CHAPTER 5
GROUND WATER SOURCE EVALUATION

5.1 BACKGROUND

The City of Gillette is located in the Powder River structural basin of northeastern Wyoming. The term Powder River Basin as used herein refers to the structural basin, not to the surface water drainage basin of the same name.

Figure 5-1 is a generalized geologic map showing the axis of the Powder River Basin and the distribution of the geologic units where they are exposed at the land surface. Figure 5-2 is a generalized stratigraphic column of the above strata through the base of the Fox Hills Formation. Figure 5-3 is a highly generalized geologic cross section from west to east across the Powder River Basin in the vicinity of Gillette. The cross section shows the relationships between the various geologic units; however, on a greatly exaggerated vertical scale. Only the uppermost 3,000 feet or less of the strata in the vicinity of Gillette are of interest to this study.

The deepest aquifer, the Madison Limestone, is not shown on the above maps and cross sections due to its much greater depth in the vicinity of Gillette. That portion of the Madison Limestone of interest to this investigation is 35 to 40 miles east of Gillette, on the east flank of the basin, where the top of the Madison Limestone is generally less than 2,400 feet below the land surface and the total depths of wells producing from the Madison aquifer less than 3,000 feet.

Previous reports prepared for the City of Gillette by their consultants have described the local details and history of wells completed in the various groundwater sources listed above and will not be repeated herein. Other literature about the geology of the Powder River Basin is extensive. A brief description of the geologic framework is quoted below from USGS Scientific Investigations Report 2005-5008, Hinaman (2005), for an overview of the information pertinent to the distribution and availability of groundwater from the various strata.

Figure 5-1
Generalized Geologic Map of the Powder River Structural Basin

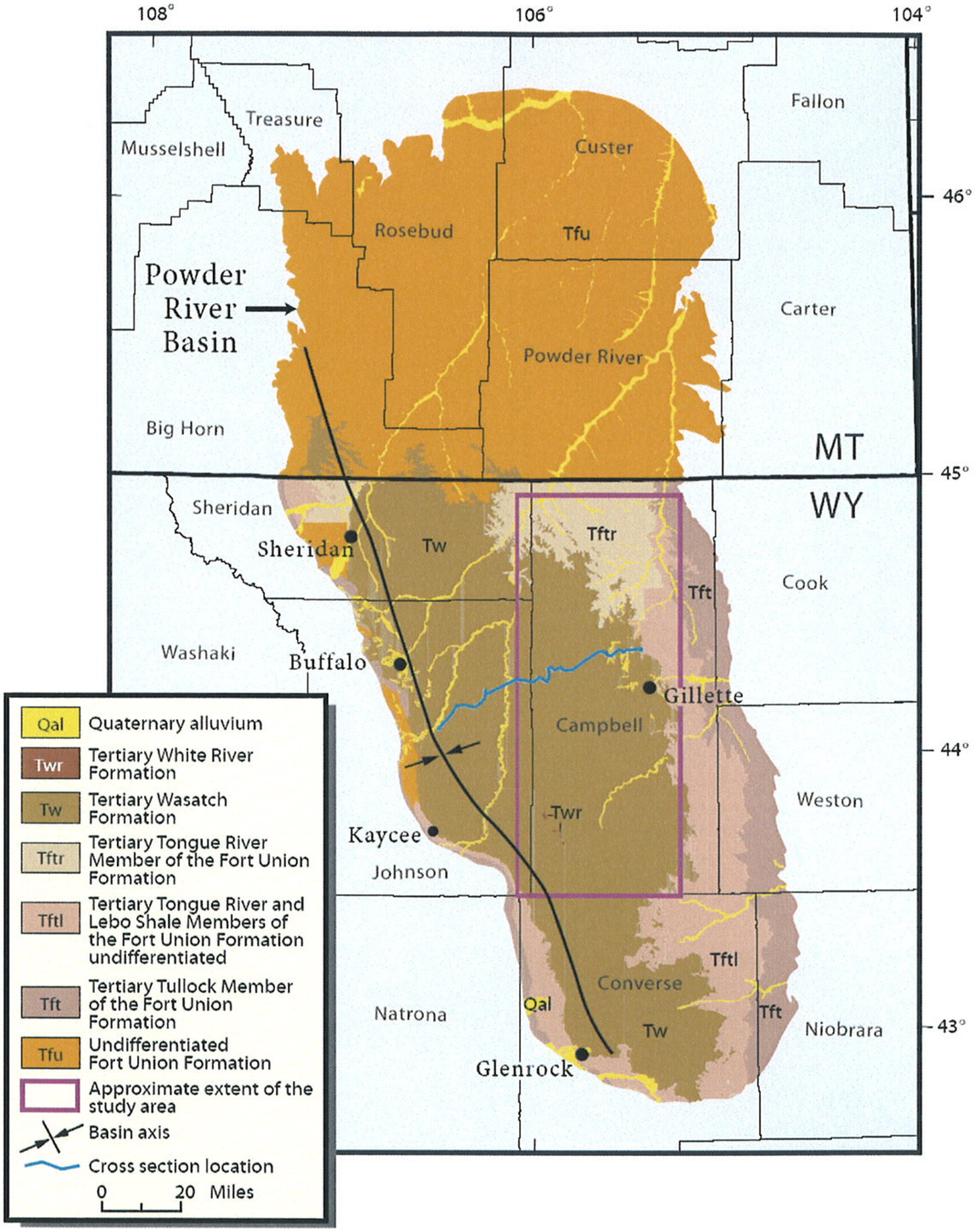


Figure 1. Generalized geologic map of the Powder River Basin, Wyoming and Montana showing the basin axis, counties, major cities, location of cross section (fig. 2), and approximate extent of the study area (modified from Flores and Bader, 1999).

Figure 5-1 reproduced from:
USGS Open-File Report 00-372

Figure 5-2
Generalized Stratigraphic Column Of Tertiary and Upper Cretaceous Formations Including Hydrogeologic Subdivisions

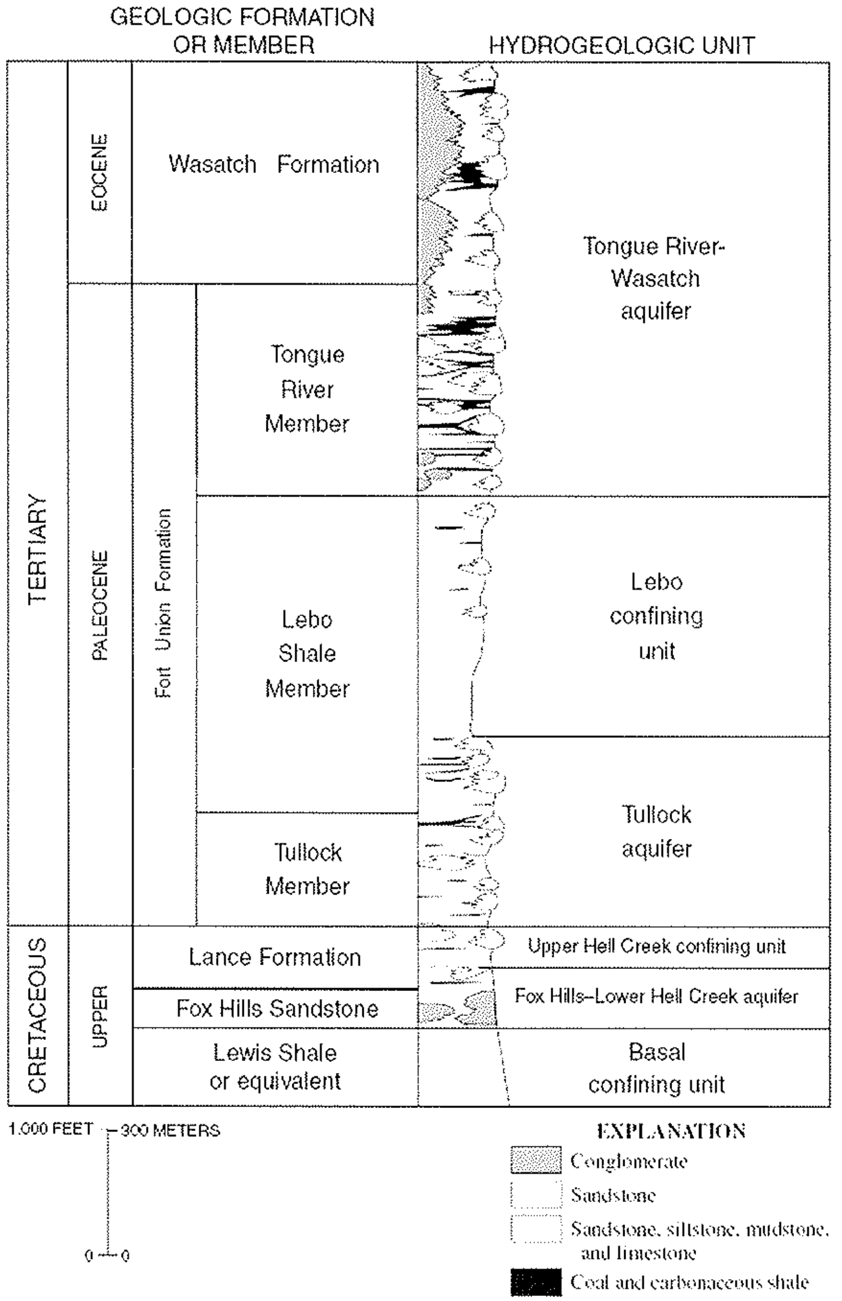


Figure 2. Generalized stratigraphic column of Tertiary and upper Cretaceous formations in the southern Powder River structural basin, Wyoming (modified from Flores and Bader, 1999).

Figure 5-2 reproduced from:
USGS Scientific Investigations Report 2005-5008

Figure 5-3
Generalized Hydrogeologic Section Of The Hydrogeologic Units Of The Powder River Structural Basin

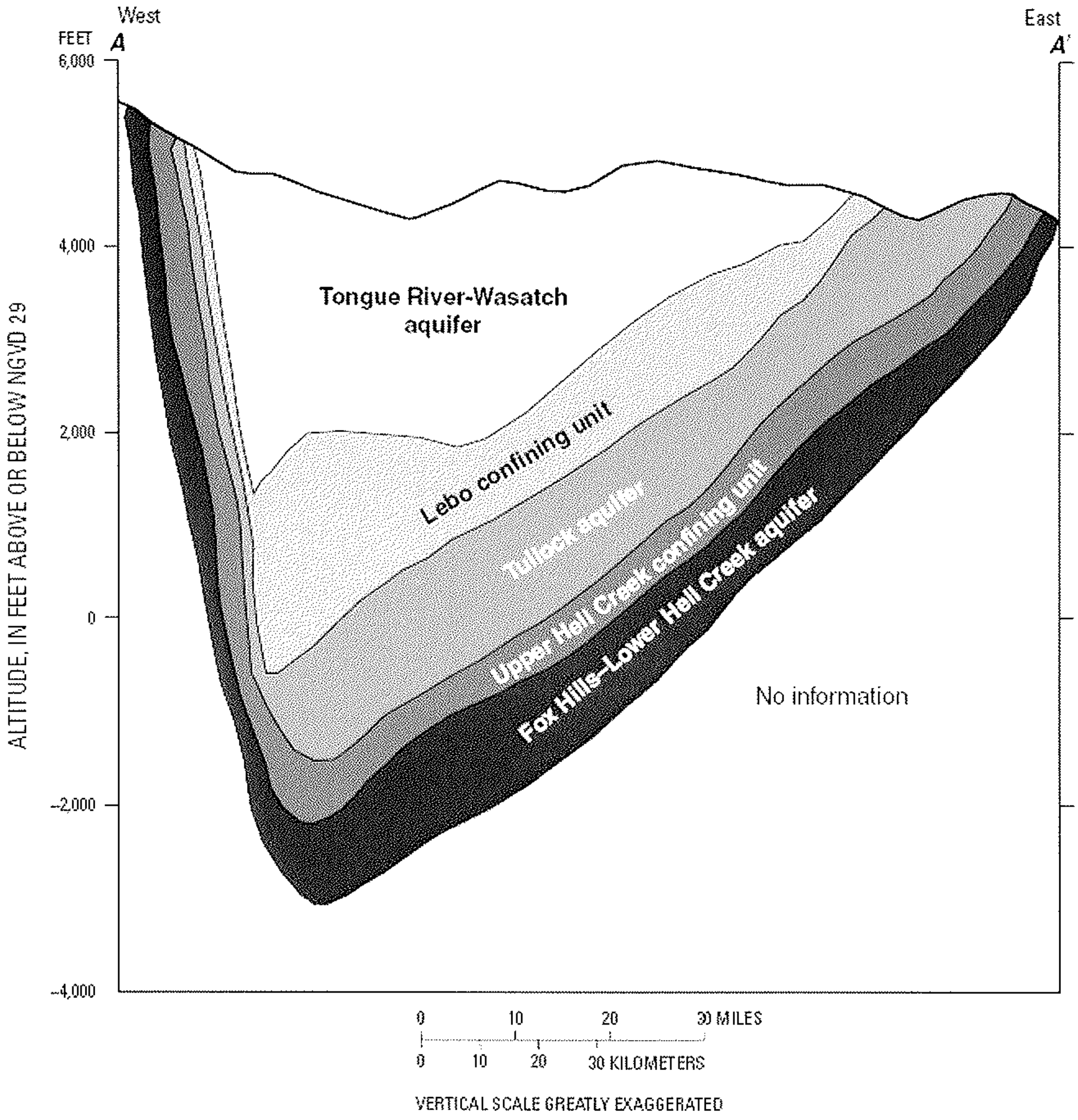


Figure 3. West to east hydrogeologic section of the Tertiary and upper Cretaceous hydrogeologic units showing the asymmetric nature of the Powder River structural basin, Wyoming.

Figure 5-3 reproduced from:
USGS Scientific Investigations
Report 2005-5008

"*. . . The Powder River Basin is an asymmetrical syncline formed during the Laramide orogeny (Late Cretaceous to early Tertiary age). The axis of the basin trends from southeast to northwest near the western margin of the basin* [Figure 5-1], *far from its geographic center. In Wyoming, the Powder River Basin is bounded by the Black Hills uplift in the north east, the Hartville uplift in the southeast, the Laramie Mountains in the south, the Casper arch in the southwest, and the Bighorn Mountains in the west . . . The basin continues northward into Montana, where another structural feature, the Cedar Ridge anticline, separates it from the Williston Basin . . . The entire basin covers an area of more than 22,000 square miles (Flores and others, 1999).*

Tertiary and upper Cretaceous hydrogeologic units within the Powder River Basin in Wyoming are the focus of this study. The Tertiary and upper Cretaceous formations that contain the hydrogeologic units of interest are described first in this section of the report.

The Lewis Shale and its equivalents [Figure 5-2], *which are upper Cretaceous-age marine units (Steidtmann, 1993; Merewether, 1996), form the basal confining layer. Overlying this shale is the upper Cretaceous-age Fox Hills Sandstone, which represents a sandy shoreline as the seaway retreated to the north east (Steidtmann, 1993; Mereweather, 1996), and the Lance Formation of which about one-third is composed of channel sandstones and the rest of the formation is finergrained interfluvial sedimentary rock (Conner, 1992). Crossbedding, channel orientation, and orientation of well-cemented sandstones having log-like forms show deposition from eastward-flowing streams on both sides of the Bighorn Mountains and on the east side of the Powder River Basin, indicating that those mountains had yet to emerge, and that the Powder River Basin had yet to form by the end of the Cretaceous period (Conner, 1992).*

The lowest Tertiary unit is the Paleocene-age Fort Union Formation, which is subdivided into three members—the Tullock, Lebo Shale, and Tongue River Members [Figure 5-2]. *The Tullock Member is composed of about one-third channel sands and two-thirds finer grained overbank deposits (Brown, 1993). In the northwest part of the Powder River Basin, the Tullock Member contains carbonate clasts in its lowermost part, which are interpreted to mean that the Bighorn Mountains were starting to emerge during the time of its deposition (Brown, 1993). At the same time, coal-forming environments were forming and being deposited from the southeast to the northwest (Brown, 1993). The Lebo Shale Member is mainly a mudstone with some minor channel sands (Whipkey and others, 1991). The depositional environment for the Lebo Member was described as lacustrine by Ayers (1986), whereas others (summarized in Flores and Bader, 1999) indicate that fine-grained sediments are from the shedding of fine-grained marine Cretaceous shales from the rising Bighorn Mountains. The Tongue River Member contains considerable channel sandstones and coals. The Tongue River Member transitions to the Eocene-age Wasatch Formation* *with no distinct marker or lithologic change at the contact between the two units (Bartos and Ogle, 2002).*

Some of the hydrogeologic units described in this report correspond with geologic formations, whereas other hydrogeologic units combine geologic formations, and other hydrogeologic units have contacts within stratigraphic units [Figures 2 and 3]. *This study uses terms for hydrogeologic units defined by Lewis and Hotchkiss (1981) and further described by Hotchkiss and Levings (1986). The uppermost hydrogeologic unit is the Tongue River-Wasatch aquifer, which averages 55 percent sand (Lewis and Hotchkiss, 1981). Below the Tongue River-Wasatch aquifer is the Lebo confining unit, which averages 31 percent sand (Lewis and Hotchkiss, 1981); however, predominant mudstones give it confining unit characteristics. Below the Lebo confining unit is the Tullock aquifer, which includes basal sands of the Lebo Shale Member. This aquifer averages 53 percent sand (Lewis and Hotchkiss, 1981) and is confined at its base by the upper Hell Creek confining unit. This unit typically acts as a confining unit, but with sand content ranging from 9 to 88 percent, and a mean of 35 percent, wells completed in it can flow (Hotchkiss and Levings, 1986). This confining unit is the upper part of the Lance Formation in Wyoming. Below this unit is the basal aquifer of interest in this report, the Fox Hills-Lower Hell Creek aquifer. Average sand content is 50 percent and it is a good source of water in the Powder River Basin (Hotchkiss and Levings, 1986). This aquifer includes the lower part of the Lance Formation and the Fox Hills Sandstone. At its base is the basal confining unit, the Lewis Shale, which is equivalent to the Bearpaw Shale in Hotchkiss and Levings (1986).*

For all of the hydrogeologic units of interest, the general direction of ground-water flow is from south to north with the influence of the major rivers as discharge areas (Hotchkiss and Levings, 1986; Whitehead, 1996). In parts of the basin, this general direction of flow is influenced by a flow component from topographically high areas, such as the Bighorn Mountains to the west, Laramie Range to the south, and Black Hills to the east. An example is shown on the potentiometric map of the Wyodak-Anderson coal bed by Daddow (1986), which only shows the eastern limb of the flow system and has a more southeast to northwest flow pattern. From a more regional view, the general flow is from south to north." (Hinaman, 2005; pp. 2-5)

5.2 AQUIFER STRATA

The principal sources of groundwater in the Powder River Basin at Gillette and the surrounding area, from shallowest to deepest, are as follows:

- Wasatch Formation
- Tongue River Member of Fort Union Formation

- Lebo Member of Fort Union Formation
- Tullock Member of Fort Union Formation
- Lance/Fox Hills Aquifer
- Madison Aquifer

5.2.1 Wasatch Formation

Very little quantitative information about the Wasatch Formation in the Gillette area was found in the existing records. The report titled, "City of Gillette – Water Master Plan Report", dated December 2004, by Wester-Wetstein and Associates provides the following information about the Wasatch Formation:

> ***"Although the city of Gillette has not produced water from the Wasatch Formation since 1981, at one time the Wasatch Formation supplied a large portion of the City's water needs. Water from the Wasatch Formation is characterized as "very hard" with hardness levels of 2,000 mg/L as $CaCO_3$. Wasatch Formation water also has increased levels of TDS, sulfates, manganese, and iron concentrations that exceed the USEPA SMCL's. Wasatch Formation water has TDS, sulfates, manganese, and iron concentrations at 2,800 mg/L, 1,700 mg/L, 0.4 mg/L, and 0.4 mg/L, respectively. The City of Gillette treated the Wasatch Formation water at an electro dialysis plant and the treated water was then blended with water from the Fort Union Formation and Lance/Fox Hills Formation. Development of the Madison Formation Wells allowed the City of Gillette to discontinue producing water from the Wasatch Formation."***

The City of Gillette stopped using groundwater from the Wasatch Formation when groundwater from the Madison aquifer became available in 1981 and has not produced from the Wasatch since 1982. The change from the groundwater source in the Wasatch Formation to the Madison aquifer source was motivated by the relatively poor water quality in the Wasatch aquifer and associated water treatment requirements as well as declining groundwater levels and combined with relatively small and decreasing well yields from the Wasatch.

5.2.2 Tongue River Member of the Fort Union

Multiple sandstone lenses penetrated by wells in the Tongue River Member of the Fort Union Formation are one of three principal sources of groundwater currently used by the City of

Gillette. In general, groundwater from the Tongue River Member does not offer concentrations of chemical constituents exceeding MCL's for primary USEDA drinking water standards. Static water levels in most of the City of Gillette wells completed in the Fort Union Formation, including the Tongue River Member, range from slightly more than 700 feet to nearly 1,000 feet below ground surface, depending on the well. Well yields range from approximately 100 to 160 gallons per minute (gpm) and are probably affected by well construction such that properly constructed wells may provide greater yields per well.

5.2.3 Tullock and Lebo Members

Some of the City of Gillette wells have historically produced from the Tullock and Lebo Members of the Fort Union Formation, usually in conjunction with simultaneous production from the Tongue River Member. Production from the Lebo Shale has been from a few stray sandstone lenses. Production from the Tullock has historically resulted in elevated concentrations of fluoride above the US EPA secondary standards in the water such that production zones in the Tullock have been sealed off and new wells are not being completed in the Tullock Member.

5.2.4 Lance/Fox Hills Formations

The sandstone portions of the Lance/Fox Hills Formations have historically been an important source of groundwater to the City of Gillette. Two of the wells produce groundwater in the range of 500 to 650 gpm, making the source attractive from the standpoint of yield. However, the water is more mineralized than the other groundwater sources used by the City of Gillette and requires blending or treatment when it is in use. Most recently, use of this source has been limited to periods of highest demand when it can be blended with water from the other two sources. This aquifer remains under consideration as a potential source of groundwater in view of its relatively high well yields. Continued use of this aquifer as a primary source will continue to depend on the ability to either blend the water with better quality groundwater from the other sources or subject it to treatment as a less expensive alternative to some of the other options.

5.2.5 Madison Limestone

The Madison Limestone consists of a sequence of marine carbonates (limestone and dolomite) and some evaporates (anhydrite and gypsum) approximately 650 feet thick in the area east of Gillette. Where individual subdivisions of the unit are recognized, it consists of a lower unit called the Lodgepole Limestone, a middle unit called the Mission Canyon Limestone, and an upper unit of evaporates called the Charles Formation, all collectively referred to as the Madison Group. In this report, the Madison is referred to as either the Madison Limestone or the Madison aquifer. Figure 5-4 shows the extent and general thickness of the Madison Limestone.

The Madison aquifer is the third principal source of groundwater for the City of Gillette and has been in production through a well field approximately 35 miles east of Gillette since 1982. Potential production from the well field is currently limited by the condition and hydraulic capacity of the transmission line to Gillette.

Water quality in the Madison aquifer is generally very good; however, the water is very hard, ranging from 470 to 521 mg/L hardness and averaging 488 mg/L hardness as $CaCO_3$. The hardness of the Madison groundwater has been an undesirable factor in part of the City of Gillette service area, most likely because of the seasonal changes from very soft water from the other sources to very hard water when the Madison well field is put into service. Another source of consumer complaints about the hardness of the Madison water is the fact that when the Madison source is in use, some parts of the Gillette service area receive hard water whereas others continue to receive soft water. The contrast in hardness of water received by different customers in close proximity to one another has raised consumer questions about the water quality. It is thought that many of these questions will cease if a consistent quality of water is delivered to the customers throughout the service area and if large seasonal changes in the characteristics of the water delivered to the customer are eliminated.

Five of ten Madison wells have historically produced water with fluoride concentrations ranging from 1.10 to 2.03 mg/L. The fluoride concentrations in water from the other five wells have been less than 1.0 mg/L, ranging from 0.63 to 0.68 mg/L as summarized in the December 2004 Water Master Plan Report by Wester-Wetstein and Associates.

Figure 5-4
Thickness And Structural Contours on the Madison Limestone

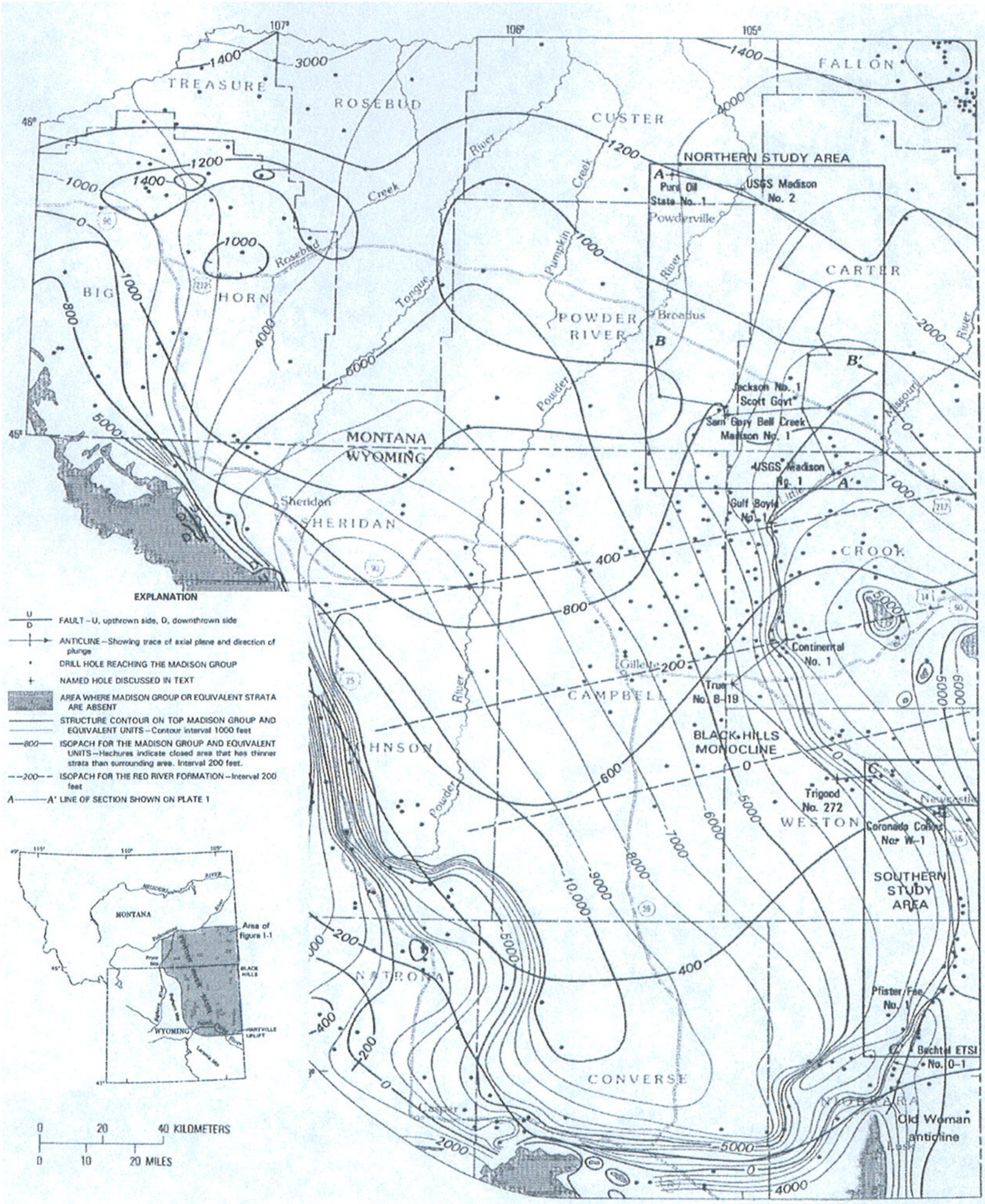


Figure 5-4 reproduced from:
USGS Professional Paper 1330

The completion reports prepared in 1980 and 1981 for the first eight Madison wells by Anderson Kelly, a consulting engineering firm subcontracted to James. M. Montgomery Consulting Engineers, indicated that wells M-2 and M-4 penetrated distinct voids when they were drilled. The results of those reports suggest those wells penetrated enlarged solution cavities in the limestone. The other wells did not penetrate voids and, according to Anderson Kelly consultants, presented different hydraulic response than did the wells that penetrated voids.

Subsequently, well M-3 was hydraulically fractured with gelled sand to prop the fractures open. This increased the performance of well M-3 to equal that of wells M-2 and M-4 which penetrated caverns, presumable by propagating fractures from the well to existing caverns in the limestone. Wells, M-2, M-3, and M-4 produce the lowest concentrations of fluoride of all of the original eight wells. The remaining five wells that did not penetrate cavernous limestone or were not hydraulically fractured to connect them to the zones of enhanced permeability in the limestone produce the highest concentrations of fluoride.

Two additional wells, M-9 and M-10, were drilled into the Madison aquifer in 1995. Both of these wells penetrated voids, presumably solution caverns. Well M-9 provided hydraulic performance similar to that obtained by the other wells that penetrated caverns. Well M-10 provided good yield, but with more drawdown than M-9. Accordingly, it was stimulated by acid fracturing, a process that improved its hydraulic performance to match that of well M-10. Both of these wells provide water with concentrations of fluoride equal to 0.63 mg/L.

These results support the conclusion that groundwater flowing through interconnected caverns in the Madison aquifer has lower concentrations of fluoride than water flowing through non-cavernous porosity in the limestone. This conclusion is consistent with the idea that fluoride concentrations in the groundwater result from dissolution of fluoride minerals in the mineral matrix of the aquifer. Accordingly, the longer the water is in contact with the rock, the more fluoride it will dissolve. If water flowing through caverns is moving at a greater velocity than the flow through non-cavernous limestone, it will have less contact time with the rock and less surface area for contact. Therefore, it should have lower concentrations of fluoride due to less contact time and less surface area for dissolution than water flowing in small pores in the limestone.

If the above conclusion is correct, it indicates that fluoride concentrations in Madison wells at the well field area east of Gillette may be reduced by hydraulically fracturing the wells to connect them with the areas of greatest void openings and, presumably, highest flow velocities in the Madison aquifer.

5.3 RECHARGE CONSIDERATIONS

Shortly after the beginning of extensive mining of coal in the Powder River Basin, the Federal government initiated baseline investigations of surface water and groundwater resources in the region with respect to the potential impact of mining on those resources. In several studies over the years, the USGS attempted to determine the amount of groundwater flow through the region, particularly in the Fort Union Formation which is a major aquifer and is the principal target of the coal mining and in the Madison aquifer which was perceived as a potential source of unappropriated water. The investigations were expanded in recent years with the onset of coal bed methane (CBM) development which involves pumping large amounts of groundwater out of the coal beds in the Fort Union Formation.

5.3.1 Recharge to Tertiary and Cretaceous Strata

The initial investigations of regional and local groundwater flow, principally in the Fort Union Formation, were based on statistical analysis of stream base flow and the chemistry of stream base flow. It was anticipated that the uppermost reaches of the streams and rivers in the Powder River structural basin might exhibit loss of base flow corresponding to stream reaches crossing potential aquifer recharge areas. Likewise, it was anticipated that gains in base flow and changes in surface water chemistry in parts of the surface channels crossing the down-gradient outcrops of the main aquifer strata would reflect discharge of groundwater out of those strata and into the surface streams. The foregoing types of relationships are usually indicative of a major component of regional groundwater flow.

The results of the investigations for the 1956 through 1977, pre-development water years, were reported in a 1985 USGS Open-File Report 85-4229 that was released in 1990. The conclusions of the investigations are quoted below:

"*The types of streams in the area are perennial, ephemeral, and interrupted. Some of the perennial streams in the basin originate in nearby mountains.*

Northward regional ground-water flow that is stratigraphically controlled can be inferred from potentiometric data, but discharge areas in the northern part of the basin could not be identified on the basis of chemical quality of water from springs and shallow wells. The chemical quality of ground water from shallow depth in the northern part of the basin is affected more by local conditions than by regional flow.

Potentiometric data indicate that most streams in the Powder River structural basin should receive base flow from a regional ground-water system. However, such base flow is not evident in streamflow records. Streamflow data collected at fourteen stations on eight streams show that base flow occurs at six of the stations, but base flow during the nongrowing season occurred only in Otter Creek and the Little Powder River. The three largest streams included in the analysis were the Powder, Belle Fourche, and Cheyenne Rivers. Of the three, only the Belle Fourche had base flow, and it was present only during the period of largest precipitation, but not during the period of minimum evapotranspiration. The locations of the streams that do not have base flow and the period of base flow that occurs in most streams indicate that base flow, where present, is from local systems rather than a regional system.

The absence of base flow in streams derived from ground water moving through the regional system is the result of the nonhomogeneity of the formations. The nonhomogeneity of the formations precludes the use of simple water-level maps as a substitute for sets of stratigraphically based potentiometric maps.

Analysis of streamflow records indicates that alluvial and clinker aquifers have more measurable effect on flow at the stations analyzed than do bedrock aquifers. The alluvium contributes flow to some streams, but most streams probably lose water to the alluvium to replace water discharged by evapotranspiration.

The existence of those areas of natural ground-water discharge from a regional ground-water system consisting of the Wasatch-Fox Hills sequence in the Powder River structural basin that would be inferred from potentiometric data could not be substantiated. Therefore, it is concluded that the regional flow system may have a smaller flow than previously thought, and that measurable effects from surface mining and water development will affect mostly local flow systems. However, more data are necessary to describe local and subregional flow systems and their relation to the regional system. (Rankl and Lowry, 1990)

The natural discharge of groundwater from the regional aquifers in the Powder River Basin to surface streamflow was undetectable as either a change in flow or a change in water chemistry at or below where the streams crossed the outcrops of the major groundwater-bearing units. This indicates groundwater flow through the regional aquifers is very slow and the total flow volume is small.

A similar conclusion about the rate of recharge and flow through the regional aquifers in the Wasatch and Fort Union aquifers (and by extension, the Lance/Fox Hills aquifer) was reached in subsequent regional groundwater flow investigations related to CBM development and mining of groundwater to depressurize coal beds so that methane gas will be released. The recent study, based on interpretation of chemical and isotope data for the groundwater, concludes that vertical flow of recharge from the land surface, through the Wasatch beds, and into the multiple layers of sandstone stacked in the Fort Union Formation is very small to nil. One interpretation of the data is that the upper part of the Wasatch is a shallow groundwater system with local circulation, including recharge and discharge areas, that are separate from a nearly stagnant groundwater flow system in the lower part of the Wasatch and the deeper Fort Union strata. Alternative interpretations are that vertical flow is so limited that very little recharge enters the deeper system or that a combination of the latter two flow mechanisms is in effect, with small variations from one location to another.

Figure 5-5 is a depiction of part of the stratigraphic column shown on shown on Figure 5-2, but identifying the coal beds targeted for CBM development. As shown on Figure 5-5, most of the groundwater that will be mined for CBM development will abstracted from groundwater stored in the coal beds in the Tongue River Member of the Fort Union Formation. The sandstone units interbedded with the coal, shale, and mudstone in the Tongue River Member are one of the principal sources of groundwater for the City of Gillette.

Figure 5-6 shows the distribution of coal-lease areas, CBM wells, and permitted locations for CBM wells in the vicinity of Gillette as of June 2000. Subsequent updates of these maps (not shown here) by the Wyoming State Geological Survey dated August 2002 show a large increase in the number of CBM wells permitted in the Gillette area compared to those shown on Figure 5-6. Estimated water production from CBM pumping in the Powder River Basin, most of which is around Gillette, was 400 barrels per day per well for 2,737 wells as of November 2000

(Rice and Nuccio, 2000). This equates to an average daily pumping rate of approximately 31,900 gpm or 71 cubic feet per second (cfs).

Figure 5-7 is a surface geology map showing the clinker outcrops where the principal coal beds in the region have burned at the outcrop. Groundwater stored in the clinker is thought to be the principal source of recharge to the regional aquifer system in the Fort Union strata. Figure 5-8 shows the 1986 potentiometric contours for the groundwater levels in the Wyodak-Anderson coal bed (Figure 5-5) near the top of the Tongue River Member. Potentiometric surface maps for the deeper coal beds and/or sandstone strata have not been prepared for lack of a sufficient number of wells to provide the necessary data; however, the groundwater is presumed to flow generally north.

The conclusion that the volume and rate of recharge entering the regional aquifer system, particularly the deeper part below the uppermost part of the Wasatch, is based on observation of significant differences in groundwater chemistry and the processes that produced that chemistry, between the shallow and deep parts of the aquifer system.

Figure 5-5
Geologic Section With Principal CBM Coal Beds Identified

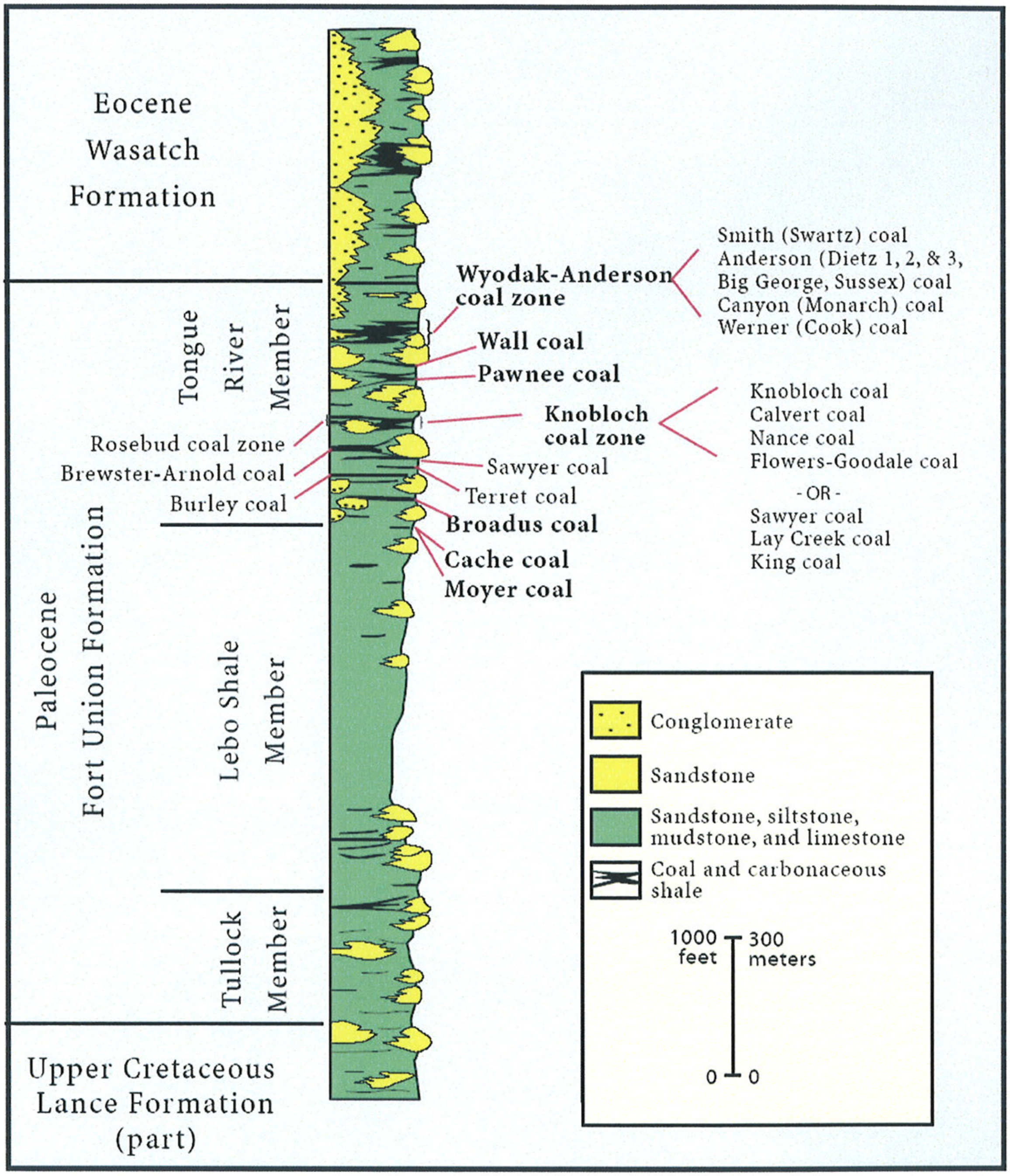
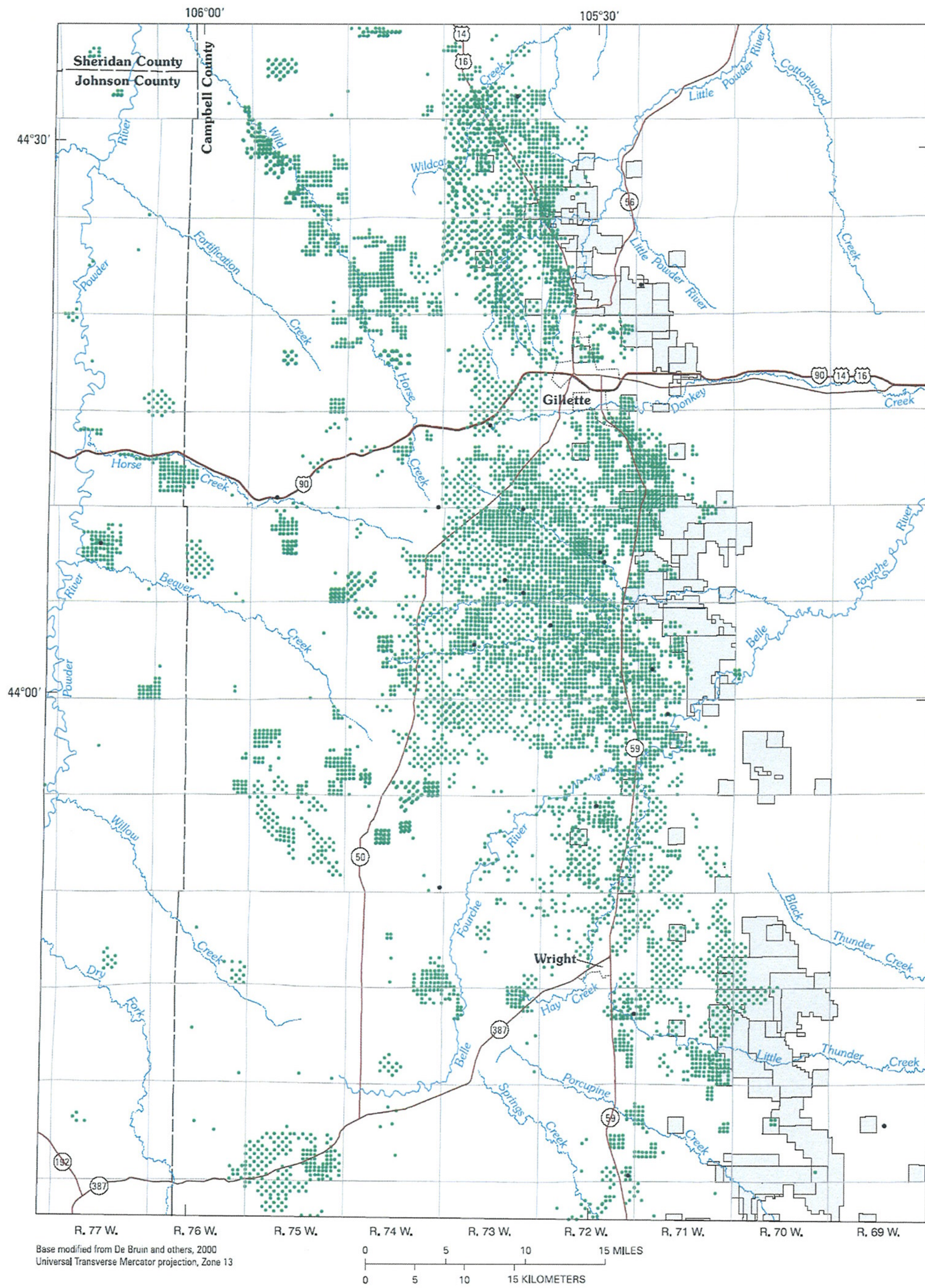


Figure 3. Composite stratigraphic column showing the Upper Cretaceous Lance Formation (part), and Tertiary Fort Union and Wasatch Formations in the Powder River Basin, Wyomin and Montana. Major coal beds and zones in the Fort Union Formation are identified. Coal zones or beds targeted for coalbed methane are bold. (Modified from Flores and Bader, 199

Figure 5-5 reproduced from:
USGS Open-File Report 00-372

Figure 5-6
Coalbed Methane Wells and Permitted Coalbed Methane Well Sites Near Gillette



EXPLANATION

- Coal-lease area
- **Coalbed methane well**--Permitted or production well (as of June 2000)
- **Ground-water sampling site**--spring discharging from clinker, monitoring well, or coalbed methane production well sampled as part of this study

Figure 9. Coal-lease areas, coalbed methane permitted or production wells, and ground-water sampling sites in the study area, eastern Powder River Basin, Wyoming, 1999.

WATER-LEVEL MEASUREI

Figure 5-6 reproduced from:
USGS Water-Resources
Investigations Report 02-4045

Figure 5-7
Generalized geology and location of clinker beds near Gillette

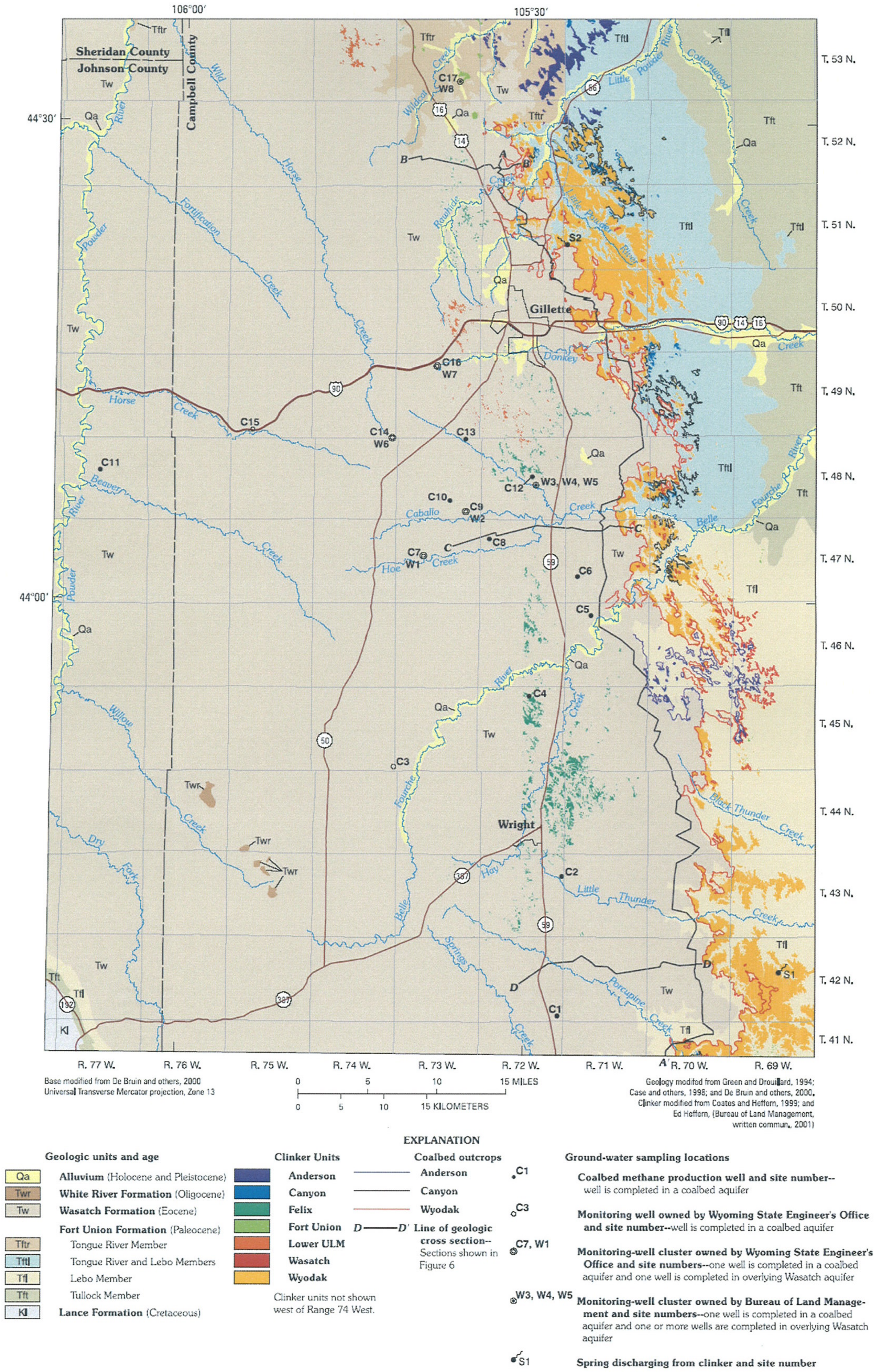
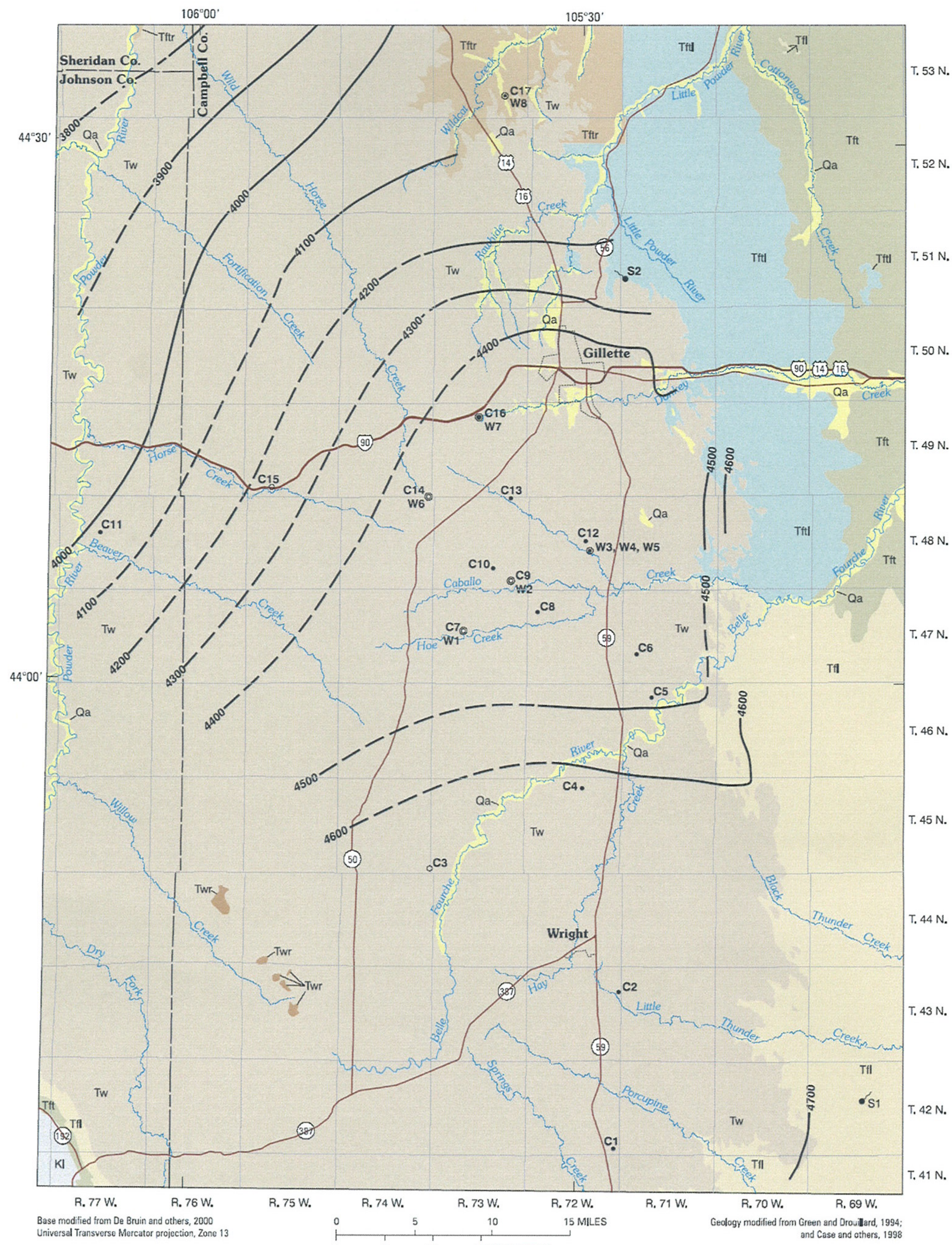


Figure 4. Generalized geology, clinker in the study area, lines of geologic cross sections, and ground-water sampling locations in the eastern Powder River Basin, Wyoming, 1999.

Figure 5-7 reproduced from:
USGS Water-Resources
Investigations Report 02-4045

Figure 5-8
Potentiometric Contours for Wyodak-Anderson Coal Bed in 1986



EXPLANATION

Geologic units and age

- Qa — **Alluvium** (Holocene and Pleistocene)
- Twr — **White River Formation** (Oligocene)
- Tw — **Wasatch Formation** (Eocene)
- **Fort Union Formation** (Paleocene)
  - Tftr — Tongue River Member
  - Tftl — Tongue River and Lebo Members
  - Tfl — Lebo Member
  - Tft — Tullock Member
- Kl — **Lance Formation** (Cretaceous)

4700 — **Generalized potentiometric contour**--Shows altitude at which water level would have stood in tightly cased well completed in the Wyodak-Anderson coalbed aquifer. Represents pre-coal mining and pre-coalbed methane development. Potentiometric surface based on water levels measured 1973-84 (Daddow, 1986). Dashed where approximately located. Contour interval 100 feet. Datum is sea level

Ground-water sampling locations

- C1 — **Coalbed methane production well and site number**--well is completed in a coalbed aquifer
- C3 — **Monitoring well owned by Wyoming State Engineer's Office and site number**--well is completed in a coalbed aquifer
- C7, W1 — **Monitoring-well cluster owned by Wyoming State Engineer's Office and site numbers**--one well is completed in a coalbed aquifer and one well is completed in overlying Wasatch aquifer
- W3, W4, W5 — **Monitoring-well cluster owned by Bureau of Land Management and site numbers**--one well is completed in a coalbed aquifer and one or more wells are completed in overlying Wasatch aquifer
- S1 — **Spring discharging from clinker and site number**

Figure 8. Generalized geology, potentiometric contours for Wyodak-Anderson coalbed aquifer, and ground-water sampling locations in the study area, eastern Powder River Basin, Wyoming, 1999. Potentiometric contours from Daddow (1986).

Figure 5-8 reproduced from:
USGS Water-Resources
Investigation s Report 02-4045

The conclusions drawn from those observations are summarized as follows:

> *"... two possible concepts were hypothesized to explain the observed major-ion chemistry and stable isotope values at three locations with monitoring-well clusters in the study area. The first concept proposes that the changes observed with depth at the three monitoring-well cluster locations are the result of geochemical processes that occur as ground water moves vertically through successively deeper, hydraulically connected sandstone lenses in the Wasatch Formation and finally, into the underlying coalbed aquifer. This investigation, along with earlier investigations described herein, has noted a hydraulic potential for downward ground-water flow within the Wasatch Formation. If geologic conditions are favorable to vertical ground-water flow, geochemical processes such as dissolution, precipitation, ion exchange, sulfate reduction, and mixing of waters are the processes that may occur as ground water moves downward through the Wasatch Formation and evolves the water to the sodium-bicarbonate type observed in the deeper part of the Wasatch Formation and the coalbed aquifers.*
>
> *The second concept assumes the presence of two different aquifers or aquifer systems to explain the differences in major-ion chemistry and stable isotope values observed at the three monitoring-well clusters. Three ground-water samples were collected from shallow wells in this study and all three were collected from the shallow part of the Wasatch Formation (less than about 200 ft below land surface). The wells had mixed cation composition (but generally dominant in calcium and magnesium) with either sulfate or bicarbonate as the dominant anion; all three wells were located at the monitoring-well clusters previously discussed. These wells could be part of a shallow aquifer or aquifer system represented by the "shallow geochemical zone" discussed previously. All ground-water samples collected from wells completed deeper in the Wasatch Formation and the underlying coal beds were sodiumbicarbonate-type waters; these wells could be representative of the underlying, deeper, chemically stagnant geochemical system described by Lee (1981) (described herein as the "deep geochemical zone" composed of the deep sandstone lenses present in the Wasatch Formation and underlying coal beds). In this explanation, little vertical hydraulic connection is present between successively deeper sandstone lenses in the Wasatch Formation and between the shallow sandstone lenses and the underlying coalbed aquifers; very little vertical flow, and therefore, intermixing of waters between the shallow and deep geochemical zones would occur. Heterogeneity and anisotropy, related to discontinuous sandstone lenses surrounded by a predominantly finegrained lithology present in the Wasatch Formation, could have a large effect relative to the actual groundwater flow direction and result in ground-water flow that is primarily horizontal. Other investigators such as Feathers and others (1981) and Lowry and others (1993) have suggested that there is very little vertical ground-water flow in the Wasatch and Fort Union Formations because of the predominantly fine-grained*

lithology, and that ground-water flow in these formations is primarily horizontal through the discontinuous sandstone lenses present. Therefore, well depth (and consequently, differences in ionic composition and stable isotope values) may simply reflect the relative distance water has flowed through the aquifer and different hydrological and geochemical origins and evolutional paths. Waters from the "shallow geochemical zone" may represent waters in local ground-water flow systems with relatively short flowpaths, whereas waters from the "deep geochemical zone" may be representative of a deep, regional ground-water flow system. This explanation also is consistent with differences in water chemistry noted in this and earlier studies, assuming very little vertical ground-water flow through the Wasatch Formation and into underlying hydrogeologic units such as the coalbed aquifers.

Both of the proposed concepts can explain the observed composition of waters in the Wasatch Formation and sodium-bicarbonate composition of waters in the coalbed aquifers at the locations examined. In addition, the concept proposed by Heffern and Coates (1999) discussed earlier also can evolve the water in the coalbed aquifers to a sodium-bicarbonate type. At the basin scale, it is possible, and perhaps most likely, that all three concepts of the ground-water system are correct – the predominant hydrogeologic and geochemical processes at any given location are probably dependant on site-specific geologic and hydrogeologic conditions. In areas where many sandstone lenses are "vertically stacked" above coal beds and the hydraulic gradient allows for downward vertical flow, ground water may move downward through the Wasatch Formation and into the underlying coal beds. In other areas where the sandstones are relatively isolated with limited hydraulic connection, vertical ground-water flow is restricted and flow is primarily horizontal. Despite the localized differences in processes, the overall net effect at the basin scale is the system currently observed. The number of locations where vertical changes in major-ion chemistry and stable isotope values were examined during this study was limited to three locations. Examination of both major-ion chemistry and stable isotope values at additional locations throughout the Powder River Basin may help to refine or alter these proposed concepts of the ground-water system.

Six of eight wells completed in the Wasatch aquifer had no post-bomb water, and two of the eight wells (W2 and W5) had concentrations suggesting a mixture of pre- and post-bomb water, although the low concentrations are suggestive of very little modern water (table 10). One of these two wells (W5) is the shallowest well completed in the Wasatch aquifer (probably the only water-table well). This well had very low concentrations of tritium, indicating that some post-bomb water may be present near the water table. Additional samples at the water table in the Wasatch aquifer should be collected to determine if some modern water is present at or near the water table at more locations in the basin. However, if there was a significant amount of areal recharge, it would be expected that post-bomb water would be distributed throughout the shallow zone of the Wasatch aquifer. The absence of post-bomb water in the shallow *zone would suggest that processes responsible for recharge to the Wasatch aquifer in the Powder River Basin are probably very slow. As discussed earlier, most recharge to the coalbed aquifers is suspected to occur in or near clinker. It is possible that the majority of recharge from precipitation to the Wasatch aquifer may occur in the highly permeable clinker scattered throughout the basin; additional recharge also probably occurs from surface-water drainages in the study area. More accurate age-dating techniques and measurement of recharge rates would be required to understand recharge processes to the Wasatch aquifer.*

Based on the absence of any post-bomb water in samples collected from the coalbed aquifers, it appears that ground water may be flowing very slowly away from the suspected source of recharge, the clinker (which has modern or post-bomb water). Since no tritium data has been collected from wells completed in the coalbed aquifer adjacent to and immediately downgradient of suspected recharge areas (i.e., clinker), the rate at which water enters the coalbed aquifers from its recharge areas is not known." (Bartos and Ogle, 2002)

It is clear from the various studies of flow through the Fort Union aquifer, some based on evaluation of gain in streamflow and changes in water quality at the downstream end of the aquifer system and some based on potentiometric surface elevations and geochemistry of the water at the upstream end of the system, that very little annual recharge or associated groundwater flow takes place on a regional basis. The foregoing investigations conclude that geochemical and isotopic data show that detectable groundwater circulation patterns with recharge areas and discharge areas are limited to relatively small, localized flow systems. They likewise suggest the possibility that groundwater circulation in the deep part of the Fort Union aquifer may be essentially stagnant, thus indicating no recharge is entering or flowing though the system. In any event, such recharge and associated flow to natural discharge areas is too small to detect with existing techniques and records. Certainly, such recharge, if any at all, is less than insignificant compared to demands for community drinking water supplies, oil-field flooding supplies, and CBM dewatering abstractions.

5.3.2 Recharge to the Madison Aquifer

Available information does not provide much factual data about regional recharge to the Madison aquifer, with the exception of studies in the Black Hills of South Dakota which do not directly apply to the area around the City of Gillette's Madison aquifer source area. However, the available information is adequate to support general opinions about the recharge to the regional aquifer as well as the direction of groundwater flow and natural discharge areas.

Figure 5-9 shows a construction of the predevelopment approximately 1950 potentiometric surface of the Madison aquifer copied from (Downey (1984)). The direction of groundwater flow is from recharge areas along the mountain flanks in Montana and Wyoming, including the Black Hills, eastward into North and South Dakota and northeastward into Manitoba, where the Madison limestone terminates in the subsurface. Accordingly, it is concluded that the flow of groundwater to the eastern limit of the Madison Limestone flows vertically upward into overlying units and eastward, through the base of the formation, into older Cambrian-Ordovician aquifer strata that underlie the Madison and extend east of the Madison aquifer limits.

Figure 5-9
Predevelopment Potentiometric Surface of the Madison Aquifer

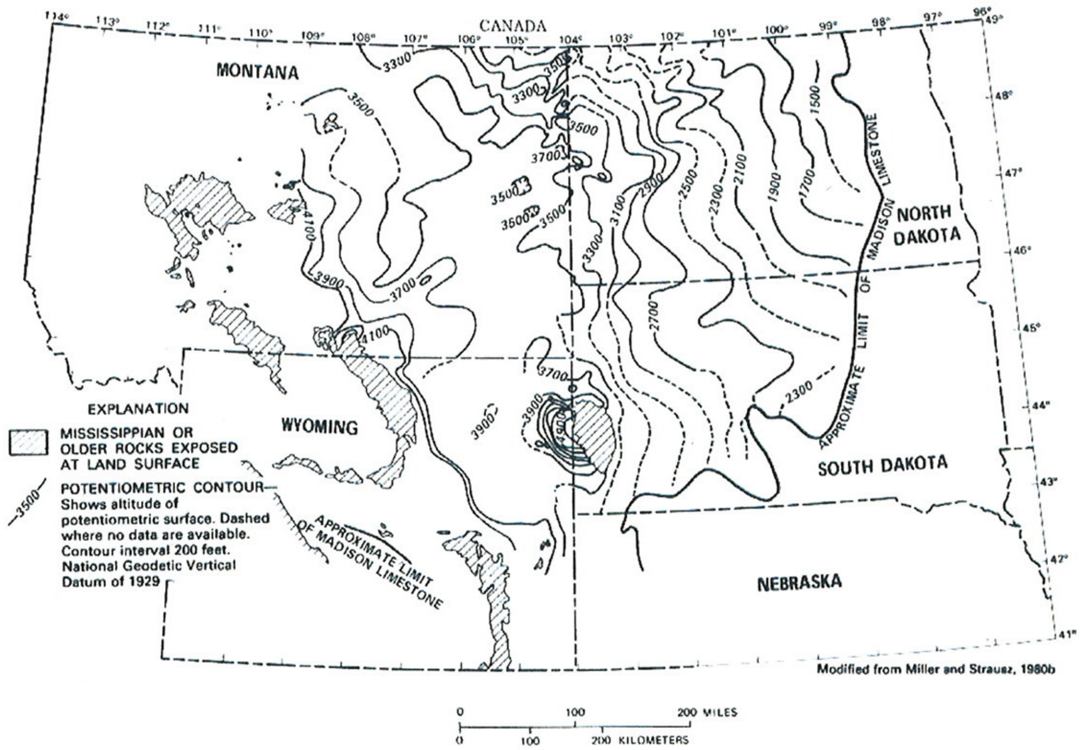


FIGURE 24. —Predevelopment potentiometric surface of the Madison aquifer.

Figure 5-9 reproduced from:
USGS Professional Paper 1273-G

Doney (1984) sums up knowledge as of 1984 about recharge to the Madison aquifer as follows:

> "***Streamflow measurements on several streams draining the east side of the Black Hills . . . indicated that as much as 10 ft^3/s*** [cubic feet per second] ***was being lost from the streams as they crossed the outcrop of the Madison Limestone (Swenson, 1968). Prior to a program of stream-channel sealing during 1937, streamflow losses of about 100 ft^3/s were reported by Powell (1940). On the basis of similar aquifer lithology and degree of weathering, it is reasonable to assume that most streams draining the western mountainous areas, such as the Bighorn Mountains, would lose similar amounts of flow as they cross aquifer outcrops of comparable area.***" (Doney, 1984; pp. G29)

Carter and others (2001) and Driscoll and others (2002), determined an average net groundwater recharge to and net outflow in the Madison and Minnelusa aquifers for the period 1987-1996 of 100 cfs. Figure 5-10 shows distribution of the net groundwater flow around the flanks of the Black Hills. Figure 5-10 also shows the direction of groundwater flow. It should be noted that recharge entering the Madison aquifer along the west flank of the Black Hills does not flow into Wyoming as far as the location of the well field operated by the City of Gillette, but instead turns to flow around the north and south ends of the Black Hills and into the Dakotas.

The 100 cfs of net recharge calculated by Carter and others (2001) and Driscoll and others (2002) does not compare directly to the 100 cfs of streamflow loss referred to by Doney (1984) in the preceding quotation because much of the 100 cfs of streamflow loss cited by Doney (1984) from a secondary reference could have discharged back to the streamflow through so-called headwater springs. Therefore, probably not all of the 100 cfs referred to by Doney (1984) was net recharge to the aquifer. For example, total recharge from combined precipitation and streamflow loss into the Madison aquifer for 1987-1996 ranged from 141 cfs in 1988 to 847 cfs in 1995 and averaged 395 cfs (Carter and others, 2001). Of the average gross recharge of 395 cfs, some 78 cfs discharged through headwater springs back to surface water flow, 189 cfs discharged through artesian springs, and 28 cfs was pumped from wells leaving an average net recharge to the aquifer of 100 cfs for the period.

Figure 5-10
Generalized Groundwater Flow Directions and Flow Rates in the Madison Aquifer in the Black Hills Area

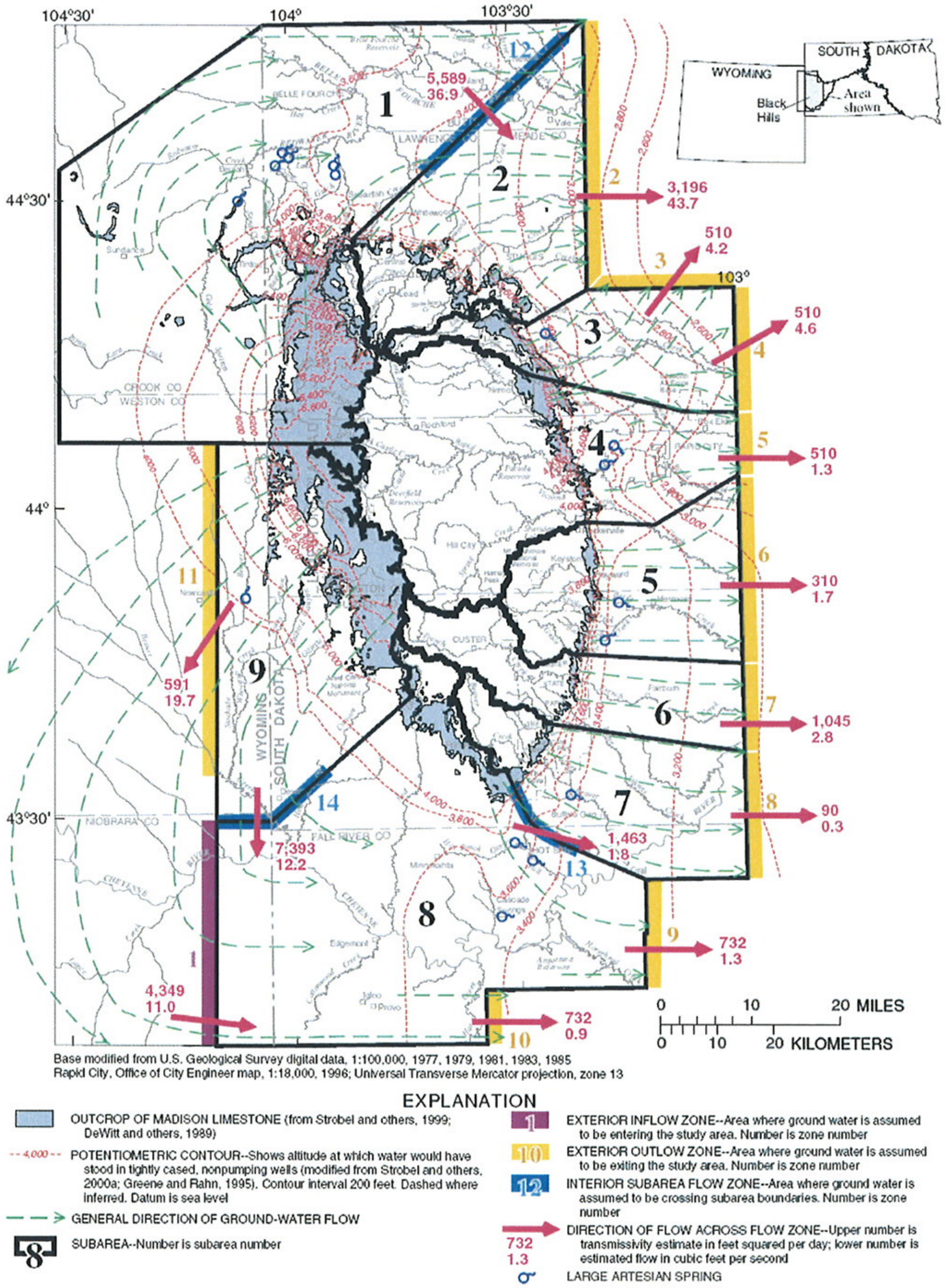


Figure 69. Subareas, generalized ground-water flow directions, and flow zones for the Madison aquifer. Estimated transmissivities and flow components for flow zones also are shown (from Carter, Driscoll, Hamade, and Jarrell, 2001).

Figure 5-10 reproduced from:
USGS Water-Resources
Investigations Report 02-4094

In comparison to the detailed studies indicating a net recharge of 100 cfs to the Madison aquifer in just the Black Hills area for 1987-1996 (Driscoll and others, 2002), the earlier regional study (Doney, 1984) estimated a net groundwater flow through all of the Madison aquifer in the region of 47 cfs. Figure 5-11 summarizes the results of the regional groundwater model presented in Doney (1984). The basis for this estimate is partly described in the citation above, which states how recharge rates were determined, but is ultimately based on a numerical simulation Doney (1984) of steady state flows through the aquifer. Numerical simulations of this nature require broad generalizations and assumptions about the aquifer hydraulic properties, recharge rates, leakage into and out of boundaries, and a host of other factors.

Obviously, the numerical simulation estimating a constant flow of 47 cfs through the entire Madison aquifer does not compare favorably with the detailed studies that determine a flow of 100 cfs in just the Black Hills portion of the aquifer alone.

For example, measurements of momentary streamflow loss into the Madison aquifer in the Judith Basin of Montana in 1963 through 1971 ranged from 13 cfs to 56 cfs on four different stream reaches (Feltis, 1980), again suggesting much more recharge than estimated by the 1984 regional model. While the results of the studies by Driscoll and others (2002) and Feltis (1980) cannot be compared directly to long-term average steady-state flow through the Madison aquifer, because they represent recharge for limited periods of time that may represent fluctuations one way or another compared to long-term average recharge, they do give some indication that the early numerical simulation by Doney (1984) is a conservative underestimation of regional flow through the aquifer.

Figure 5-11
Predevelopment Recharge and Discharge Rates in the Madison Aquifer

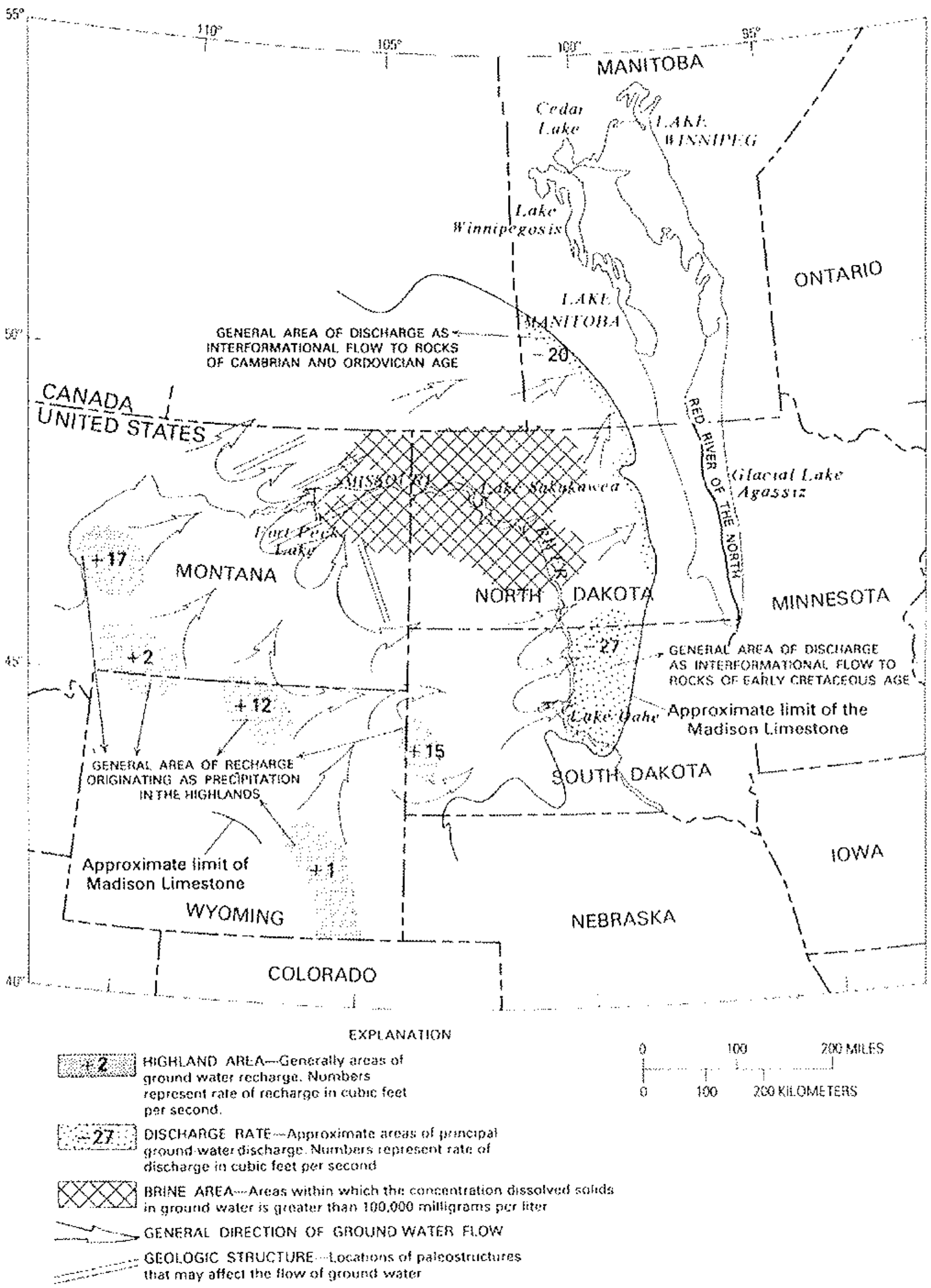


FIGURE 39.—Predevelopment flow directions, recharge areas, and discharge areas in the Madison aquifer (AQ2) as determined through simulation.

Figure 5-11 reproduced from:
USGS Professional Paper 1273-G.

However, it is equally clear that the results of the model in Doney (1984) were more than adequate for their intended purpose which was a preliminary assessment of the magnitude of the groundwater resource that might be available for new groundwater appropriations from the Madison aquifer. Even if the results of the model underestimate the flow through the aquifer by 100, 200, or 400 percent, the long-term steady-state flow through the aquifer is relatively small compared to the potential demands of coal-fired generating plants, agricultural irrigation, municipal water use, and other demands for water in the regional aquifer area and its subareas, such as the Powder River Basin.

It is therefore inevitable that any substantial development of the Madison aquifer will eventually lead to mining of groundwater in excess of the sustainable yield of the aquifer. Therefore, the sustainability of the groundwater developments will depend on the local hydraulic characteristics of the aquifer and the volume of stored groundwater available for mining, not the long-term recharge. This means that the Madison aquifer is not conceptually different from the Fort Union and Lance/Fox Hills aquifers regarding long-term sustainability. Long-term recharge is going to be less than long-term demands at some time in the future.

This latter realization shifts the area of concern from long-term sustainable recharge to the aquifer as a basis for sustainable yield to the issue of long-term sustainability of groundwater development. Sustainable yield and sustainable development are two completely different concepts which must be understood by planners for the City of Gillette's water supply.

5.4 SOURCE RELIABILITY

The City of Gillette has experience with each of the groundwater sources under consideration in this study, having used them in various combinations since the early 1970's. Engineering consultants' reports to the City since at least 1980 describe the various wells in the different aquifers at different times, including the water quality issues and blending of water from the different sources to meet regulatory requirements. A good summary of this history is provided by Wester-Wetstein and Associates in December 2004 report titled, *Coal Bed Methane – Aquifer Storage and Retrieval Project, Level II, Southern Ft. Union Well Field Exploration and Development Study* (Wester-Wetstein, 2004).

The report was one product of a water master planning process conducted by the City of Gillette extending back to at least 1993. The report (Wester-Wetstein, 2004) recommended that future municipal water supplies for Gillette be obtained from a well field to be completed into the upper part of the Fort Union Formation in an area south of Gillette. The recommendation was based in part on identification of a favorable thickness of potentially water-bearing sandstones in the Fort Union Formation at the proposed well field area and a test well that verified acceptable groundwater yield, as well as groundwater quality that was better than that produced by the existing City of Gillette Fort Union wells. The cost for the southern well field in the Fort Union aquifer was estimated in 2004 to be approximately $36.8M as compared to a cursory estimate of $53M to obtain the same amount of water by expanding the existing well field in the Madison aquifer east of Gillette.

However, the history of declining groundwater levels in Fort Union wells operated by the City of Gillette and in surrounding communities, combined with the current regional production of groundwater from the Fort Union Formation by coal-bed methane wells, raised questions about the sustainability of a water supply developed from the Fort Union aquifer south of Gillette. Accordingly, the City of Gillette sponsored this study to review the considerations affecting the sustainability of groundwater supplies developed from both the Fort Union and Madison groundwater sources, as well as to review of how the other groundwater sources in the area around Gillette might play a role in the future.

The sustainability issue is not discussed in detail herein for the aquifer systems in the Wasatch Formation and Lance/Fox Hills strata. The re-worked oil and gas wells that produced groundwater from the Wasatch Formation in the 1970's were abandoned by the City of Gillette in 1980 (Wester-Wetstein, 2004). The hydraulic parameters of the Lance/Fox Hills aquifer are not well enough defined for the purpose of projecting the response of the aquifer to pumping in order to assess the magnitude of development that the aquifer might support. However, it is recognized that the Lance/Fox Hills remains as a potential groundwater source under consideration in this study.

5.5 SUSTAINABILITY CONCEPTS

The term "sustainable yield" is in danger of becoming the modern equivalent of the term "safe yield" used in the past. "Safe yield", originally defined by Meinzer (1920) as "***... the rate at***

which the groundwater can be withdrawn year after year, for generations to come, without depleting the supply", fell into disrepute as a workable concept by the 1950's because, as stated by Lohman (1972), "[hydrogeologists] ***began redefining the concept in more and more precise terms to suit themselves or to suit the particular ground-water conditions with which they were concerned.***" The number of definitions and lack of precision in the many interpretations proliferated made "safe yield" a useless term because it meant essentially whatever its user chose.

Theis (1940) described the natural conditions in an aquifer prior to development of groundwater by wells as follows:

> "***All groundwater of economic importance is in process of movement through a porous rock stratum from a place of intake to a place of disposal. Velocities of a few tens or a few hundreds of feet a year are probably those most commonly met with in aquifers not affected by wells. This movement has been going on through a part of geologic time. It is evident that on the average the rate of discharge from the aquifer during recent geologic time has been equal to the rate of input into it. Comparatively small changes in water level, may occur as the result of temporary unbalance between discharge by natural processes and recharge, but such fluctuations balance each other over a complete season or climatic cycle. Under natural conditions, therefore, previous to development by wells, aquifers are in a state of approximate dynamic equilibrium.***"

Theis (1940) went on to describe the effect that development of groundwater through pumped wells would have on an aquifer:

> "***Discharge by wells is thus a new discharge superimposed upon a previously stable system, and it must be balanced by an increase in the recharge of the aquifer, or by a decrease in the old natural discharge, or by loss of storage in the aquifer, or by a combination of these.***"

5.5.1 Sustainable Yield as Capture

The foregoing statements by Theis (1940) recognized that all water discharged by wells is balanced by a loss of water somewhere in the aquifer system and, therefore, established the hydrologic principal that "sustainable yield", i.e., the magnitude of sustainable groundwater pumpage, depends on the amount of natural discharge that can be captured, the amount of new recharge that can be induced (if possible), or both. In aquifer systems such as those in the Powder River Basin around Gillette, the water table is separated from surface water by considerable thicknesses of unsaturated materials and, in most cases, the bedrock aquifers are confined by beds of very low permeability. Accordingly, fluctuations in the groundwater levels in such aquifers do not influence the amount of recharge or induce recharge by pumping wells. This limits the "loss" or redistribution of water in the aquifer system in response to pumping withdrawals to:

1. Mining of water from groundwater storage and/or

2. Capture of groundwater flow away from natural discharge areas to be redistributed to the pumped wells.

Therefore, it follows that if the amount of pumping remains less than or equal the amount of groundwater flow that can be captured from the natural discharge areas of the aquifer, the pumping is a "**sustainable yield**".

If the pumping does not exceed the amount of flow captured from the natural discharge areas, the aquifer adjusts towards a new dynamic equilibrium, with redistribution of internal hydraulic gradients in response to redistribution of part or all of the aquifer discharge to the pumped wells. The amount of groundwater flow that can be captured, as previously mentioned, depends on the hydraulic parameters of the aquifer, the boundaries of the aquifer system, and the position of the groundwater developments with respect to the aquifer boundaries, particularly with respect to the natural discharge areas.

5.5.2 Groundwater Mining and Sustainability

If the long-term pumping abstractions exceed the amount of groundwater captured away from flow to the natural discharge areas, the difference between pumping and capture must be satisfied by "mining" of water from storage in the aquifer. Mining of groundwater storage is a net depletion of the natural aquifer system and results in continued decline of the water level in the aquifer so long as the mining continues. The magnitude and extent of the decline in groundwater levels is controlled by the hydraulic conductivity, saturated thickness, and storage coefficients of the aquifer for a given rate and duration of groundwater abstraction.

The strictly defined concept of sustainable yield – collective pumped well yield in an aquifer system that does not exceed the amount of groundwater flow that can be captured from the natural discharge of the aquifer – has proven to be a difficult concept to apply in practice. The difficulty arises from the time required for an aquifer to achieve a new "dynamic equilibrium" in response to pumping and capture. In large, complex regional aquifer systems such as those in the Powder River Basin, the time for the cones of depression around centers of pumping to expand to the discharge areas and begin capture of natural discharge may take decades to centuries, depending on the locations of the pumped wells with respect to the natural discharge areas. Additional decades to centuries may be required before the water captured from the natural discharge area becomes a significant component of the pumped water, assuming the collective pumping rate is not greater than the natural discharge rate.

Other complications are that it is often nearly impossible to reasonably quantify the natural discharge from the aquifer system (or the natural recharge) in order to gain some idea of the total rate of pumping that might be sustained from a groundwater basin. The latter problem clearly applies to the Fort Union Formation where it has already been shown that recharge and discharge cannot be quantified and may in fact amount to nearly nothing compared to projected pumping abstractions from the regional aquifer. It therefore becomes impractical or impossible to distinguish between aquifer drawdown associated with sustainable pumping rates (and the initial mining of groundwater storage necessary to establish groundwater flow toward the pumped wells) and drawdown associated with mining of groundwater storage and aquifer depletion for both of the above reasons.

5.5.3 Sustainable Development versus Sustainable Yield

It therefore follows that the entire water budget approach based on attempts to quantify aquifer recharge, natural discharge, and changes in storage is essentially meaningless in large, regional groundwater basins. This condition is because groundwater development will depend on mining of the groundwater storage, not capture of water from discharge areas (sustainable yield), for periods of time that will exceed any anticipated project life or foreseeable planning projections.

So long as mining of groundwater is the principal mechanism of groundwater production, the water budget or water balance approach cannot be used to determine the magnitude of development that an aquifer system can support. Likewise, the magnitude of groundwater development that can be achieved is unlikely to be related to sustainable yield, but instead will be a function of the local hydraulic properties of the aquifer and an acceptable amount of drawdown over a given period of time. In many cases, the planning will be such that the acceptable decline of groundwater levels over a designated period of time will result in groundwater mining and aquifer depletion, simply because the sustainable yield is unknown and our planning concepts do not embrace the amount of time required for aquifer systems to adjust to a new steady-state condition or "dynamic equilibrium" in response to our pumping withdrawals. These considerations were recognized by Bredehoeft et al. (1982), as paraphrased below:

1. The magnitude of sustainable groundwater pumpage depends on how much of the natural discharge from the aquifer can be captured.

2. The magnitude of a "sustainable development" depends on hydrologic effects that you are willing to tolerate, ultimately or at any given time, depending on economics, environmental concerns, water rights issues, or other factors.

The latter definition of "sustainable development" is different than the definition of "sustainable yield". Whereas the sustainable yield of the aquifers in the Powder River Basin will likely remain unknown, as will be discussed later in this report, the amount of groundwater development the aquifers will support for a given period of time with groundwater decline limited to a pre-defined amount can be determined. The magnitude of a sustainable development will depend on the local hydraulic properties of the aquifer and the amount of groundwater level decline that will be allowed by the State regulatory agencies; in this case the Wyoming State Engineer's Office.

The difference between sustainable yield and sustainable development has been recognized for quite a number of years, but has not been applied evenly in the different western states. Some states, such as Montana, have continued to write legislation that is based on the idea that pumping within a groundwater basin shall not exceed the recharge. Bredehoeft (2002), recently revisited this issue with a technical paper on the matter that starts with the following introduction:

"The idea persists within the ground water community that if one can determine the recharge to an aquifer system then one can determine the maximum magnitude of a sustainable development. One commonly hears the statement, "the pumping must not exceed the recharge (if the development is to be sustainable)."

The idea that the recharge (by which one usually means the virgin recharge before development) is important in determining the magnitude of sustainable development is a myth. A number of hydrogeologists have tried to debunk the myth, starting with Theis (1940) in a paper titled "The Source of Water Derived from Wells: Essential Factors Controlling the Response of an Aquifer to Development." Brown (1963) and Bredehoeft et al. (1982) wrote papers debunking the myth. Unfortunately, the message in Brown's paper was apparent only to those deeply schooled in ground water hydrology. The Bredehoeft et al. paper, while more readily understandable, was published in an obscure National Academy of Science publication that is out of print. At the time the Bredehoeft et al. paper was published, Theis congratulated the authors, commenting that he had intended to write another paper on the subject, but now he did not see the need. Needless to say, in spite of these efforts, the myth goes on; it is so ingrained in the community's collective thinking that nothing seems to derail it.

It is presumptuous and perhaps arrogant of me to imply that the entire community of ground water hydrologists does not understand the principles first set forth by Theis in 1940; clearly this is not the situation. There are good discussions in recent papers that indicate other hydrogeologists understand Theis' message. The 1999 USGS Circular 1186, Sustainability of Ground-Water Resources (Alley et al. 1999), state the ideas lucidly. Sophocleous and his colleagues at the Kansas Geological Survey have published extensively on the concept of ground water sustainability; Sophocleous (2000) presents a summary of his ideas that contain the essence of Theis' principles.

On the other hand, I do not find Theis' principles on sustainability expressed clearly in the texts on groundwater. These ideas were taught to me, early in my career, by my mentors at the U.S. Geological Survey. Also I find in discussions with other ground water professionals that these ideas, even though they are 60 years old, are not clearly understood by many individuals. It is my purpose in this paper to address again the myth that recharge is all important in determining the size of a sustainable ground water development, and show that this idea has no basis in fact."

The foregoing comments by Bredehoeft (2002) echo the Bredehoeft et al. (1982) paper which states "***Perhaps the most common misconception in groundwater hydrology is that a water budget of an area determines the magnitude of possible groundwater***

development. Bredehoeft et al. (1982) provide a detailed discussion of sustainable yield from an aquifer, concluding as follows:

> "***The ultimate production of groundwater depends on how much the rate of recharge and (or) discharge can be changed – how much water can be captured. Although knowledge of the virgin rates of recharge and discharge is interesting, such knowledge is almost irrelevant in determining the sustained yield of a particular groundwater reservoir. We recognize that such a statement is contrary to much common doctrine. Somehow, we have lost or misplaced the ideas Theis stated in 1940 and before.***" (Bredehoeft et al., 1982; pp. 54-55)

and

> "***Magnitude of development depends on hydrologic effects that you want to tolerate, ultimately or at any given time (which could be dictated by economics or other factors). To calculate hydrologic effects you need to know the hydraulic properties and boundaries of the aquifer. Natural recharge and discharge at no time enter these calculations. Hence, a water budget is of little use in determining magnitude of development.***" (Bredehoeft et al., 1982; pp.56)

The "magnitude of development" referred to above is, of course, the magnitude of groundwater pumpage that can be sustained under defined conditions. The limitations imposed on sustainable development by the local and regional hydraulic properties of an aquifer must be considered carefully in selecting a tolerable amount of groundwater level decline, a conclusion reinforced by the example of the Ogallala Formation of the High Plains Aquifer. The Ogallala Formation in the southern high plains of Texas and New Mexico is described by Lohman (1972) as follows:

> "***Water is in Tertiary deposits (Ogallala Formation), which have a maximum thickness of about 600 ft and an average thickness of about 300 ft. The material is moderately permeable and rests on relatively impermeable rocks. The recharge, which is derived solely from scanty precipitation, is estimated to range from 1/20 to ½ inch per year, or of the order of $3X10^9$ ft^2 $year^{-1}$*** [3 billion cubic feet per year]. ***The natural discharge, of the same estimated order, is from seeps and springs along the eastern escarpment. The storage of ground water prior to development was very large, of the order of $2X10^{13}$ ft^3*** [20 trillion cubic feet]. ***The withdrawal by pumping has increased from about $4X10^9$ ft^3 $year^{-1}$*** [4 billion cubic feet per year] ***in 1934 to more than $2X10^{11}$ ft^3 $year^{-1}$*** [200 billion cubic feet per year in 1972] ***and is used mainly for irrigation. . . .***

Salvaged natural discharge [capture of natural discharge] ***virtually none; gradient toward eastern escarpment has been virtually unchanged but even if all*** [natural] ***discharge could be salvaged*** [captured]***, it would only amount to 1 or 2 percent of the withdrawal rate.***

. . . virtually all water is being mined from storage and that equilibrium is not being reestablished. Because ground water is a mineral that is being mined without hope of natural replacement, the Federal courts have affirmed the right of eligible ground-water users (those who have, in effect, paid for the water in the form of land prices higher than that of land lacking a good supply) to claim a depletion allowance for Federal income-tax purposes.

Possible remedial measures. – (1) In the Texas section of the region, a water conservation district, to which most affected counties belong, has sought to retard depletion by encouraging water-saving practices and by requiring proper spacing of wells. In the New Mexico section, the State law based on prior appropriation is applied by allowing, in a particular area, appropriations until the remaining supply is judged sufficient for an additional period (such as 30 or 40 years) to enable recovery of investments in land and wells and the creation of wealth through extraction of this "minable" (sic) ***resource . . .*** " (Lohman, 1972; pp.66)

Recognizing the potential for the type of problem experienced due to unregulated mining of groundwater from the Ogallala Formation, a number of western States established definitions of sustainable development (although it may not be called that) for use by planners, engineers, hydrogeologists, and regulators. For example, Arizona passed legislation defining an "Assured Water Supply" as the combined amount of groundwater capture and mining of groundwater storage that would support the proposed magnitude of groundwater development for 100 years. This concept, also referred to as a "100-year water supply", was additionally modified in so-called "Active Management Areas" in Arizona to prohibit lowering of the groundwater levels below a certain depth or elevation in the local aquifer. In other variations to this approach, sustainable development has been defined as the amount of pumping that can be supported for 100 years without decreasing the available water level in the aquifer by more than 50 percent.

Unfortunately, many hydrogeologists have adopted the practice of calling the above approaches to sustainable development the "sustainable yield" of the well or aquifer. This misuse of the term "sustainable yield" blurs the difference between sustainable yield and sustainable development. The difference between the terms is important because it is nearly impossible to arrive at a practical definition of sustainable yield in a large, complex groundwater basin such as the Powder River Basin whereas sustainable development is always based on some assumption about the amount of allowable depletion of the aquifer, usually as established by a regulatory entity.

It is important for groundwater users such as the City of Gillette to recognize that their present and future development of municipal water supplies from groundwater sources in the Powder River Basin, including the Madison aquifer, is based on groundwater mining. It is therefore necessary for them to develop an acceptable definition of sustainable development of groundwater from each aquifer in terms of production rates and an acceptable amount of groundwater level decline over time. It is necessary for the City of Gillette to determine the sustainable development because the State of Wyoming has not formulated guidelines such as a 100-year supply concept or other limits on pumping rates and associated drawdown of groundwater levels.

5.6 EXAMPLES OF SUSTAINABLE DEVELOPMENT CONCEPTS

Based on the foregoing considerations, a brief analysis is provided herein to put the concept of a sustainable groundwater development (not sustainable yield) into perspective with the future needs of the City of Gillette. The analysis is based on a hypothetical example of a Fort Union aquifer well field using the concepts of the Wester-Wetstein (2004) recommendations for a Southern Well Field.

In the hypothetical well field described by Wester-Wetstein (2004), it is assumed that projected future City of Gillette water demand through year 2037 is a peak demand of 21,300 gpm and an associated average daily demand of 6,600 gpm. It is assumed that a 1:1 mixing ratio will be maintained between Madison aquifer water and all other water sources. It is assumed that all other sources besides the Madison will be in the Fort Union aquifer. Therefore, the Fort Union wells must collectively provide 3,300 gpm for the average day. The capacity of the existing City of Gillette Fort Union wells is about 1,115 gpm, leaving a balance of 2,185 gpm of average daily demand to be satisfied by new Fort Union wells in the future.

It is assumed that the new Fort Union wells to provide an additional 2,185 gpm will be located in the Southern Well Field in the pattern laid out with approximately one-mile spacing between wells as described in the December 2004 coal bed methane ARS well study (Wester-Wetstein, 2004; Figure 6-1). In the Wester-Wetstein (2004) study, a test well was completed with 197.6

feet of well screen distributed across 1,937 feet of borehole penetrating the Tongue River and Lebo members of the Fort Union aquifer. Additional screen in the Lebo was plugged off due to unacceptable water quality from that part of the aquifer system. The static water level in the well was 531.6 feet below ground surface when the well was completed and the top of the uppermost well screen was at 1,383.8 feet, leaving an available water column of approximately 852 feet that can be used without lowering the water level in the well to where the uppermost well screen is dewatered.

The Wester-Wetstein tests obtained a 24-hour constant rate yield of 260 gpm from the well with approximately 400 feet of drawdown. Accordingly, a well yield of 268.4 gpm is used in the example as a reasonable well yield for each well in the hypothetical well field. The 15 wells, each pumped at 268.4 gpm for 12 hours per day, will provide the required 2,185 gpm. In the model, an average daily pumping rate of 134.2 gpm, 24 hours per day, is assumed to obtain the average long-term drawdown. For the purpose of defining a sustainable level of development, it is assumed that the water level in each well must be maintained above the uppermost well screen using the static water level and well screen interval described above, i.e., the usable water column in the well for sustained development is 850 feet. This assumption does not take into account the ongoing decline of groundwater levels in the Gillette area observed in dedicated monitoring wells for the past 10 years.

Projection of drawdown in each well uses the hydraulic properties of the aquifer determined by Wester-Wetstein (2004) from an observation well used in conjunction with their test well. They found that the confined aquifer response was radial flow conforming to the Theis analytical model, providing an apparent aquifer transmissivity of 241 ft^2/day (1,800 gpd/ft) and a confined storage coefficient (storativity) of 1×10^{-4} for the collective water-bearing zones producing to the well.

The foregoing hydraulic constants were used in a forward solution to the Theis equation, implemented though a simple spreadsheet model, to predict the drawdown that would occur in each of 15 wells in the well field array adopted from the Wester-Wetstein (2004) report. The model assumes that the 15 wells will provide the average daily demand with an average long-term pumping rate of 134.2 gpm. This pumping rate is equivalent to pumping each well 12 hours per day at 268.4 gpm and allowing them to recover 12 hours each day at zero gpm. This approach does not address the issue of peak water demands and it does not include the maximum momentary drawdown during each pumping cycle. A well loss of 13 feet is added to the drawdown at each pumped well location to adjust the predicted aquifer drawdown to reflect the equivalent water level in a pumped well.

Simulations are provided for 5 years, 10 years, and 20 years of operation under all the foregoing assumptions. The simulations underestimate the maximum momentary drawdown that will occur during real-world well operations and only provide the projection of average long-term groundwater level decline that will occur in association with the assumed operating scenario. Short-term drawdown for operating cycles must be superimposed over the long-term drawdown to determine maximum momentary drawdown. Maximum momentary drawdown was not determined because the limitations imposed on sustainable groundwater development are evident from the average drawdown projections.

For example, Figure 5-12 shows projected average drawdown at the end of 5 years of an average pumping rate of 134.2 gpm in each of the 15 wells. As described above, sustainable development of groundwater in this scenario is limited by an allowable drawdown of 850 feet. As shown on Figure 5-12, more than 750 feet of drawdown is predicted for the center of the wellfield after only 5 years of pumping. This prediction does not include the ongoing decline of groundwater levels in the area.

Figure 5-13 shows the simulated drawdown after 10 years of pumping. The projected drawdown in the middle of the well field equals the allowable drawdown of 850 feet, not taking into consideration on-going groundwater level decline in the area.

Figure 5-14 shows the simulated drawdown after 20 years of pumping. The projected drawdown in the middle of the well field exceeds the allowable drawdown of 850 feet in 11 of the 15 production wells, not considering the existing groundwater recession rate. The foregoing analysis, although cursory, indicates that groundwater production under the stated assumptions is sustainable for less than 10 years. The 10-year simulation on Figure 5-2 shows drawdown at or slightly below the allowable drawdown of 850 feet in two of the wells. The drawdown is between 800 and 850 feet in six additional wells. When on-going groundwater recession rates from existing pumped wells and increased drawdown and well loss associated with 268.4-gpm momentary pumping rates are superimposed on the drawdown shown on Figure 5-2, eight of the fifteen wells in the well field will exhibit drawdown exceeding the allowable limit of 850 feet for sustained groundwater development. Accordingly, the simulations show that the development of 2,185 gpm is not sustainable for 10 years using 15 Fort Union wells on approximately 1-mile spacings or more, under the assumptions used in this modeling simulation.

Figure 5-12
5-Year Average Drawdown for 15 Fort Union Wells

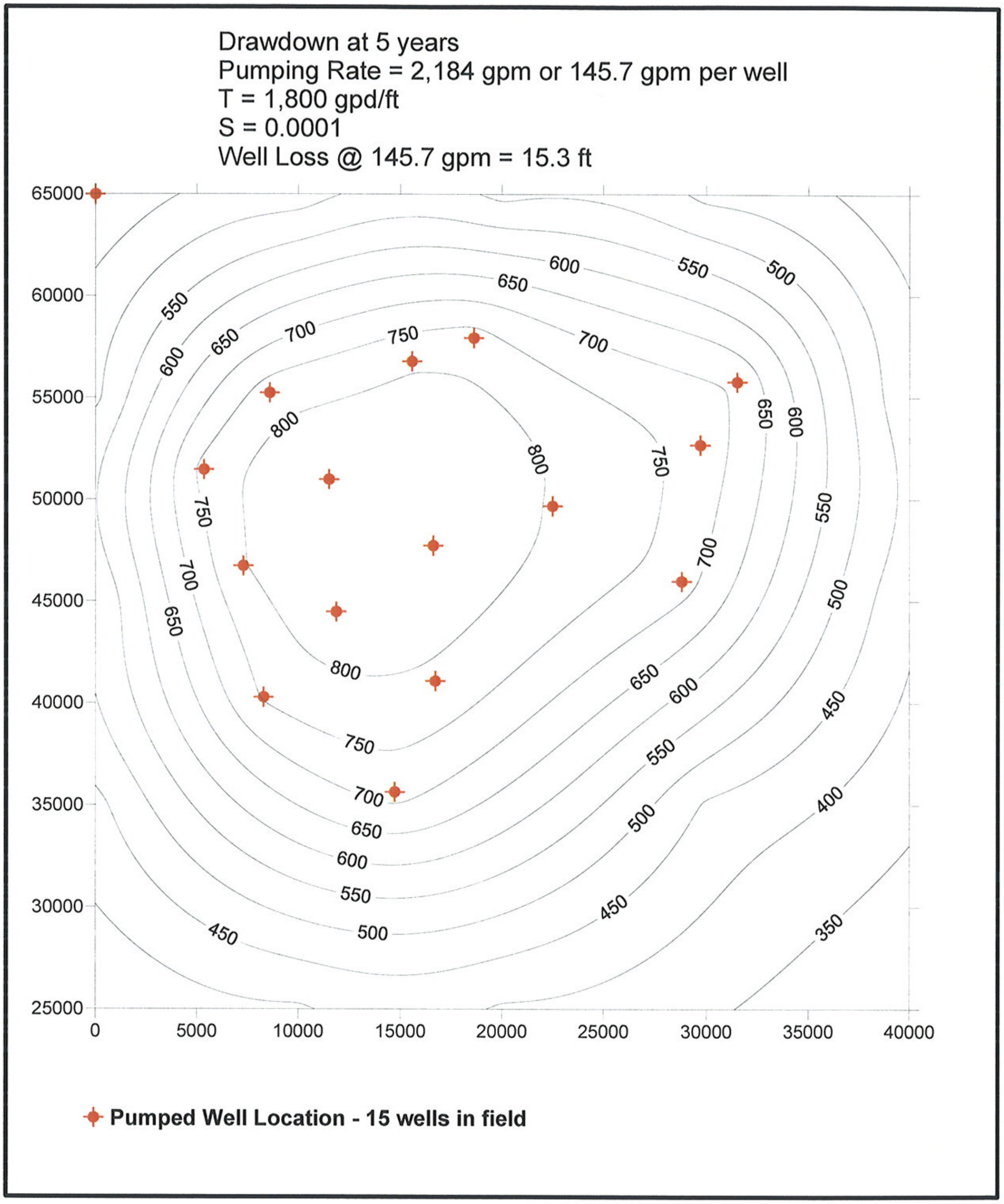


Figure 5-13
10-Year Average Drawdown for 15 Fort Union Wells

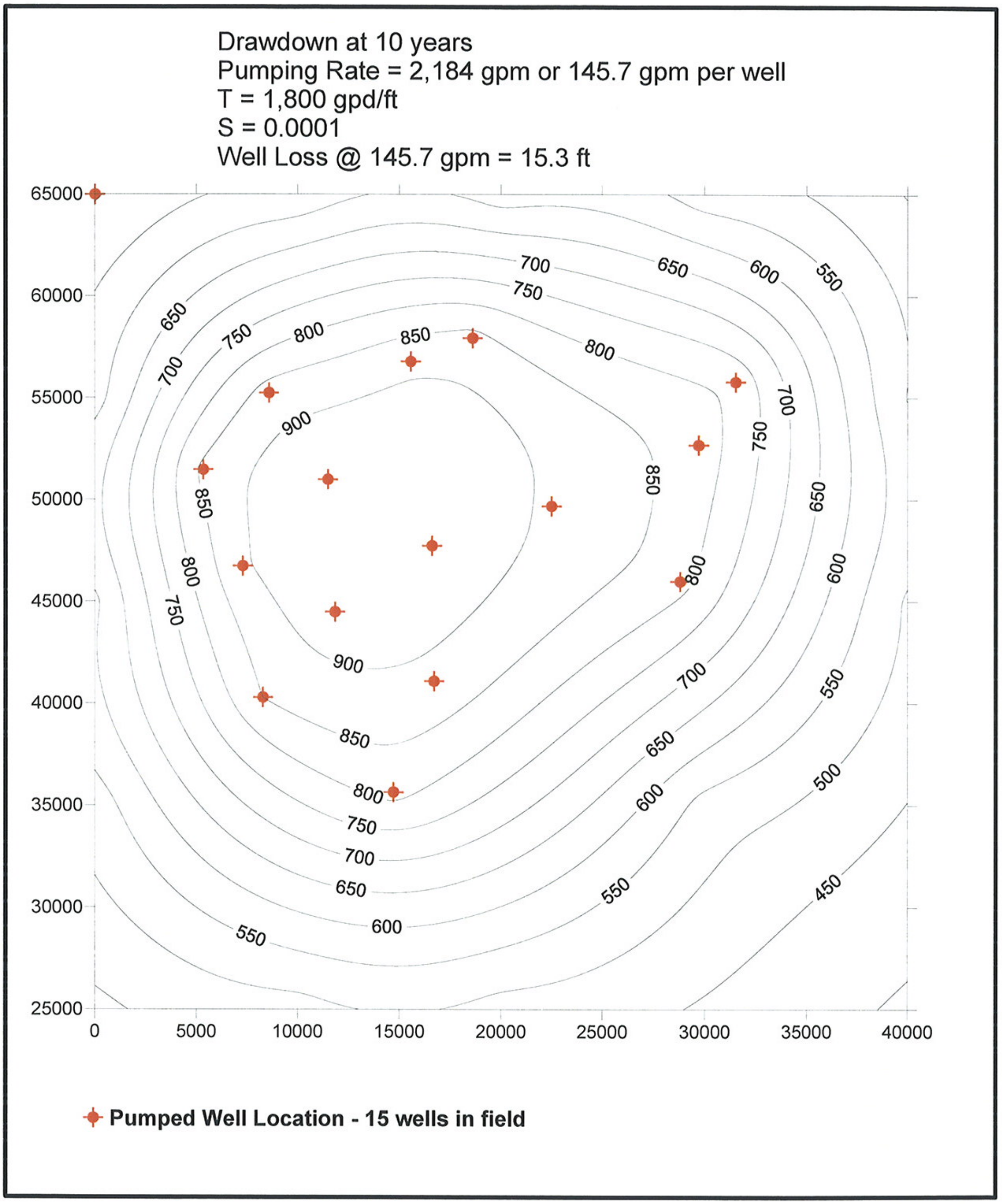
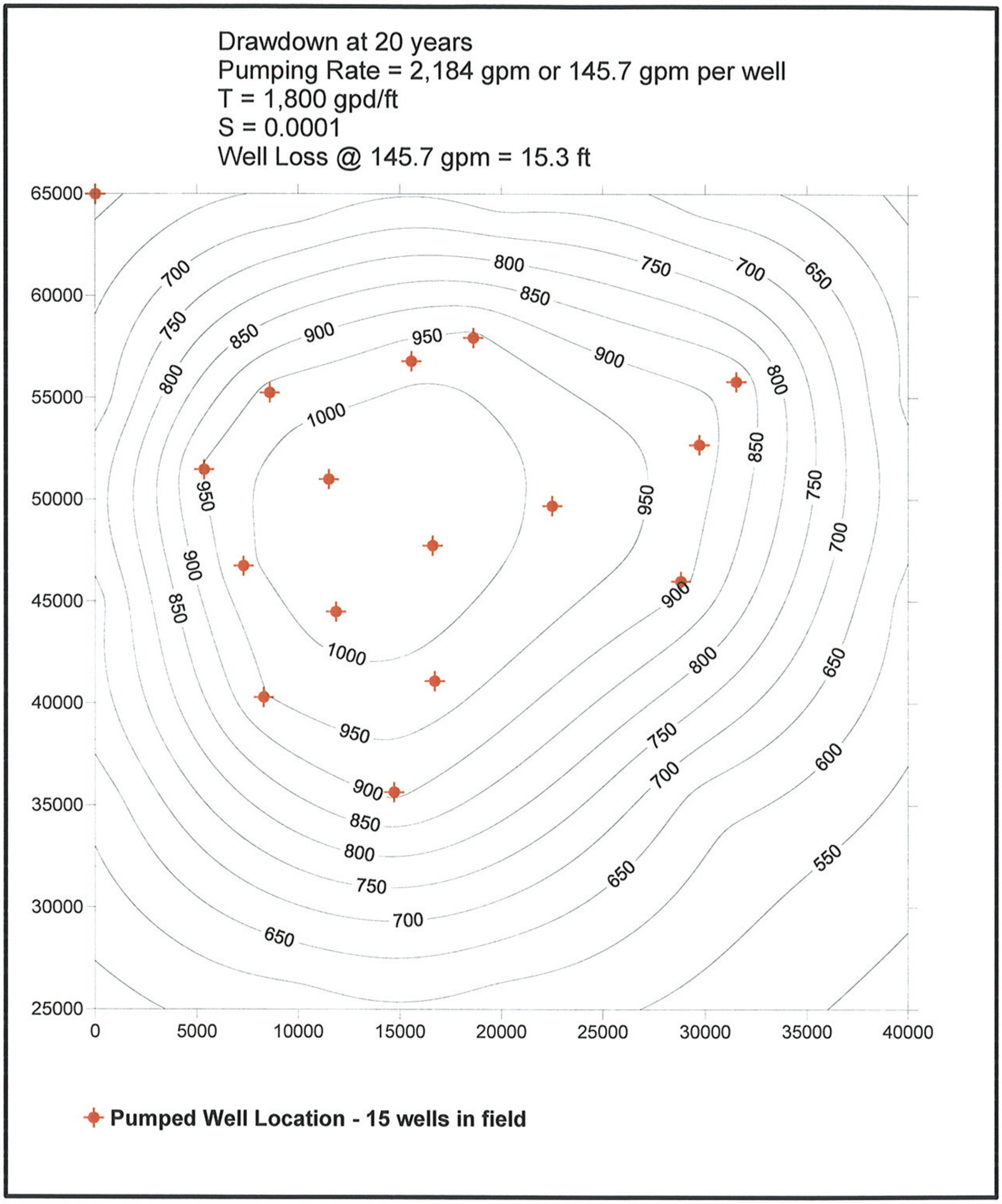


Figure 5-14
20-Year Average Drawdown for 15 Fort Union Wells



In actual operation, a 15-well well field would likely be constructed in phases and the total demand of 2,185 gpm would not be imposed on the well field until quite a few years after the first wells were constructed. Therefore, it could be argued that in actual operation, such a well field could be used for a longer period of time than predicted by the simulations. While this might be true, the implications of a finite well field life, i.e., that production of 2,185 gpm from such a well field could not be maintained for 10 years, after the full pumping demand was imposed, cannot be ignored.

The scenario presented above is not an exhaustive analysis of the potential performance of the proposed Southern Well Field in the Fort Union aquifer; however, it provides a first order insight into the potential aquifer response. The potential aquifer response simulated by the model is consistent with the historic and on-going decline of groundwater levels in the Fort Union in this area of groundwater use by Gillette and surrounding communities. The fact the model predicts greater amounts of drawdown than the historic groundwater recession rate is assumed to be due to the concentration of long-term groundwater withdrawal in the model compared to the relatively more widely spread distribution and variable frequency of pumping in the historic wells.

5.6.1 Sources of Uncertainty

The hypothetical example presented above, based on actual Fort Union aquifer hydraulic parameters measured in the Southern Well Field area, can also be used to demonstrate some of the uncertainty associated with this type of analysis and prediction of future response of the aquifer. Several fundamental sources of uncertainty include natural variability in the aquifer hydraulic parameters, difficulty in interpreting the results of aquifer tests, uncertainty about how to project on-going groundwater level declines in the Gillette area, and uncertainties about the future effects of coal bed methane pumping. An example of each type of uncertainty is discussed below.

5.6.1.1 Uncertainty Caused by Natural Variability

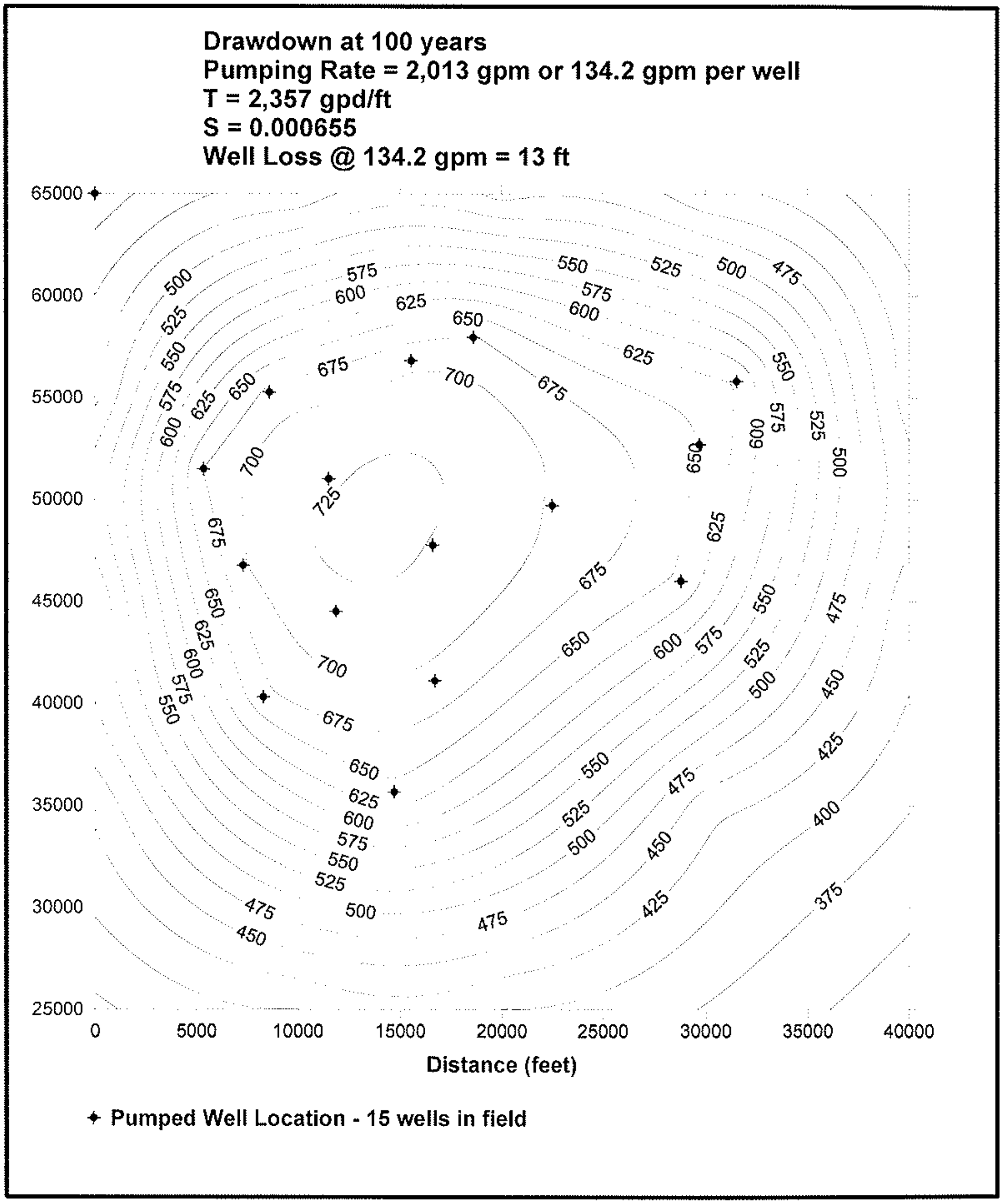
The groundwater model used in the Fort Union well field example above is sensitive to the value of aquifer transmissivity assumed to represent the aquifer. The model as presented above does not take into account potential variability in the distribution of aquifer properties. This situation is due to the fact there are no existing data to quantify the variability in aquifer transmissivity and storativity with the possible exception of the recently conducted test of the Sleepy Hollow Well No. 6, which is located outside of and east of the proposed Southern Well Field area.

Tests of the Sleepy Hollow No. 6 well provided a transmissivity value of 315 ft^2/day or about 2,357 gpd/ft (Brown and Caldwell, 2005). The Sleepy Hollow No. 6 well is in an area outside of the largest collective thickness of sandstone identified on isopach maps compiled by Wester-Wetstein (2004). Assuming there is any correlation between collective sandstone thickness and aquifer transmissivity, the results of the Sleepy Hollow No. 6 well test are inconsistent with the results of the Wester-Wetstein (2004) isopach study and ASR well test. One possible explanation for the smaller value of transmissivity obtained at the ASR test site in the proposed Southern Well Field area is that a suction pipe from the test pump inlet to near the bottom of the well was not used on the ASR test well. It is possible that this resulted in calculation of an erroneously low value of transmissivity for the aquifer at this site due to inefficient energy distribution in the test well, i.e., not all of the water-bearing zones penetrated by the ASR test well may have provided water to the test; therefore, the calculated transmissivity would not have included all available water-bearing zones. However, the Sleepy Hollow well test was also conducted without a long suction pipe and therefore might indicate lower transmissivity than the true value. Other explanations include the possibility that the transmissivity at the Sleepy Hollow site is greater than at the ASR test site or that the interpretation of the Sleepy Hollow test is flawed.

Substitution of the Sleepy Hollow transmissivity value into the model simulation reduces the predicted drawdown significantly, and changes the basic conclusions indicated by the preliminary modeling simulations. There is no reason to reject the ASR test results and replace them with the Sleepy Hollow test results. It is reasonable to conclude that aquifer properties are simply different at the two sites. However, the significant difference between aquifer transmissivity at the two sites also indicates it is reasonable to conclude that a significant range of transmissivity may exist over the extent of an area as large as the proposed well field.

The result of increasing the transmissivity value in the model by 31 percent (1,800 gpd/ft increased to 2,357 gpd/ft) and increasing the storativity to that determined at the Sleepy Hollow No. 6 well completely changes the conclusions with regards to the sustainable development of 2,185 gpm from the Fort Union aquifer. This sensitivity underscores the significance about uncertainty regarding the variability in aquifer hydraulic parameters within the modeled area. Figure 5-15 shows the simulated 100-year drawdown of the Southern Well Field based on transmissivity equal to 2,357 gpd/ft and storativity equal to 6.55×10^{-4}.

Figure 5-15
100-Year Average Drawdown for 15 Fort Union Wells With Sleepy Hollow Aquifer Parameters



As shown on Figure 5-15, a sustainable development of 2,185 gpm is projected for a maximum allowable drawdown of 850 feet over 100 years if the Sleepy Hollow aquifer parameters are applied in the model to simulate future aquifer drawdown.

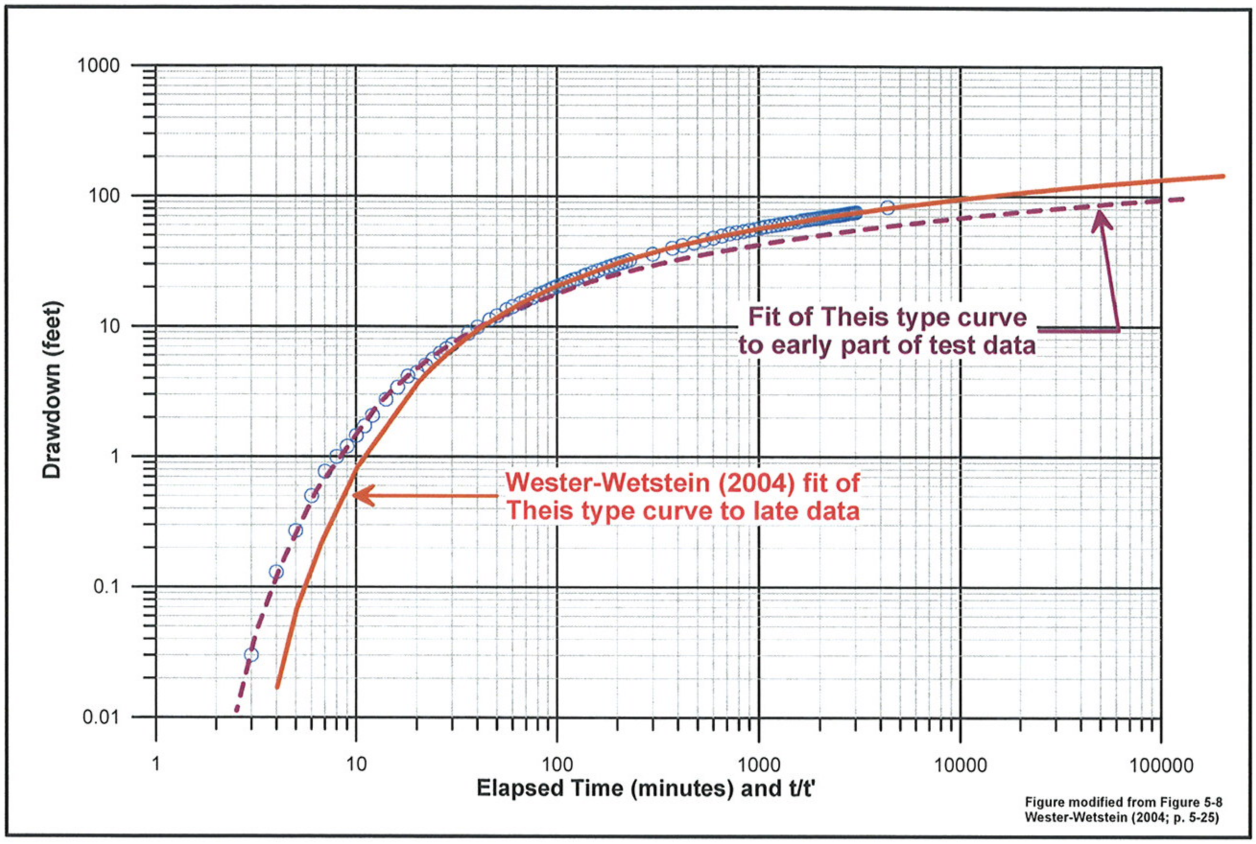
The strikingly different conclusions provided by the foregoing examples of sustainable development evaluation, using a reasonable range of aquifer parameters, simply demonstrate the uncertainty in this type of analysis when very little factual information is available to support the analysis. For example, the projected duration of the sustainable development is changed from less than 10 years to more than 20 years by simply increasing the value of storativity for the ASR site from 1.0×10^{-4} to 5.0×10^{-4}. The latter range of storativity values derived from multiple aquifer tests in the same aquifer are typical, rather than an exception, and they can produce a great amount of uncertainty in projections of the nature presented herein.

5.6.1.2 Uncertainty Caused by Aquifer Test Interpretation

Aquifer test interpretation is another source of uncertainty and may reveal other conditions strongly influencing long-term sustainable development projections. For example, Figure 5-16 below is modified from Figure 5-8 on page 5-25 of the Wester-Wetstein (2004) report wherein the Theis type curve is fit to the data observed in the ASR Monitoring Well while the ASR Test Well was pumped at 260 gpm for 24 hours. The figure shows that the type curve could be fit to the early part of the data with the result that it would not fit the late part of the data and would not accurately project the late aquifer response into the future. Wester-Wetstein (2004) fit the type curve to the late data where it appears to most accurately project the late part of the aquifer test response into the future. Of course, the aquifer transmissivity and storativity values derived from the match between the type curve and the data are significantly different between the alternative fits. The question is which set of values is most representative of the aquifer hydraulics, the early data or the late data?

The fact that not all the aquifer response data fit the type curve indicates something in the aquifer causes the aquifer response to deviate from the Theis type curve. The Theis type curve (Theis, 1940) is the standard analytical model for radial flow to a pumped well in a confined aquifer such as the Fort Union sandstone units penetrated by this well. Hence, the departure of the aquifer response from the type curve indicates that the radial cone of depression expanding away from the pumped well encounters an area in the aquifer where the aquifer properties change significantly or a boundary to the aquifer is reached. Conditions causing real-world aquifer response to depart from the mathematical time-drawdown prediction described by the type curves for different types of aquifer flow are referred to as "boundary conditions".

Figure 5-16
Alternate Fits of Theis Type Curve to ASR Monitoring Well Data



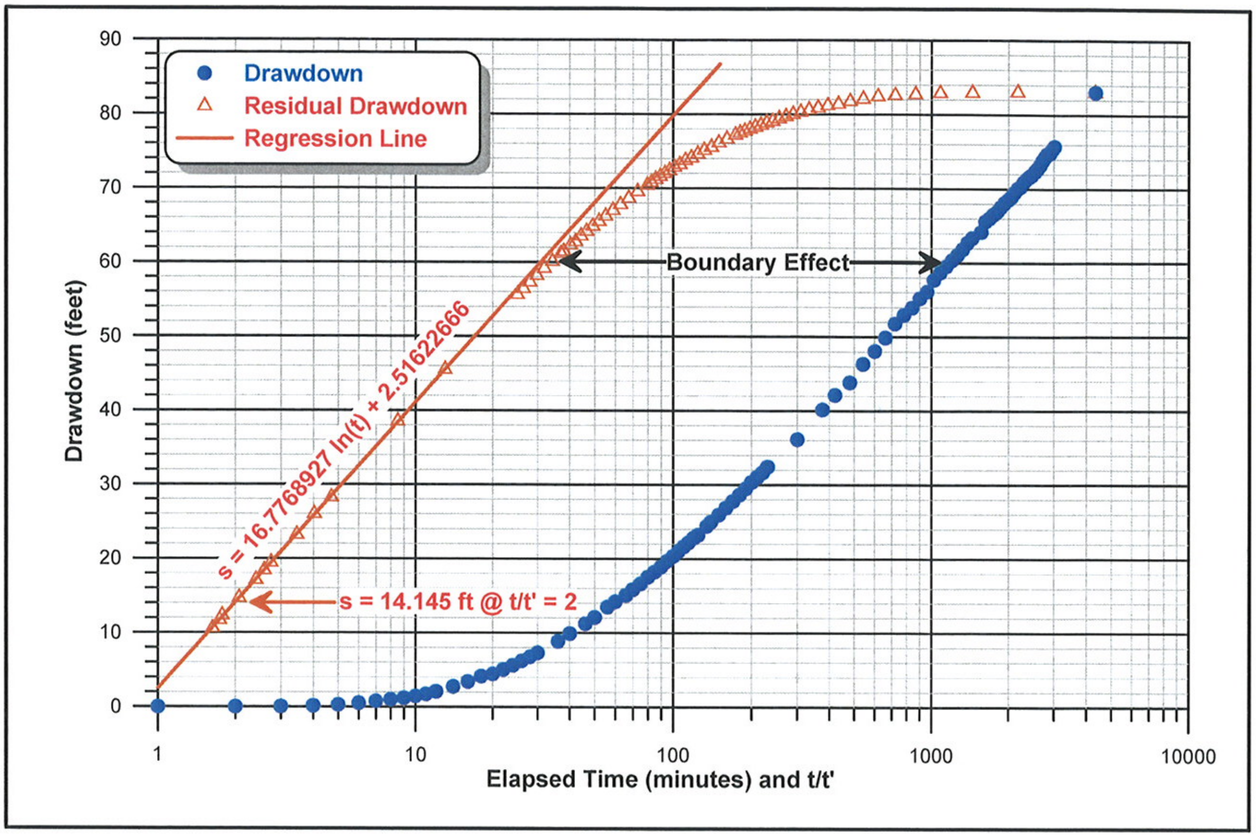
Boundary conditions may include no-flow boundaries (negative boundary conditions), decreases in aquifer transmissivity (partial negative boundary), increases in aquifer transmissivity (partial positive boundary), and constant head recharge to the aquifer (positive boundary) from surface water sources, to name a few. Recognizing that the departure of the ASR Monitoring Well data from the Theis type curve indicates the presence of a boundary condition in the aquifer, the question then becomes how that modification of the radial flow response affects the long-term aquifer response in the future, if at all. Additionally, does the Wester-Wetstein (2004) fit of the type curve to the late aquifer response provide the correct solution to adequately describe the hydraulic parameters of the aquifer and the aquifer's response to future pumping?

Because boundary conditions external to the well are encountered as the cone of depression expands away from a pumped well, type curves are generally fit to the early part of the aquifer response, prior to the onset of the boundary condition that modifies the aquifer response and causes the departure from the type curve. Exceptions to this general statement are caused in single pumped wells where damage to the formation near the well bore and non-laminar flow make the early test data to depart from the early part of the type curve. The ASR Monitoring Well is an observation well, not a pumped well, and therefore is not subject to the latter effects. Accordingly, the logical fit of the Theis type curve to the ASR Monitoring Well response is to the early part of the data as shown on Figure 5-16.

Figure 5-16 shows that after about 30 minutes of elapsed pumping time, the aquifer response to the test begins to depart from the Theis type curve, with the drawdown proceeding at a greater rate than predicted by the type curve. This relationship is typical of the cone of depression around the pumped well encountering a partial or full negative boundary, i.e., a decrease in aquifer transmissivity due to decreased thickness and/or decreased hydraulic conductivity or a no-flow boundary. The fact that the aquifer response observed is obtained from multiple sandstone beds in the Fort Union strata indicates that either one or two water-bearing sandstone units control the aquifer response or the cause of the boundary condition is intrinsic to most or all of the sandstone strata. The next step is to attempt to determine what type of negative boundary affects the aquifer response to pumping.

Various types of analytical plots of time-drawdown data from aquifer tests are diagnostic of specific types of aquifer flow and/or boundary conditions. Figure 5-17 is a semi-logarithmic plot of arithmetic drawdown versus logarithmic time. The plot is referred to in groundwater hydrology literature as the Cooper-Jacob straight-line solution wherein a semi-logarithmic straight-line response is diagnostic of confined radial flow to the pumped well or unconfined radial flow with insignificant dewatering of the total saturated thickness of aquifer at the well. The Cooper-Jacob plot is a special condition of the Theis type curve and simply represents that part of the Theis type curve during late aquifer response where the Theis type curve is changing so slowly that it can be represented by a semi-logarithmic straight-line plot.

Figure 5-17
Cooper-Jacob Plot of ASR Monitoring Well Response



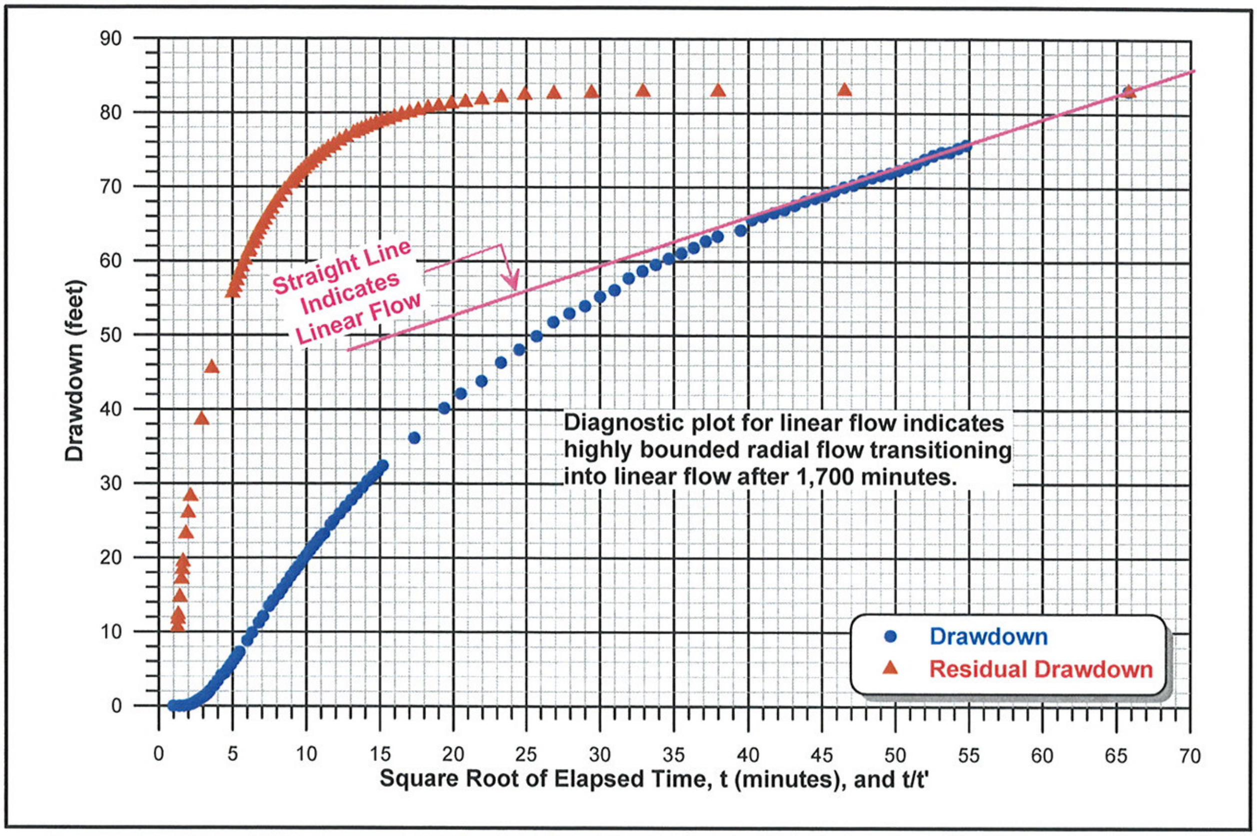
For example, the aquifer test data on Figure 5-16 are approaching a semi-log straight-line response after about 1,000 minutes of pumping as shown by the decrease in the rate of drawdown and the progressive straightening of the type curve after 1,000 minutes. However, the fact that the semi-logarithmic straight-line does not represent the aquifer response until after the onset of the boundary effects, which start after about 30 minutes of pumping, indicates that any solution for aquifer parameters derived from either the late data straight-line solution or the Theis type curve fit to the late data, as in Wester-Wetstein (2004), will provide inaccurate values of aquifer transmissivity and storativity because the shape of the time-drawdown plot has been modified by the boundary effects. Therefore, the conclusions of this study differ from those presented in the Wester-Wetstein (2004) report regarding the aquifer hydraulic properties. These conclusion may be verified with the plot on Figure 5-17 and subsequent specialized plots diagnostic of certain types of aquifer flow conditions.

One of the strong suits of the Cooper-Jacob plot as a diagnostic tool is the ability to compare time-drawdown response during pumping to time-residual drawdown response during recovery from pumping. Observation well drawdown plotted versus time since pumping started, t, and the residual drawdown plotted versus t/t', where t' is time since pumping stopped, should provide curves that coincide, assuming an isotropic and homogenous aquifer of infinite extent (at least with respect to the effects of pumping). The drawdown and recovery curves plotted on Figure 5-17 do not coincide. The recovery curve is shifted to the left of the drawdown curve – the typical effect of a partial or full no-flow boundary encountered by the cone of depression during the pumping period.

The effect of the negative boundary condition is demonstrated by the recovery curve. The well was pumped for 72 hours. Without a negative boundary effect, the water level in the well should recovery fully from pumping within 72 hours; however, as shown on Figure 5-6, 14.145 feet of residual drawdown remain after 72 hours of recovery (t/t' = 2). This condition indicates that if the well is pumped on a cyclic schedule, it will not fully recover from pumping between cycles unless the recovery periods are substantially longer than the pumping periods. Any residual drawdown remaining in the well at the onset of each pumping cycle will be cumulative over the lift of the well. For example, if the well is pumped for 72 hours and then allowed to recover for 72 hours, 14.145 feet of additional residual drawdown will remain at the beginning of each new pumping period such that after five pumping cycles, residual drawdown will have accumulated to 70.725 feet. The accumulated residual drawdown will be different for different pumping rates and durations.

Figure 5-18 is a diagnostic plot for linear flow. Linear flow in an aquifer occurs where flow to the pumped well is controlled by parallel or nearly parallel boundaries such as fractures or shear zones, strip aquifers bounded by less permeable material, or flow through elongate lenses embedded in less permeable matrix. Linear flow produces a straight line where arithmetic drawdown and residual drawdown are plotted versus the square root of elapsed pumping time or t/t', respectively, whereas radial flow or non-linear flow produces a curved line. As shown on Figure 5-18, the response observed at the ASR Monitoring Well indicates the onset of linear flow after approximately 1,700 minutes of pumping time.

Figure 5-18
Drawdown Versus Square Root of Elapsed Time for ASR Monitoring Well

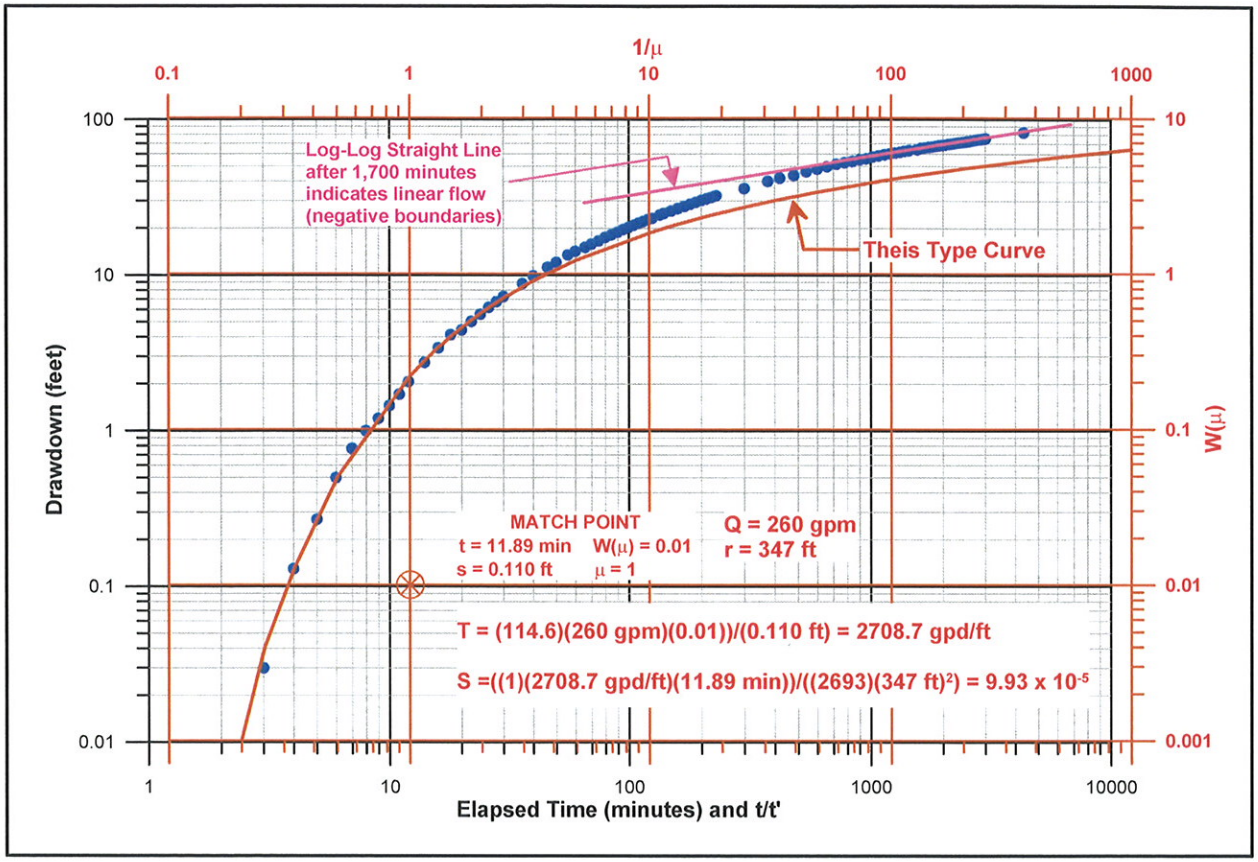


The linear flow response during the late part of the 72-hour aquifer test is consistent with the nature of the Tongue River and Lebo Members of the Fort Union strata which consist of various amounts of elongate, discontinuous lenses of channel sands embedded in finer grained overbank deposits. Expansion of a cone of depression within such lenses is initially under radial flow. When the cone of depression reaches one or another of the sides of a lens, radial flow becomes bounded and begins a transition from radial flow through bounded radial flow to linear flow when the cone of depression reaches the other side of the lens. Figure 5-18 reflects the progression from radial flow through linear flow. The fascinating aspect of this response is that it evidently occurs uniformly from multiple potential production zones from individual lenses. This phenomenon may indicate that the boundary conditions are related to the limits of the thickest cumulative sandstone percentages in the Fort Union strata as identified by Wester and Wetstein (2004) rather than individual boundaries in individual sandstone lenses.

The first hydraulic implication of linear flow response is that linear flow on a log-log plot, such as Figure 5-16, is a log-log straight line. A log-log straight line in the late aquifer response results in much more drawdown as pumping time increases than is predicted by the Theis type curve. Therefore, the fit of the Theis type curve to the late data in the Wester-Wetstein (2004) solution to the pumping test (Figure 5-16) is not a valid projection of the time-drawdown relationship into the future.

Figure 5-19 shows the Theis curve fit to the early aquifer response, before onset of negative boundary effects, as well as a log-log straight line fit to the late data showing a greater rate of drawdown than predicted by the Theis type curve. Figure 5-19 also shows a Theis non-equilibrium solution for the early transient response of the aquifer to the 260-gpm pumping rate. The solution provides transmissivity of 2,709 gpd/ft and storativity of 9.93×10^{-5}, values that are considerably different than the Wester-Wetstein (2004) solution and a transmissivity that exceeds that determined by Brown and Caldwell (2005) at the Sleepy Hollow No. 6 well.

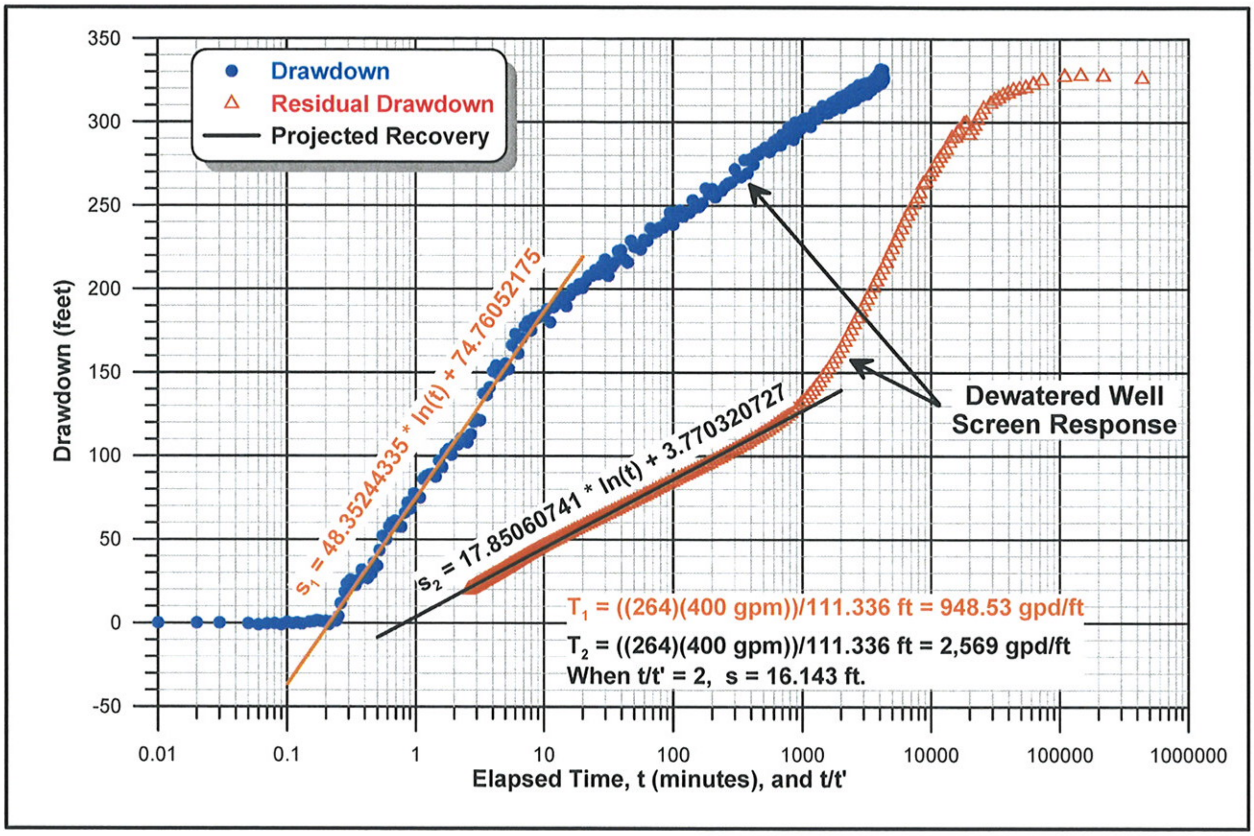
Figure 19
Theis Solution to Early Aquifer Response in ASR Monitoring Well



The second hydraulic implication of the linear flow response in the late data from the 72-hour response is that the discontinuous sandstone lenses in the Fort Union may not only be laterally bounded, as indicated by the linear flow response, but they may have finite length. In other words, future expansion of the depressurized area around a well field in the Fort Union strata in the Southern Well Field area may extend the composite "cone of depression" or depressurized area in the groundwater reservoir to additional negative boundaries where the sandstone lenses pinch out along their lengths. Although this effect is not evident in the 72-hour test response, it may occur over a period of years, as the aquifer is depressurized, and would result in considerable acceleration of the rate of drawdown in the well field area. The latter possibility is a large uncertainty in predicting future decline of groundwater levels and associated loss of well yield, based on our currently available predictive methods and observed test results.

A review of the single-well test time-drawdown data from the Brown and Caldwell (2005) test of the Sleepy Hollow No. 6 well reveals a response similar to the response at the ASR Test Well site – less than full recovery of groundwater levels after the 72-hour test – however, in a more complex aquifer response than observed at the ASR Test Well. Figure 5-20 shows the relationship between drawdown and recovery response on a semi-log plot.

Figure 5-20
Cooper-Jacob Plot of Sleepy Hollow No. 6 Aquifer Test Response



As depicted on Figure 5-20, there is an apparent shift between the drawdown and recovery curves and at t/t' = 2, the projected residual drawdown is 16.143 feet, i.e., when recovery time equals the preceding pumping time, 16.143 feet of drawdown remains in the aquifer. This initially appears to be another negative boundary effect, similar to that observed in the ASR well test; however, it is not a negative boundary and the cause is different. The cause in this case is the lowering of the pumping water level in the well below the uppermost three screened intervals in the well during the 72-hour test and the associated change from release of confined storage to the release of unconfined storage at the faces of the dewatered aquifer zones.

Specifically, when the pumping water level in the well reaches the top of the first screened interval in the well, at 762.5 feet, or at approximately 180 feet of drawdown on Figure 5-20, the rate of drawdown abruptly decreases and begins a new straight-line segment with much less slope than that in the earlier aquifer response. A second decrease in the slope of the recovery plot occurs after about 900 minutes of elapsed pumping time. The first change in the slope of the time-drawdown drawdown curve is clearly associated with the pumping water level declining below the top of the first well screen. A similar change in slope occurs in the time-residual drawdown curve as the recovering water level in the well rises above the top of the uppermost well screen. This change is not an unusual response in pumped wells and results from dewatering of the well screens.

The Fort Union aquifer penetrated by the Sleepy Hollow No. 6 well, similar to that at the ASR Test Well site, is a confined aquifer. The ASR site test revealed a confined aquifer storativity value of 9.93×10^{-5}. Typical confined aquifer storativity values range from 5.0×10^{-3} to 5.0×10^{-5}. The physical basis for confined groundwater storage is elastic compression of the aquifer and of the confined groundwater. Groundwater supports part of the lithostatic load or overburden load in a confined aquifer, resulting in some compression of the water in the aquifer. When a well extracts groundwater from storage in an aquifer, part of the load supported by the water is transferred to the mineral matrix of the aquifer. The increased load on the aquifer matrix compresses the voids in the aquifer. The water released from the voids as they compress and the expansion of the compressed water as pressure is released provide the groundwater storage released from a confined aquifer.

The volume of storage released from a unit volume of confined aquifer material under a unit change in head is the "specific storage" of the confined aquifer and the volume of storage released from the entire thickness of a confined aquifer under a unit change in head over a unit area is the confined aquifer "storativity". By comparison, the release of water from the same material under unconfined conditions is essentially equal to the effective porosity from which stored groundwater drains under gravity and is termed the "specific yield" of the unconfined aquifer material, equal to the volume of water that will drain out of a unit volume of saturated unconfined aquifer. Typical specific yield values for sandstone are 0.05 to 0.30. The change from release of storativity under confined conditions to the release of specific yield at the face of the dewatered screen intervals is the cause of the changes in the slope of the time-drawdown curve in the Sleepy Hollow No. 6 well.

When the drawdown during the test of the Sleepy Hollow No. 6 well increased to more than about 180 feet, or a depth of about 762.5 feet in the pumped well, the uppermost three well screens in the well were progressively dewatered, with the final drawdown of about 326 feet equivalent to a pumping water level depth of about 911 feet in the well, nearly at the bottom of the third well screen interval from 902.4 to 912.4 feet (Brown and Caldwell, 2005; p. 4-18). When the well screen intervals are dewatered, the faces of the dewatered water-bearing zones change from confined aquifer conditions to unconfined aquifer conditions. Accordingly, the aquifer material at the face of the dewatered zones and extending back into the aquifer for some distance begins releasing specific yield storage.

The ASR test results provided a storativity value of 9.93×10^{-5} which, rounded off, is essentially 1.0×10^{-4} or 0.0001 and is likely representative of the storativity of the thick, multiple water-bearing zone parts of the confined Fort Union aquifers, including the area around the Sleepy Hollow Well No. 6. Therefore, as the face of the uppermost water-bearing zones in the Sleepy Hollow No. 6 well were dewatered, the sandstone in the dewatered zones began releasing water from storage at a rate in the range of 0.05 to 0.3 cubic feet per foot of head change over a square foot of aquifer depressurized aquifer as compared to the previous rate of 0.0001 cubic feet per foot of head change under confined conditions. When this happens in a well with only one water-bearing zone, drawdown often ceases for a while until continued pumping begins to generate a measurable cone of depression in the unconfined aquifer storage. Subsequent drawdown is then associated with a cone of depression in which the part nearest the well is releasing unconfined storage whereas the part of the cone of depression further from the well is releasing confined storage. The initial rate of drawdown in such a well is therefore controlled by the confined storativity whereas the rate of drawdown at late pumping time is controlled by the unconfined specific yield of the unconfined part of the cone of depression.

The response of the Sleepy Hollow No. 6 well is analogous; however, the deeper water-bearing zones in the well continue to produce from confined storage while only the uppermost three zones ultimately yielded unconfined storage. Therefore, the contribution of storage from specific yield in the uppermost three water-bearing zones modified the shape of the time-drawdown curve significantly but did not totally offset the confined aquifer response from the remaining deeper zones that remained confined throughout the test. The same phenomenon prevailed during recovery, with the exception that the curve of residual drawdown versus t/t' does not exhibit a change in slope analogous to the change that occurred in the time-drawdown curve after about 900 to 1,000 minutes. This condition occurs because of the reverse of the specific yield effect, the overriding influence of the 0.05 to 0.3 cubic feet per foot of head change per square foot of depressurized aquifer was already established and continued without change until the recovering water level rose above the top of the uppermost water-bearing zone in the aquifer.

Most importantly, those portions of the cones of depression (depressurization) in the uppermost three water-bearing intervals that became unconfined were by definition dewatered by drainage of the effective porosity as a contribution to the water pumped out of the aquifer during the test. This means that when the test pump stopped and the water level in the aquifer began to recover, the volume of water to be replaced in the unconfined, dewatered portions of the three zones required that storage of 0.05 to 0.3 cubic feet of porosity per cubic feet of aquifer volume had to be replaced by flow out of the confined part of the aquifer providing approximately 0.0001 cubic feet of water per foot of head change, divided by the saturated thickness of the aquifer. Accordingly, a specific storage in the 0.00001 (1.0×10^{-5}) range was filling void volumes in the range of 0.05 to 0.3. Obviously, the much smaller specific storage volume would take a lot longer to fill up the much larger specific yield volume than it took to pump the water out of the specific yield storage. Therefore, the water levels in the cones of depression in the uppermost three water-bearing zones in the Sleepy Hollow No. 6 well did not fully recover when t/t' = 2.0. Diagnostic plots of the test are discussed below.

Figure 5-21 is a plot of arithmetic drawdown versus square root of pumping time. The plot is limited to the first 12 minutes of drawdown and the corresponding values of the recovery curve. The data plot as curved lines, not straight lines, thus indicating aquifer response is that of radial flow.

Figure 5-21
Linear Flow Diagnostic Plot for Sleepy Hollow Well No.6 Shows Radial Flow

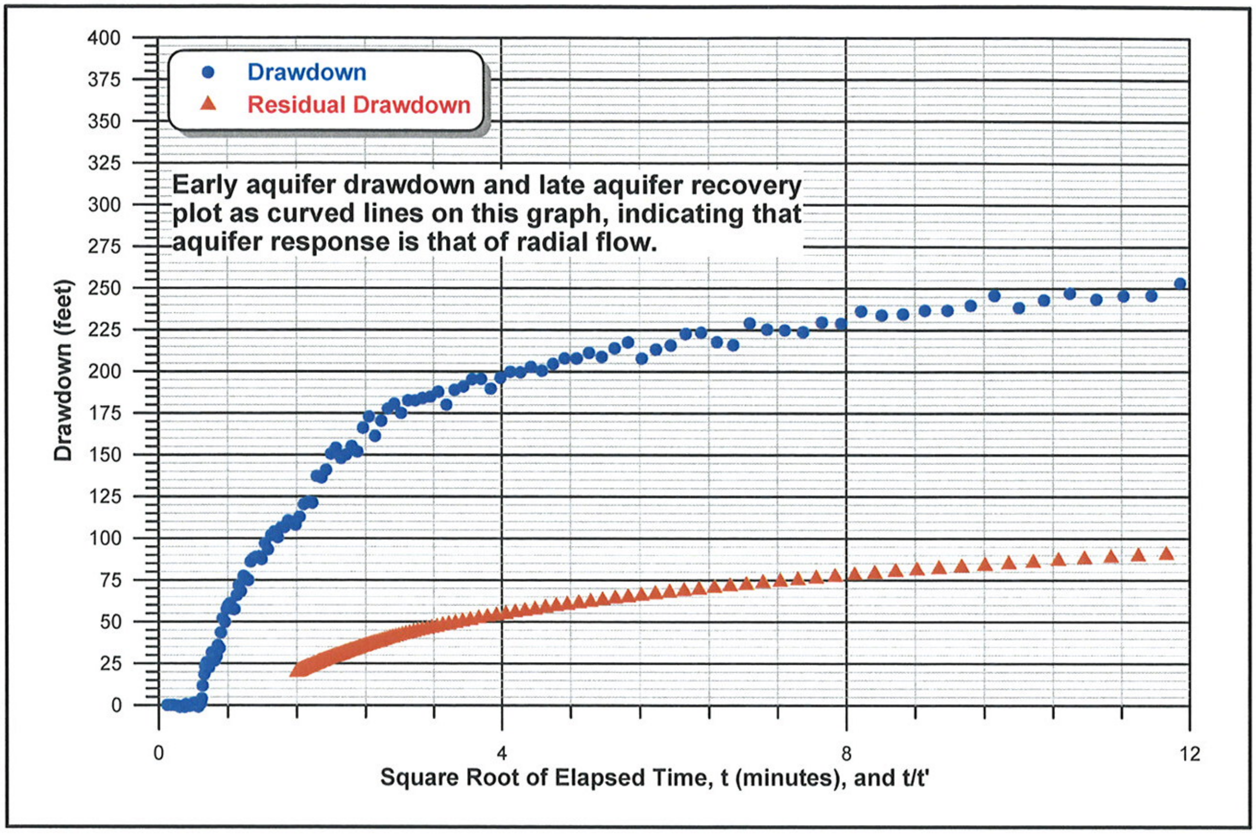
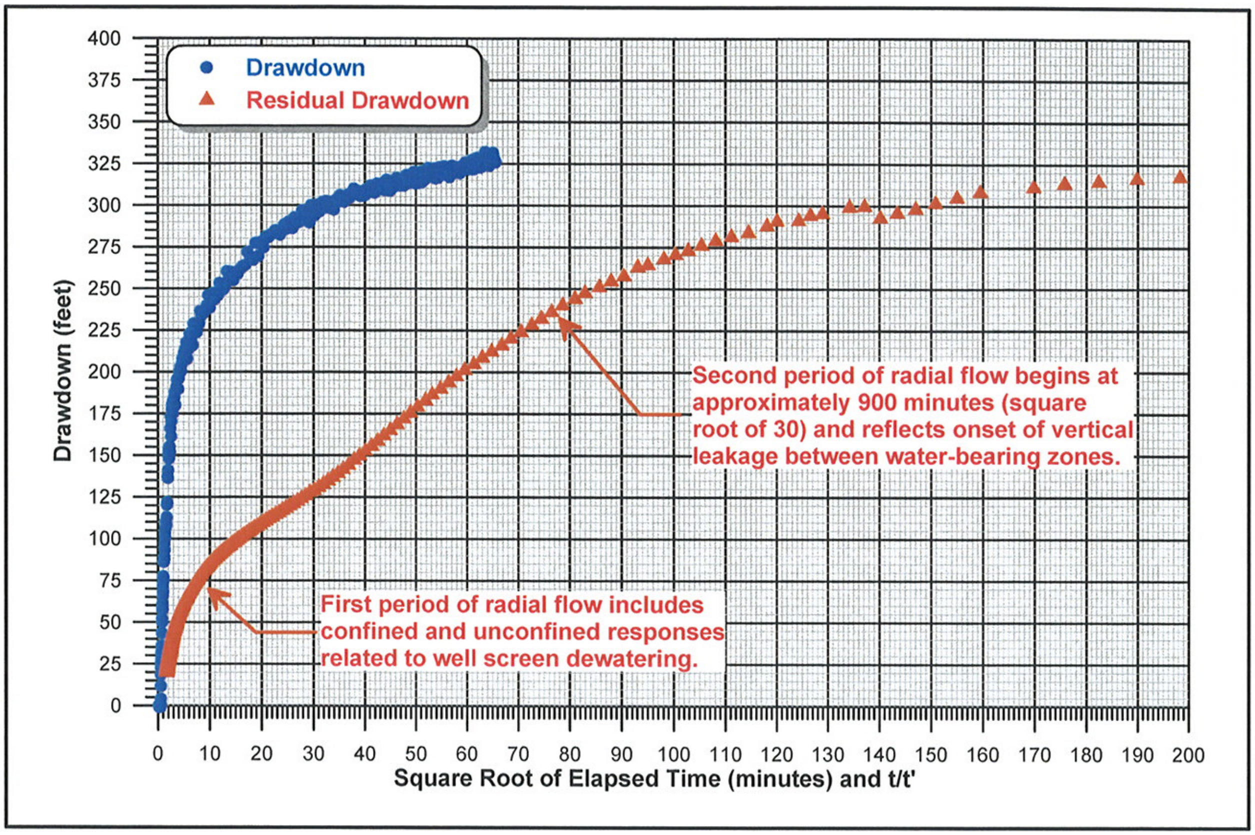


Figure 5-21 reflects a significantly different response than observed at the ASR Test Well site and one initially more favorable to long-term sustainable development of groundwater because of the absence of a negative boundary effect.

Figure 5-22 expands the diagnostic plot on Figure 5-21 to include data out to a square root time value of 200. The recovery curve clearly reflects two distinctly different periods of radial flow response. The first part of the curve reflects radial flow before and after the onset of dewatering of the uppermost well screens and associated release of unconfined storage. The second part of the curve reflects a decrease in the rate of drawdown after a period of transition from the early curve with the transition taking place over the 900- to 1,000-minute interval

Figure 5-22
Expanded Diagnostic Plot for Sleepy Hollow No. 6 Well Test



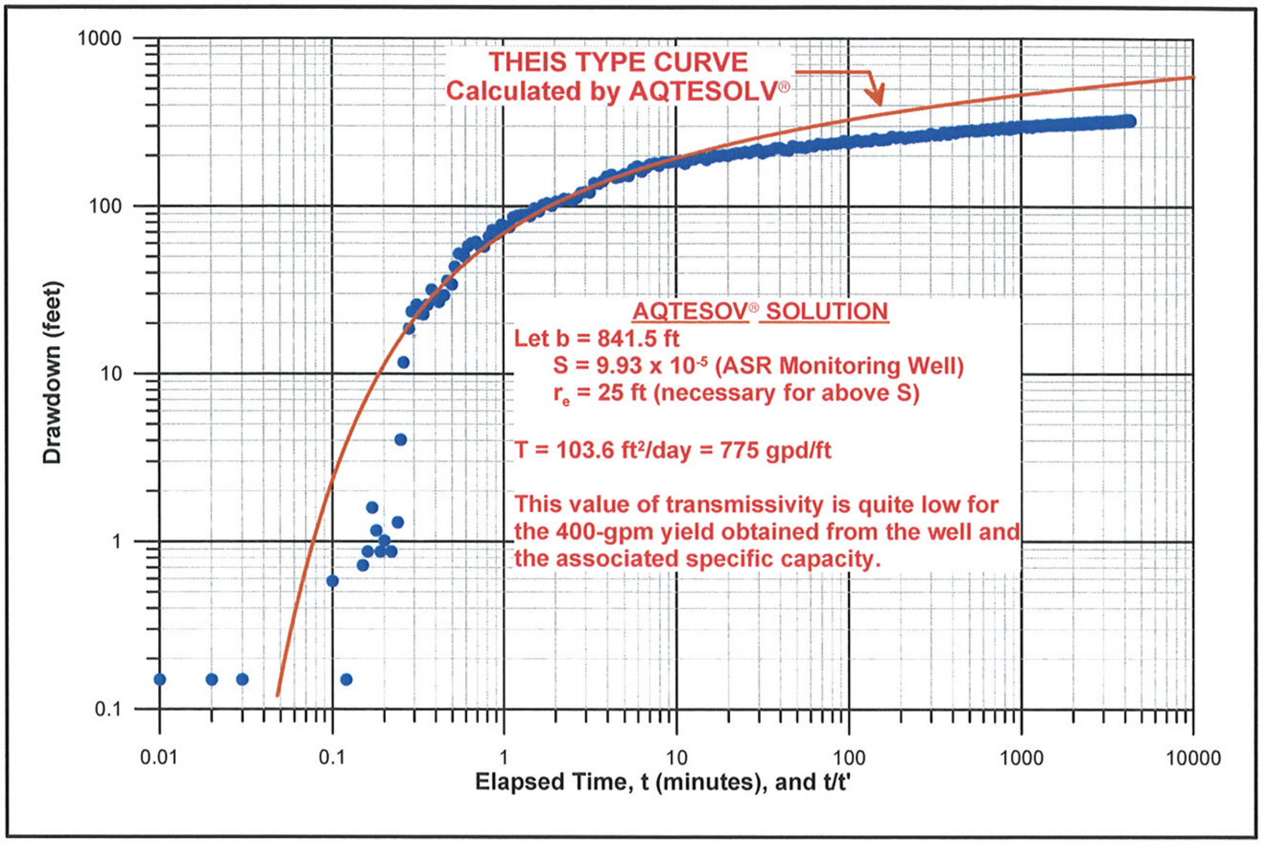
The latter decrease in the rate of drawdown begins as the dewatering of the second deepest well screen (the 866.6 to 896.6-foot interval) approaches completion and the water level continues downward toward the top of the next screened interval (902.4 to 912.4 feet). This latter decrease in the rate of drawdown is described by Brown and Caldwell (2005) "delayed head response" after 1,000 minutes caused by vertical leakage between water-bearing zones. If the latter conclusion were true, continued pumping would have resulted in stabilization of the pumping water level in the well. The fact stabilization did not occur rules out vertical leakage as an explanation for the change in the rate of drawdown after about 1,000 minutes.

In fact, the remainder of the test involves the water level in the well slowly decreasing to about 911 feet, which effectively dewaters the face of the third deepest water-bearing screened interval in the well. The net effect of the two decreases in the rate of drawdown associated with the change from confined to unconfined release of storage at the borehole face in the well is to substantially reduce the amount of drawdown that would have occurred under confined conditions. However, it also causes the rate of recovery to be slower than would happen under completely confined conditions, resulting in the 16.143 feet of residual drawdown remaining at t/t' = 2, when the recovery of the water level should have been complete.

Comparison of the test data to the Theis type curve on Figure 5-23 supports several conclusions about the aquifer response to pumping. The initial drawdown rate prior to one minute of pumping is very rapid, likely reflecting an initial pumping rate much greater than the subsequent 400-gpm rate used for the constant rate test. The initially high pumping rate may have been caused by filling of that part of the pump column above the static water level at 525.28 feet. The same initial data exhibit data scatter most likely caused by degassing and associated turbulence in the water column in the well.

From 0.3 to 8.0 minutes of pumping time, the data conform reasonably well to the Theis type curve. This portion is the only part of the data conforming to the Theis type curve and the data fit a part of the type curve where drawdown is changing rapidly; therefore, a Cooper-Jacob, semi-logarithmic straight-line solution for aquifer transmissivity is inappropriate for this part of the data. The Theis type curve on Figure 5-23 is generated with AQTESOLV® software based on a saturated thickness of 841.5 ft (Brown and Caldwell, 2005), a fixed storativity of 9.93×10^{-5} as determined at the ASR test site, and a hydraulic radius (r_e) of 25 feet as was determined by trial and error to make the type curve fit the data in the range of 0.2 to 10 minutes.

Figure 5-23
Tentative Theis Solution to Sleepy Hollow No. 6 Well Test



The value of transmissivity derived from the data fit to the transient Theis type curve for non-equilibrium drawdown is 775 gpd/ft; a value that is quite low for this aquifer and for the yield of the well. This low value suggests that the energy gradient in the well during this interval of the test had not developed sufficiently to include substantial production of groundwater from all of the water-bearing sandstone zones exposed to the well screens, a conclusion also reached by Brown and Caldwell (2005), and therefore the low transmissivity value applies only to one or a few water-bearing zones, not the collective sandstone thickness penetrated by the well.

After approximately 8 minutes pumping time, the onset of screen dewatering effects and associated release of unconfined storage causes the time-drawdown curve to depart from the Theis type curve and follow a lesser rate of drawdown associated with the release of groundwater from unconfined storage. Analytical models exist to interpret aquifer response that involves cones of depression with unconfined conditions near the well and confined conditions more distant from the well, all in the same cone of depression; however, no attempt to apply these methods is made herein due to the potential multiple-aquifer or multiple water-bearing zone nature of this well and the fact that unconfined conditions are associated with only the uppermost 3 of 14 screened intervals in this well. Moreover, Brown and Caldwell (2005) believe that only the uppermost nine screened intervals contributed flow to the well during the pumping test.

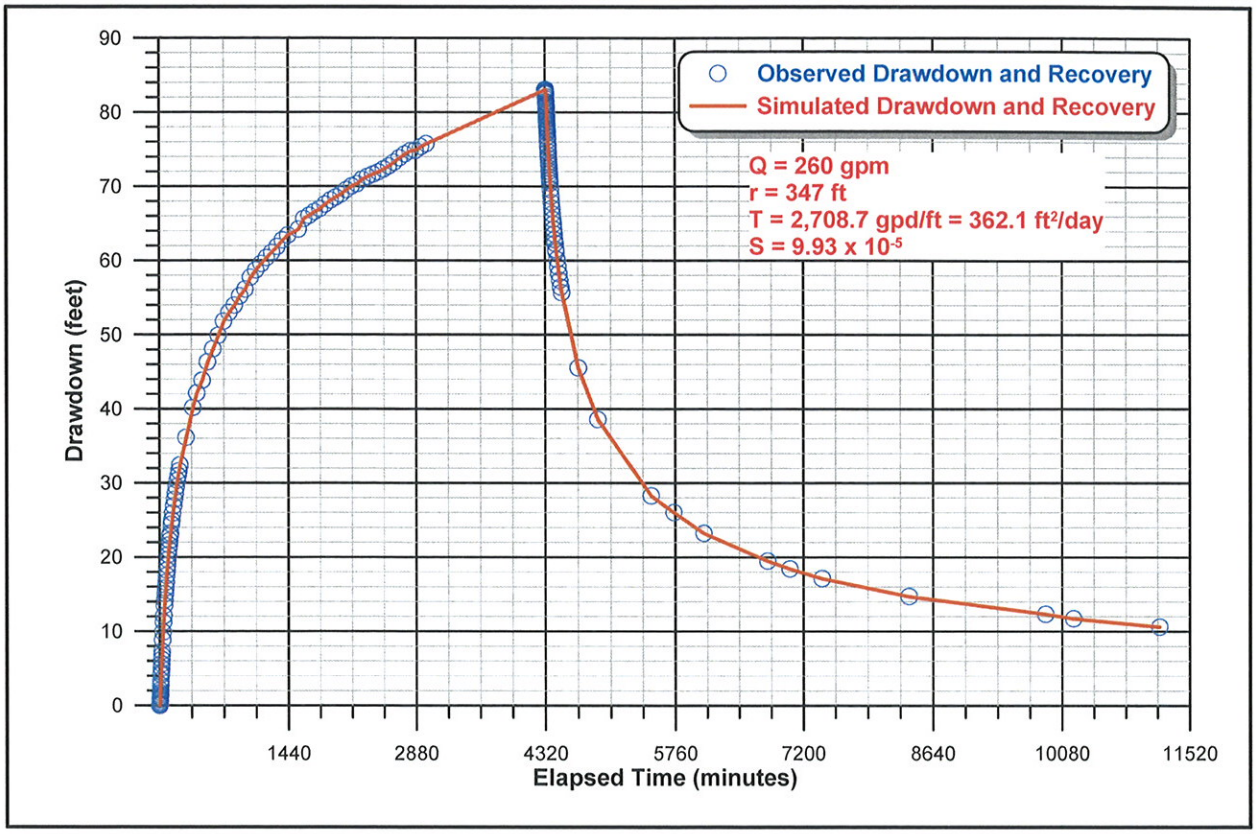
The foregoing considerations show that so many different factors complicate the time-drawdown response observed in the Sleepy Hollow No. 6 well that it is probably not possible to obtain a correct interpretation of the aquifer transmissivity from the test. The Brown and Caldwell (2005) solution of transmissivity equal to 315 ft^2/day (2,357 gpd/ft) and possibly as high as 420 ft^2/day (3,142 gpd/ft) are based on Cooper-Jacob solutions applied to time-drawdown responses that have been flattened by partially unconfined conditions in some of the water-bearing zones and which are therefore not solutions this report can agree with. The transmissivity values obtained from the Cooper-Jacob solution are directly dependent on the slope of the semi-log time-drawdown line. Therefore, an erroneously large value of transmissivity is obtained from the Cooper-Jacob solution where the time-drawdown slope is decreased by some factor such the unconfined aquifer response in this test. Likewise, the transient Theis non-equilibrium solution of 775 gpd/ft shown on Figure 5-23 is erroneously low for the reasons discussed above. The actual transmissivity is probably closer to the Brown and Caldwell (2005) solution than to 775 gpd/ft, but smaller than the Brown and Caldwell (2005) solution.

Therefore, the most important aspect of the two aquifer tests reviewed herein is the fact that neither of the tests exhibit full recovery of the cone of depression around the pumped wells during a recovery period equal to the pumping period. The presence of significant residual drawdown in both of these wells when t/t' = 2 indicates the water levels in the wells will not fully recover between pumping cycles unless the recovery periods significantly exceed the durations of the pumping periods. For example, the ASR Test Well was pumped for 3 days (72 hours). After 4.77 days of recovery, 114 hours and 25 minutes, 10.3 feet of residual drawdown remained. The projection of the recovery curve for the Sleepy Hollow No. 6 well shows more than 16 feet of residual drawdown remaining when the recovery time equals the pumping time (Figure 5-20).

As discussed above, there are too many complicating factors involved in the response at the Sleepy Hollow No. 6 well to attempt a reliable model projection of future drawdown. However, a model can replicate the conditions observed at the ASR test site. Therefore, the next question is how do negative boundary conditions potentially affect groundwater level declines associated with long-term cyclic pumping of wells in this part of the Fort Union aquifer? The first step in examining the effect of negative boundary conditions is to compile an analytical model that can reproduce the observed drawdown and recovery curves for a specific pumping rate. This step was accomplished by application of commercial AQTESOLV® software to apply the Theis solution incorporating no-flow boundaries in the aquifer. Figure 5-13 shows excellent correlation between the drawdown predicted by the AQTESOLV® solution and the observed field test data, thus demonstrating that the . AQTESOLV® solution is well-calibrated to the observed aquifer response.

The AQTESOLV® solution on Figure 5-24 assumes parallel no-flow boundaries, one 950 feet to one side of the pumped ASR Test Well and one 10,100 feet to the other side of the pumped well. The ASR Monitoring Well (observation well) is 347 feet from the pumped well towards the boundary at 950 feet on a line perpendicular to the boundaries and through the pumped well. The no flow boundaries are extended 40,000 feet past the pumped well in each direction for an 80,000-feet long strip aquifer, centered on the pumped well, with no boundaries at either end. Although it is recognized the well penetrates multiple water-bearing zones likely to offer a range of hydraulic properties, the latter no-flow boundary geometry provides a good reproduction of the observation well response and is thought to be analogous to the width of the area of thickest total sandstone content in the Fort Union around Gillette as identified by Wester-Wetstein (1999).

Figure 5-24
Calibration of AQTESOLV® Solution to ASR Monitoring Well Test Data



After the foregoing calibration process, the AQTESOLV® model was used in a forward solution to project anticipated drawdown that would result from pumping the ASR Test Well for 20 years, using a pumping schedule of 12 hours on alternating with 12 hours off, at 260 gpm (Figure 5-25). Therefore, the model assumes that the well is pumped 12 hours every day for 20 years and is allowed to recover 12 hours every day between pumping cycles. Two simulations were performed, one without boundary conditions and one with the previously described no-flow boundaries used to calibrate the model to the test data.

Figure 5-25 shows projected water level trends at the end of each 12-hour recovery period (pump off drawdown) and at the end of each 12-hour pumping period (pump on) with and without boundary effects. The hydraulic parameters applied in the model are those shown on Figure 5-24.