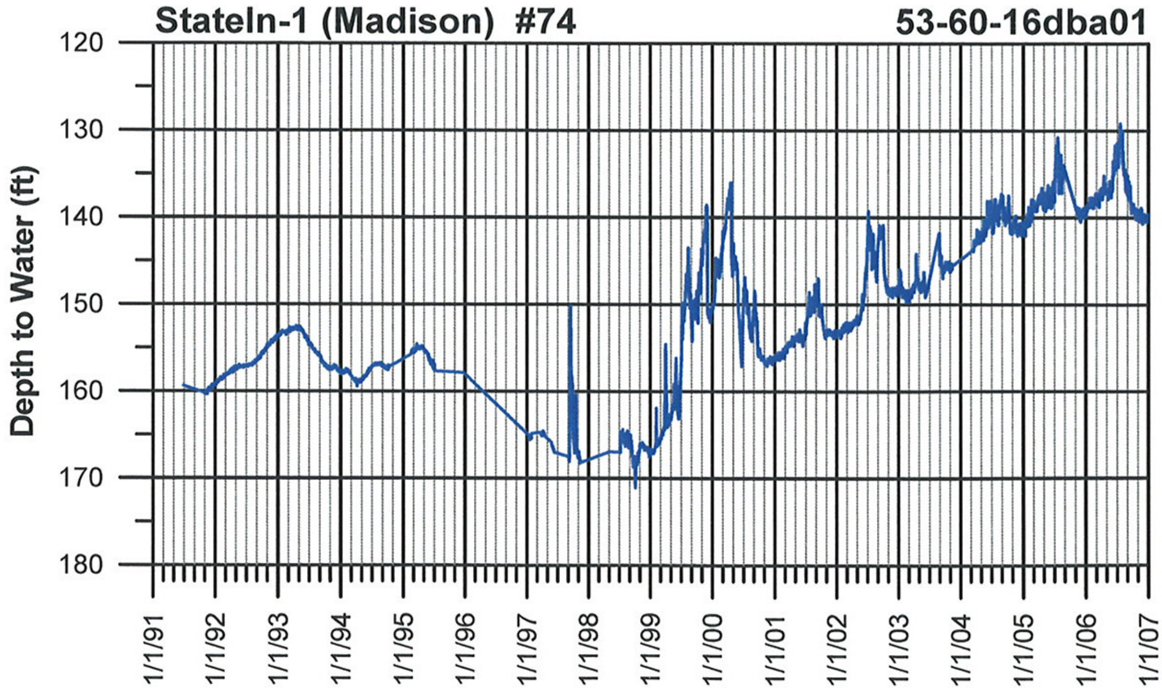
5.9.4 Madison-Minnelusa Hydrographs

Figures 5-51, 5-52 and 5-53 show hydrographs from pairs of wells completed in the Minnelusa and Madison aquifers at three different locations (Figure 5-46). The pairs of hydrographs compare the groundwater level fluctuations in the two aquifers.

The first pair of hydrographs, Figure 5-51, is from the easternmost location, near the state line between Wyoming and South Dakota, approximately 2.5 miles north northeast of Beulah, Wyoming. The hydrographs show a vertical gradient from the Minnelusa Formation to the Madison Formation depth to water in the Minnelusa aquifer averaging about 40 feet compared to 130 to more than 165 feet in the Madison aquifer. The trends of the groundwater fluctuations are the same in both aquifers; however, the two aquifers do not exhibit identical response to recharge events. As documented in Carter and others (2001) and Driscoll and others (2002), recharge to the Minnelusa and Madison aquifers is from a combination of incident precipitation and surface water flow loss into the outcrop areas of the formations. The Madison aquifer hydrograph shows groundwater levels rising in 1999 and 2000 in response to recharge events that cause corresponding, but small response in the Minnelusa whereas large responses to recharge occur in the Minnelusa in 2001 and 2002 that correspond to significant, but slightly smaller fluctuations in the Madison. The hydrographs show the Minnelusa and Figure 5-51: Minnelusa-Madison hydrographs at T53N, R60W, Section 16.

Figure 5-51
Minnelusa-Madison hydrographs at T53N, R60W, Section 16





Madison aquifers respond to the same recharge events, but in different ways, indicating differences recharge received from the same recharge events and differences in aquifer storage. Although the downward vertical gradient from the Minnelusa to the Madison aquifer indicates the potential for interformational leakage from the Minnelusa into the Madison, the hydrographs indicate that there is no direct hydraulic communication between the two aquifers and that whatever interformational leakage takes place must flow across an intervening confining layer.

Figure 5-52 compares Minnelusa and Madison aquifer hydrographs at a location along Highway 14 west of Beulah and approximately 6 to 7 miles west of the monitoring well site depicted on Figure 5-51. The hydrographs at the site depicted on Figure 5-52 show good correlation between the long-term groundwater level trends in the two aquifers and the magnitude of fluctuation is approximately the same in both aquifers. However, there is no direct correlation between individual events on the two hydrographs. The Madison aquifer hydrograph exhibits what appear to be seasonal recharge fluctuations that do not appear in the Minnelusa aquifer hydrograph, again indicating no strong hydraulic communication between the two aquifers and differences in how the two aquifers respond to the same recharge events. The vertical hydraulic gradient between the two hydrographs is upward from the Madison into the Minnelusa at this location.

The hydrographs on Figure 5-53 are from a site approximately one-quarter mile northeast of the flank of Strawberry Mountain and approximately 16 miles south southeast of Sundance, Wyoming, located on the west flank of the Black Hills. As in the other Minnelusa-Madison hydrograph pairs, there is good correlation of groundwater level trends between the two hydrographs; however, the magnitude of fluctuation in the Minnelusa aquifer is approximately seven feet compared to approximately 80 feet in the Madison aquifer. The vertical gradient between the two aquifer is downward from the Minnelusa into the Madison; however, lack of good correlation between individual short-term fluctuations in the two aquifers indicates there is no direct hydraulic communication between the two aquifers and any interformational leakage that occurs must pass through a confining layer separating the two aquifers.

Figure 5-52
Minnelusa-Madison hydrographs at T53N, R61W, Section 32