PRELIMINARY DRAFT CASE STUDY REPORT
TO THE
60TH LEGISLATURE WATER POLICY INTERIM COMMITTEE

by the

Montana Bureau of Mines and Geology

June 10, 2008

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RECOMMENDATIONS TO THE WATER POLICY INTERIM COMMITTEE

We provide recommendations within the report where they relate to the method or technique of evaluating stream depletion or mitigation, and we will continue to develop specific recommendations related to the case studies. However, we offer the following recommendations on a general but nonetheless important level.

1. The State should resist efforts to oversimplify stream depletion; it should not be treated as single value for each well. Stream depletion is not instantaneous, nor does it occur at a steady rate; it may take days, months, years, or even decades to reach its maximum. Stream depletion affects various stream reaches differently and depends on when, where, and how much water is extracted from aquifers by wells or other means. Stream depletion is not a single value for each well—it is a process that must be described with respect to both time and space.

2. The State should resist efforts to oversimplify the concept of offsetting stream depletion; it should not be treated as a single-value target. The offset of depletion is not the simple inverse or opposite of stream depletion; methods to offset depletion must be evaluated with respect to timing, location, and rate. A robust evaluation of stream depletion could lead to opportunities to offset pre-existing stream depletion or even improve aquatic ecosystems. Stream depletion offset is not a single value, but a process that must be described with respect to both time and space.

3. The evaluation of individual and cumulative stream depletion as well as methods to offset depletion is best done through investigations at a sub-basin scale. The investigations should extend beyond the property boundaries of individual applicants, especially with regard to mitigation, aquifer recharge, or other methods of offsetting stream depletion. These investigations should be aggressive, 1- to 3-year efforts that address issues specific to individual sub-basins. The investigations should focus on the science, should be practical not academic, and should include both natural and man-made changes in the hydrologic system. The level of effort to complete sub-basin investigations goes beyond the abilities of individual water appropriators and should be a part of a primary water management function assumed by government.

4. The Water Policy Interim Committee should continue beyond this interim. Changes in the regulatory environment that will be necessary to adaptively or conjunctively manage water within all of Montana's basins will not only need close legislative guidance and support, but stakeholder involvement to identify needs and concerns of inhabitants, water users, and the environment.

SECTION 1: GENERAL CONCEPTS OF STREAM–AQUIFER INTERACTION AND INTRODUCTION TO THE CLOSED BASIN AREA

by John LaFave, Associate Research Hydrogeologist

INTRODUCTION

With the passage of House Bill 831 (HB 831), the 60th Montana Legislature directed the Montana Bureau of Mines and Geology (MBMG) to study the closed basins in Montana and assess the range of potential impacts of ground-water development on surface flows.

Ground water and surface water are an interconnected resource. To date, 97 percent of the water withdrawals in Montana (and the closed basins in particular) have utilized surface water (Cannon and Johnson, 2004). Because ground-water withdrawals are minor relative to surface-water use, the interconnection between surface water and ground water has not previously presented widespread management challenges. However, basin closures, drought, and escalating demand have resulted in increased pressure to develop ground-water resources. This effort has led to water conflicts and legal action, especially within the closed basins. In 2006, the Montana Supreme court ruled that impacts to surface water and senior water-right holders from ground-water withdrawals must be evaluated (Trout Unlimited vs. DNRC, 2006). As a result the Department of Natural Resources and Conservation (DNRC) curtailed issuing any new ground-water appropriations in closed basins. The 60th Legislature approved HB 831 to provide a means for the DNRC to process ground-water appropriations, but recognized that there are challenges and limitations to developing ground water—especially in closed basins.

The management challenges facing Montana policy makers, as outlined in HB 831, revolve around how to develop ground water in a manner that (1) "protects the prior appropriation doctrine" while at the same time (2) protects "the quality of Montana's water" and (3) causes "the least possible degradation to the State's natural ecosystems" (language from HB 831). Hydrogeologic information and analysis are needed to meet these challenges and promote sound management and policy decisions.

Effective management requires a good understanding of the interaction between ground water and surface water. Depending on local factors and the scale of the drainage basin (or area of interest), these interactions can take different forms, affecting the quantity and quality of the water in both resources. The interactions can be technically challenging to quantify, and the magnitude, timing, and location of the impacts can be difficult to accurately predict. Furthermore, the effect of ground-water development on stream flow can be highly site-specific and will vary between basins depending upon each basin's unique hydrology and hydrogeology. The difficulty of measuring impacts and the heterogeneity of conditions create an uncertain management framework.

This document will review our current understanding of how ground water and surface water interact, summarize hydrogeologic conditions in the closed basins, and present specific information from "case studies" in the Beaverhead, Bitterroot, and Gallatin Valleys. This analysis will highlight the range of geologic and hydrogeologic conditions that occur and lay a foundation for understanding the range of potential impacts to be expected from ground-water development. Finally, the adequacy of the hydrogeologic assessment specified in HB 831 that applicants for ground-water appropriations must perform will be evaluated.

The Hydrologic Cycle

The hydrologic cycle describes the endless circulation of water among the ocean, atmosphere, and land. Sustainable water-resource development requires an understanding of how climate causes temporal and spatial variations in the cycle, and how geology modifies rates of flow and volumes stored in the subsurface. The basic inputs and outputs of the hydrologic cycle on a typical Montana-basin scale are shown schematically in figure 1 (hydrocycle). Energy from the sun powers the system, evaporating water from the oceans and the land surface which then condenses to form clouds. Water from clouds returns to the land surface in the form of rain or snow. Most precipitation returns to the atmosphere as evaporation from the land surface and as transpiration from plants. The process of evaporation and transpiration combined is referred to as evapotranspiration (ET). Some precipitation generates surface runoff to streams and ultimately flows back to the oceans. A final portion of the precipitation, the smallest fraction, infiltrates through the soil to the water table and becomes ground water. The movement of water from the land surface to the water table is called ground-water recharge. The amount and rate of recharge varies widely depending on the surficial geology and soil properties, topography, climate, land use, and the duration and intensity of precipitation.

Occurrence of Ground Water

Ground water is water that fills pore spaces, crevices, or fractures in soil, sediment, or rock. Ground-water occurrence can be difficult to visualize and understand because it occurs out of sight and moves at unfamiliar time scales. The amount of pore space (porosity) that water can occupy in geologic material is highly variable; unconsolidated sand and gravel may have porosities as high as 35 percent, whereas poorly sorted deposits with a high fraction of silt or clay will be significantly less. Some types of bedrock, for example unfractured granite, will have no porosity. The degree of connection between the pore spaces (permeability) determines the ease with which a material can transmit water. Ground water can move easily through well-sorted gravel deposits because they have a high degree of porosity and permeability—the open space is well connected. It is much more difficult for water to move through poorly sorted, fine-grained deposits that include large amounts of silt or clay because there is little open space and it is poorly connected.

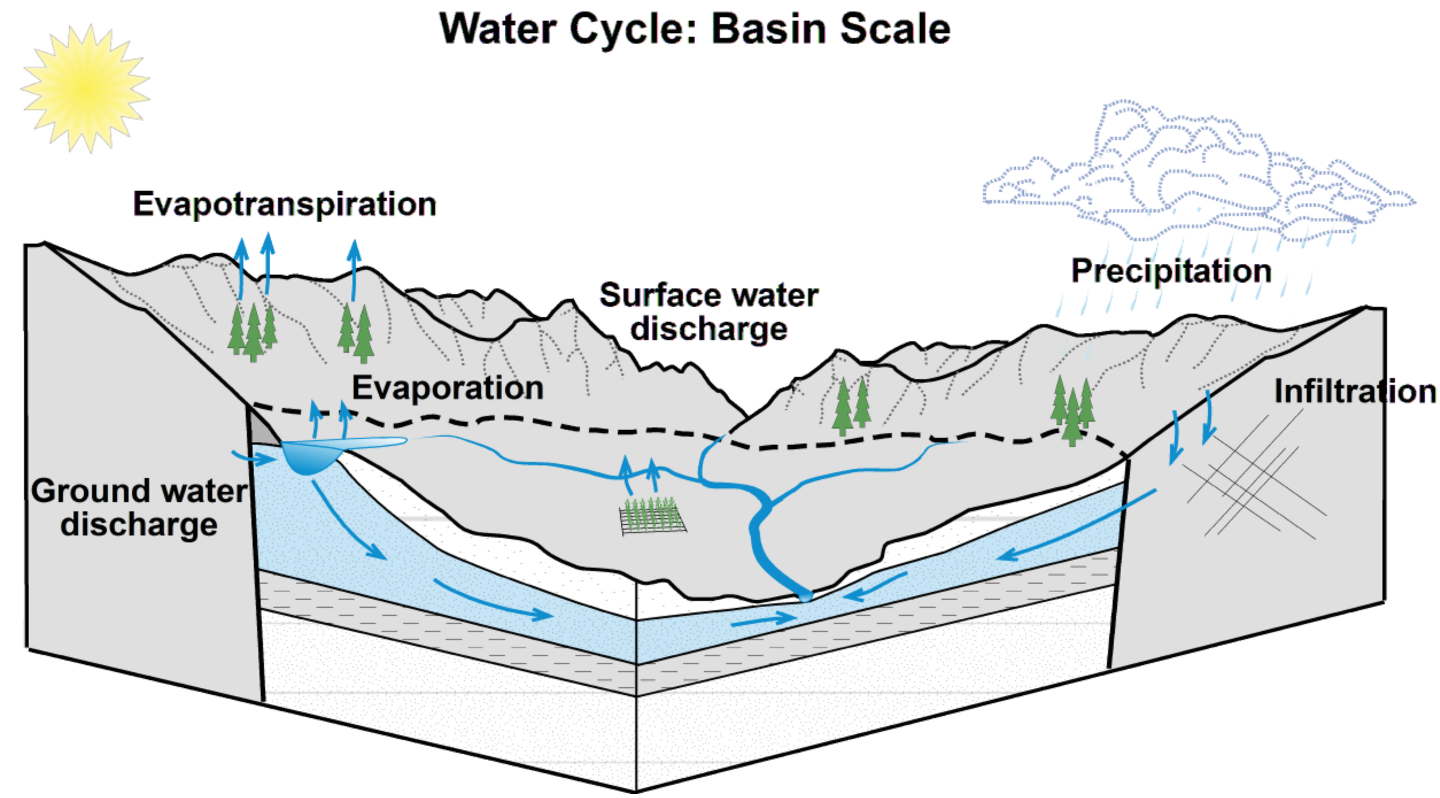


Figure 1. Schematic representation of the hydrologic cycle on a typical Montana basin scale.

Aquifers are permeable geologic units that store and transmit usable quantities of ground water. Aquifers provide two important functions: (1) they transmit water through the subsurface from areas of recharge to areas of discharge; and (2) they provide a storage reservoir for ground water. Ground water occurs either under unconfined (water table) or confined (artesian) conditions (fig. 2). In unconfined aquifers the water table represents the upper boundary of the aquifer; below the water table all the pore spaces are saturated with water, and above the water table pore spaces are filled with air and water. The water table moves upward and downward in response to water entering (infiltration/recharge to) and leaving (discharge from) the aquifer. The water level in a well completed in an unconfined aquifer will equilibrate with the water table surface. Unconfined aquifers yield water to wells by draining the pore space in the area adjacent to the well. Unconfined alluvial aquifers are generally shallow (within 100 ft of the land surface) and occur adjacent to the major streams in all the closed basins; many unconfined aquifers have direct hydraulic connection to surface water.

Confined or artesian aquifers are permeable geologic units that are: (1) completely saturated and (2) overlain or "capped" by relatively low-permeability layers such as clay or silt (confining layer). The water in confined aquifers occurs under pressure; thus, the water level in a well completed in a confined aquifer will rise above the confining layer (the top of the aquifer). The level to which water will rise in wells completed in a confined aquifer is referred to as the potentiometric surface. In flowing artesian wells the pressure in the aquifer is sufficient to raise the water level above the land surface. Confined aquifers do not yield water to wells in the same manner as unconfined aquifers. When artesian aquifers are pumped the water is released by compression of the aquifer material and expansion of the water near the pumped well; the aquifer is not drained as in the case of an unconfined aquifer. Confined aquifers occur buried at depth in many of the closed basins.

Ground water flows through aquifers towards discharge points such as rivers, wetlands, springs, and lakes. A ground-water flow system therefore consists of that part of the hydrologic cycle where water is flowing below the land surface from areas of recharge to areas of discharge. Water is constantly added to the system by recharge, and water is constantly leaving the system as discharge to surface water and as evapotranspiration. Under natural conditions aquifers are in a state of equilibrium, which means that during long time periods the amount of water flowing into an aquifer from recharge is balanced by outflow as discharge and evapotranspiration. Although ground water is flowing, it moves much more slowly than does surface water. In shallow, permeable aquifers ground water might flow on the order of 1 ft/day, and in confined aquifers the flow rate might be on the order of 1 ft/year. Surface-water flow rates are on the order of 1 ft/second.

Ground-water levels (as measured in wells) in an aquifer reflect the balance between ground-water recharge and discharge and related changes in aquifer storage. When recharge exceeds discharge, water

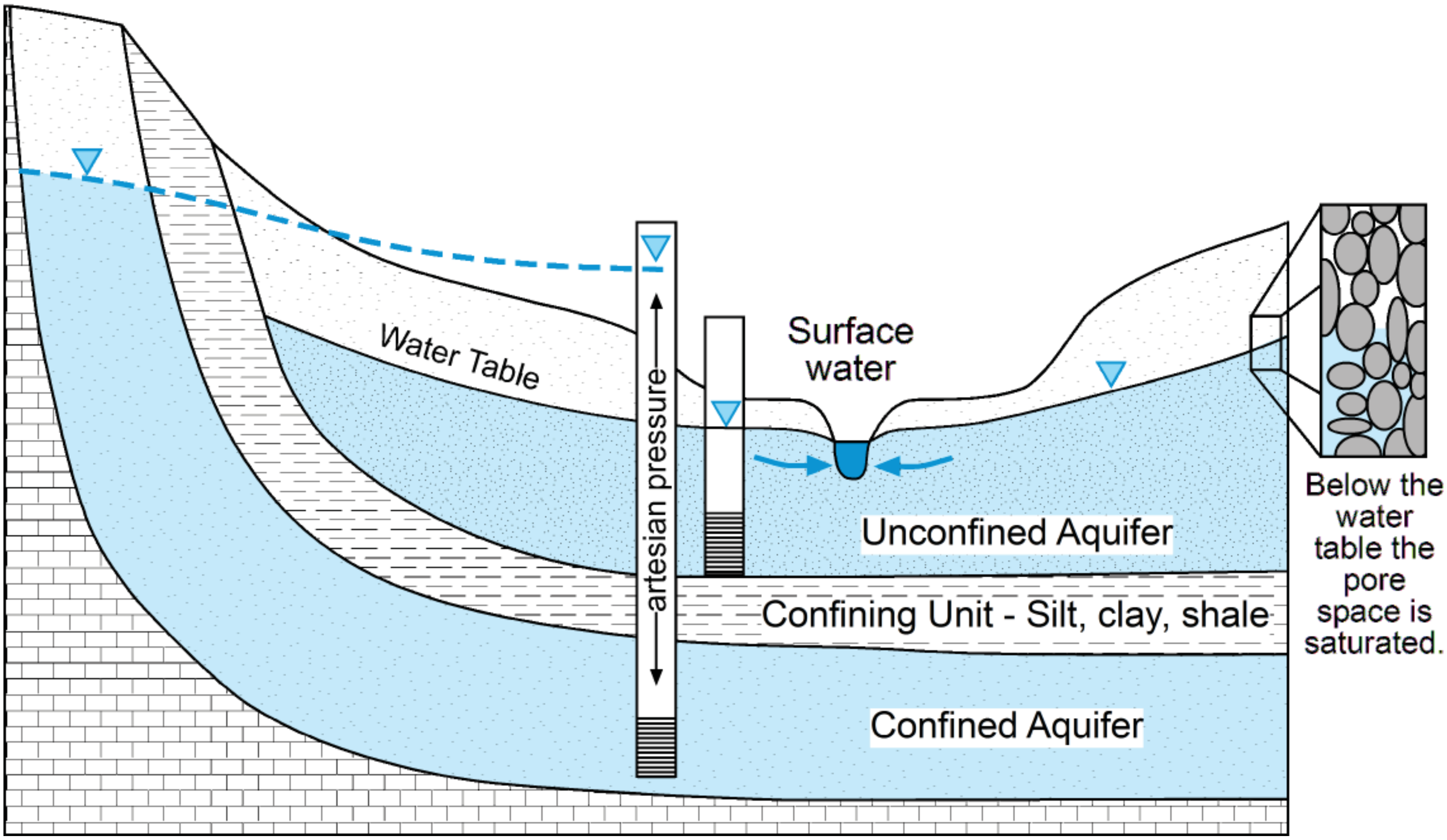


Figure 2. Ground water occurs in unconfined and confined aquifers. The water table is the upper surface of an unconfined aquifer. Confined aquifers are buried below less permeable layers and the water is under pressure.

levels rise—more water is stored in the aquifer; when discharge exceeds recharge, water levels decline—less water is stored in the aquifer. Typically, water levels will be higher in the spring, reflecting recharge from spring runoff, and lower during winter months when recharge is minimal. When recharge and discharge are balanced over the long term, steady-state conditions exist; water levels will fluctuate seasonally but the long-term average level is constant. Climatic, land-use, or water-use changes or ground-water development can disrupt the steady-state condition so that water levels either rise or fall over the long term. Depending upon the scale of the development/change, the impact might affect only a very localized area, or it might show a regional effect in an extensive aquifer. Should the climate or other change be temporally extensive, the aquifer will adjust to a new steady state with different values for storage, recharge, and discharge.

Stream–Aquifer Interaction

Nearly all surface-water features (streams, lakes, springs, wetlands, and reservoirs) interact with ground water. The interactions of surface water and ground water are governed by: (1) the position of the surface-water bodies relative to the ground-water flow system, (2) the permeability of the streambed and underlying materials, and (3) the climatic setting. In stream–aquifer systems, a stream is considered to be gaining if ground water flows from the underlying aquifer into the stream. For this to occur, the water table elevation adjacent to the stream must be higher than the surface of the stream (fig. 3). The steady flow of ground water into a stream is called base flow. On an annual basis, base flow (ground-water discharge) can account for more than 50 percent of stream flow, and it accounts for all the flow during winter months when surface water is locked up in the form of ice and snow (fig. 4).

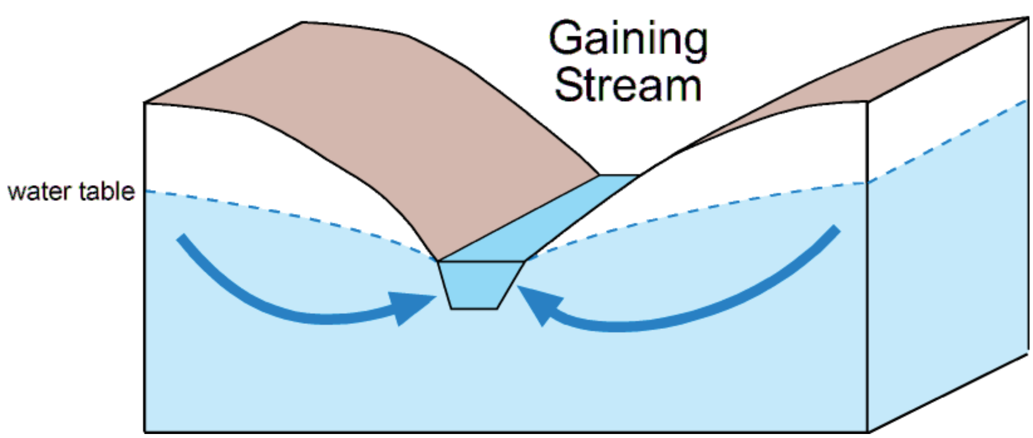


Figure 3. Shallow aquifers discharge water to gaining streams.

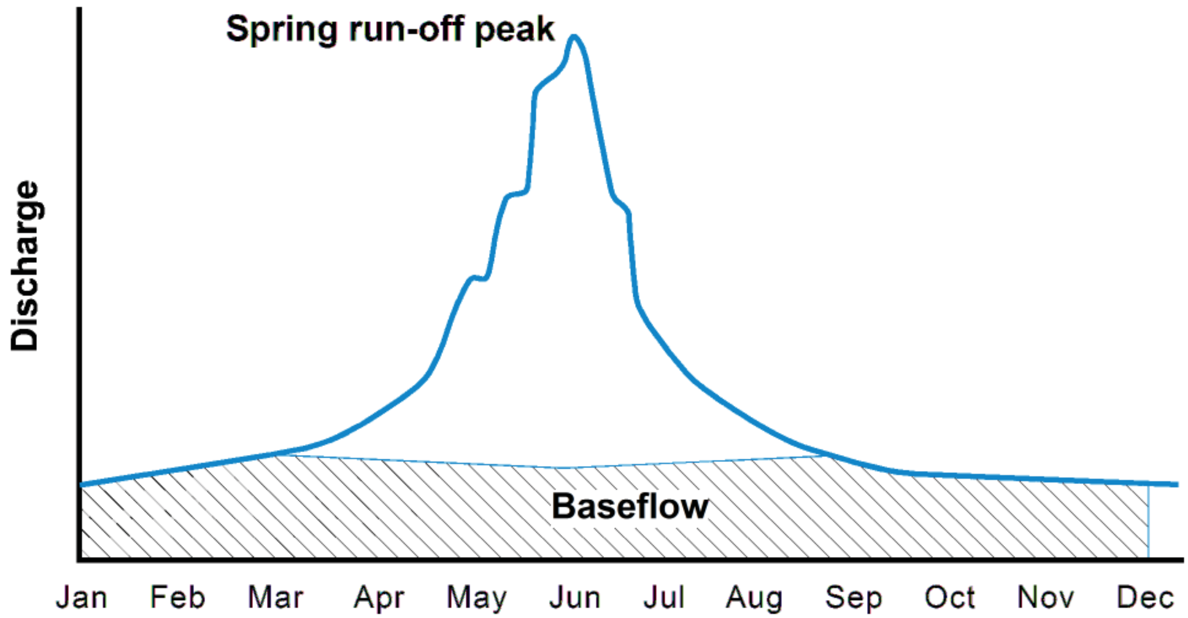


Figure 4. Schematic seasonal hydrograph showing stream discharge against time. Baseflow, the discharge of ground water into a stream, is a major component of total annual stream flow; during dry and winter months streamflow is sustained by baseflow.

underlying Missoula Valley aquifer (fig. 6). Seasonal runoff and precipitation events can change ground-water/surface-water interactions. During spring runoff or times of high-stage flows, water in a stream can rise above the adjacent water table, causing surface water to move into the streambank. After the high flows decrease, water that infiltrated the streambank is gradually released back into the stream.

Not all streams are gaining; some lose water through the streambed to the underlying aquifer. In losing streams (or stream reaches), the surface of the stream must be higher than the underlying/adjacent water table. Losing streams are a source of ground-water recharge. A losing stream can either be connected or disconnected from the underlying aquifer. If the water table intersects the stream it is connected; if the water table is below the base of the stream it is disconnected (fig. 5). In connected streams surface flow can be affected by changes in the aquifer water level or nearby ground-water pumping. In disconnected settings, water-level changes in the aquifer and/or ground-water pumping have little or no effect on stream flow.

The interaction between ground water and surface water is dynamic, and it can vary spatially and temporally. A given stream may have reaches where it is gaining and losing. For example, the Clark Fork River is gaining with the exception of the reach through the Missoula Valley. Between Hell Gate Canyon and its confluence with the Bitterroot River, the Clark Fork River loses water and is the main source of recharge to the

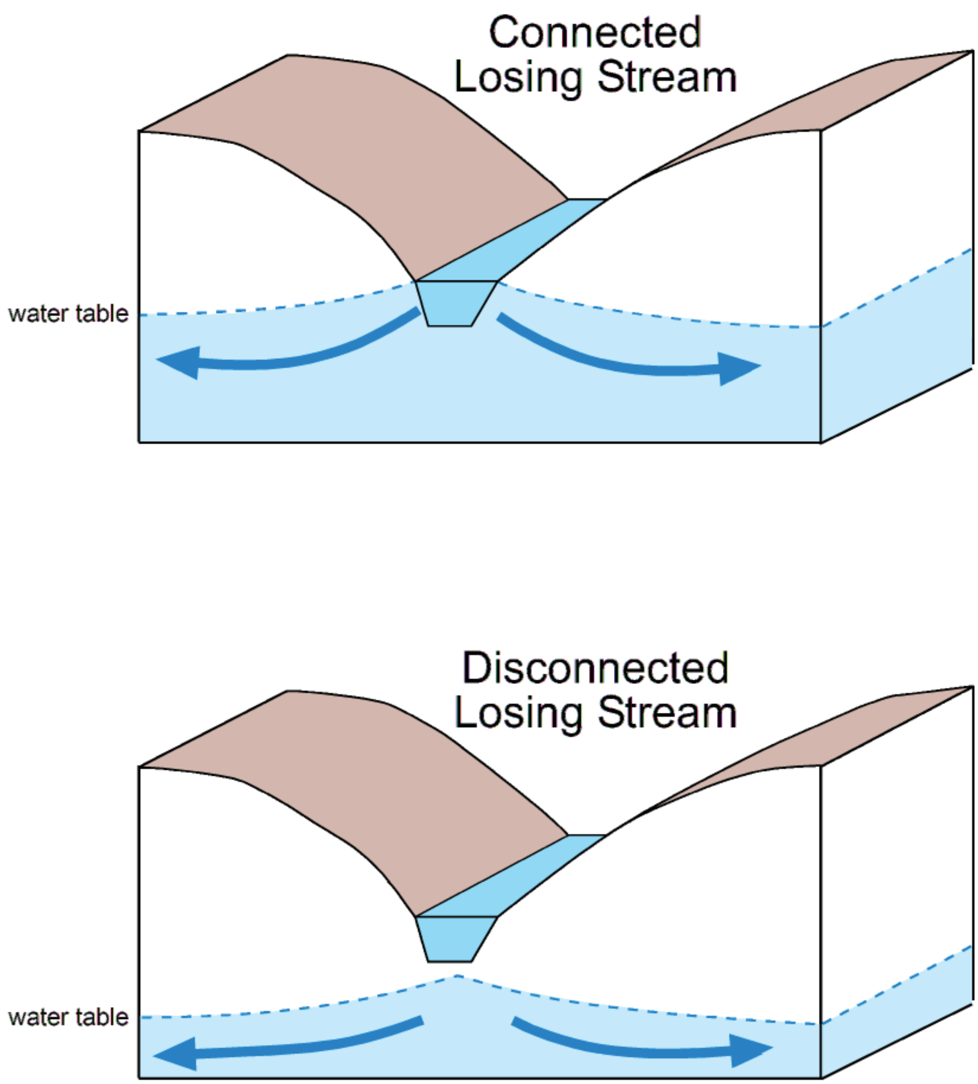


Figure 5. Losing streams lose water to the under ling aquifer; disconnected losing streams are separated from the underlying aquifer by an unsaturated zone.

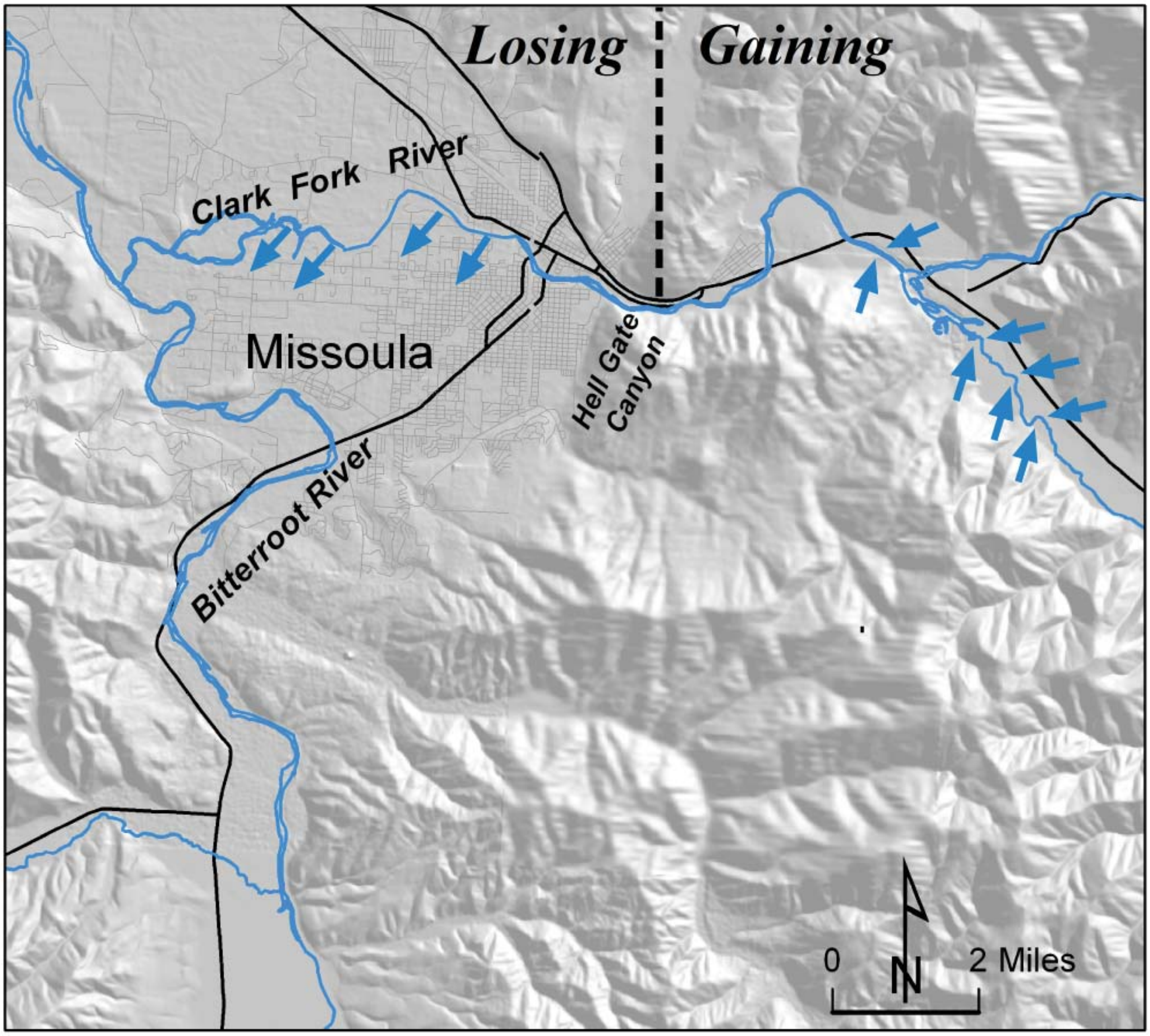


Figure 6. The Clark Fork River is a gaining stream above Hell Gate Canyon and a losing stream between Hell Gate Canyon and its confluence with the Bitterroot River. Arrows indicate flow of ground water adjacent to the stream.

Climate variability can also affect stream–aquifer interactions. Prolonged drought can reduce the amount of ground-water recharge, resulting in significant storage declines. Drought-related reductions in recharge and storage can result in stream-flow reductions and even dry up stream reaches during the hot summer months.

Ground-water development (pumping) can decrease the rate of base flow by intercepting ground water flowing to a stream and/or increase (induce) stream leakage. In an idealized stream–aquifer system where the stream is gaining, ground-water development in the shallow aquifer directly connected to the stream will progressively reduce base flow. Water-level decline in response to pumping will reduce the hydraulic gradient to the stream, thus reducing ground-water flow to the stream. If pumping continues to the point that the ground-water level adjacent to the stream falls below the stream level, then the stream becomes losing and water is induced from the stream into the aquifer (induced recharge). Figure 7 illustrates the transition from gaining to losing in response to prolonged ground-water withdrawals. There may be a considerable time lag between the start of pumping and any reduction in stream flow depending upon the location of the pumping well (distance and depth) relative to the stream, the hydraulic characteristics of the aquifer, and the pumping rate. Furthermore, the effect of ground-water pumping on stream flow may persist long after pumping has stopped. This is a simplified scenario; in the real world there will be other hydrogeologic factors such as ET, recharge variability, the presence of disconnected streams or reaches, low-permeability streambeds, and deep confined ground-water systems that complicate the stream–aquifer interactions.

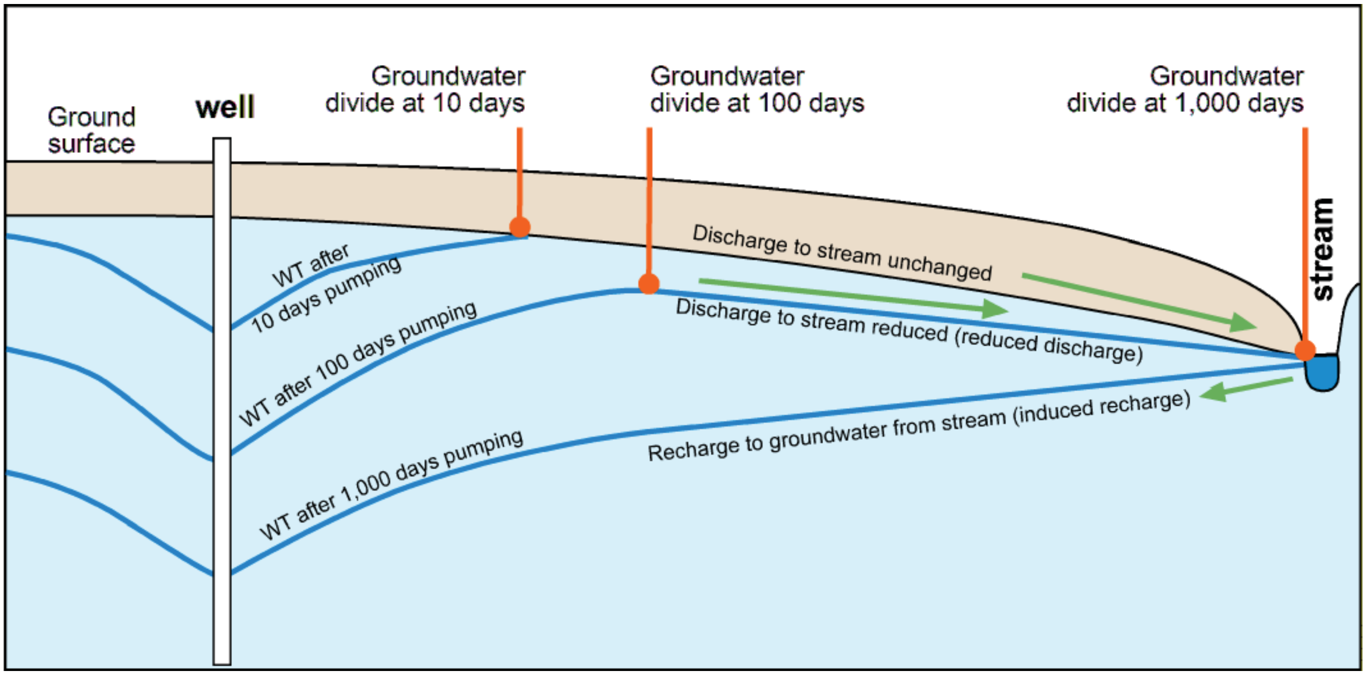


Figure 7. Schematic diagram illustrating the effect of pumping ground water from a shallow aquifer near a gaining stream. If pumping is at a high enough rate or continues long enough, the stream reach near the well may transition from a gaining to a losing stream.

CLOSED BASIN REGIONAL SUMMARY

The closed basin area, as identified in HB 831, encompasses five major drainage basins (Teton, upper Missouri, Jefferson–Madison, upper Clark Fork, and Bitterroot) that cover a 31,900 square-mile area in the central–southwestern part of Montana (fig. 8). It includes parts or all of Teton, Chouteau, Cascade, Lewis and Clark, Broadwater, Silver Bow, Deer Lodge, Beaverhead, Gallatin, Madison, Jefferson, Granite, Powell, Missoula, Ravalli, and Meagher counties. The closed basins span the Continental Divide, and include drainages in the upper Clark Fork and upper Missouri River systems. The upper Missouri, Jefferson–Madison, and upper Clark Fork basins contain multiple sub-basins (fig. 8).

The closed basin area occurs within the northern Rocky Mountain intermontane basin and northern Great Plains physiographic provinces (fig. 9). Most of the basins (Bitterroot, upper Clark Fork, Jefferson–Madison, part of the upper Missouri) are characterized by broad valleys that are separated by mountain ranges. The valleys are drained by main-stem perennial streams. Mountain tributary streams drain the upland headwater areas. The northern part of the upper Missouri and the Teton basins are bounded by the Rocky Mountain front on the west and the Little Belt Mountains to the south, but most of the area is characterized by flat to rolling prairie.

The rivers in the closed basins are an important source of water for public supply, agricultural, hydropower, and industrial uses. Surface water dominates the water withdrawals (fig. 10); an estimated 4,434 million gallons per day (MGD) of surface water is withdrawn for use as compared to 86 MGD of ground water (Cannon and Johnson, 2004). Most of the surface water has long been appropriated for agricultural use, primarily irrigation. The irrigated lands are generally along or within a few miles of major streams. Surface water is diverted from streams or reservoirs and is transported through canals to the fields (fig. 11).

Ground water in the closed basins is obtained primarily from wells completed in unconsolidated basin-fill aquifers that consist mostly of sand and gravel, and from wells completed in sedimentary-rock aquifers, chiefly sandstone and limestone. Some wells withdraw water from volcanic rocks and fractured meta-sedimentary rocks; however, well yields from these aquifers are generally low. Most wells in closed basins are reportedly used for domestic purposes (so-called exempt wells), and the number and location of wells are directly related to the distribution of people—the densely populated valleys contain the greatest number of wells (fig. 12). However, irrigation is the largest user of ground water (fig. 13): of the estimated 86 MGD of ground-water withdrawals, 60 percent is for irrigation; municipal use accounts for 23 percent and private domestic wells account for 11 percent; and stock watering and industrial use accounts for the remainder.

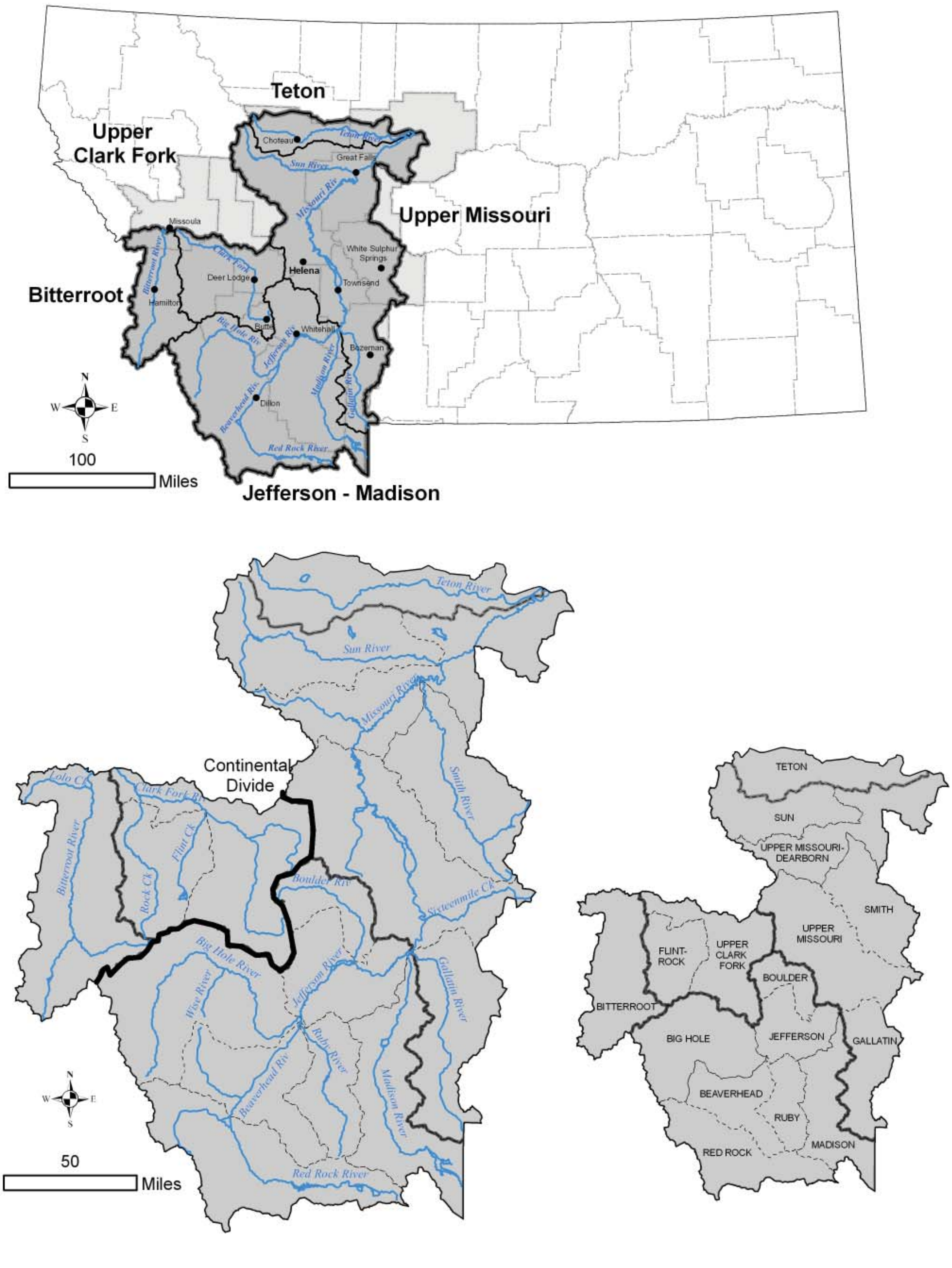


Figure 8. The five closed basins cover a large part of central and southwest Montana. With the exception of the Bitterroot and Teton, the basins have several sub-basins.

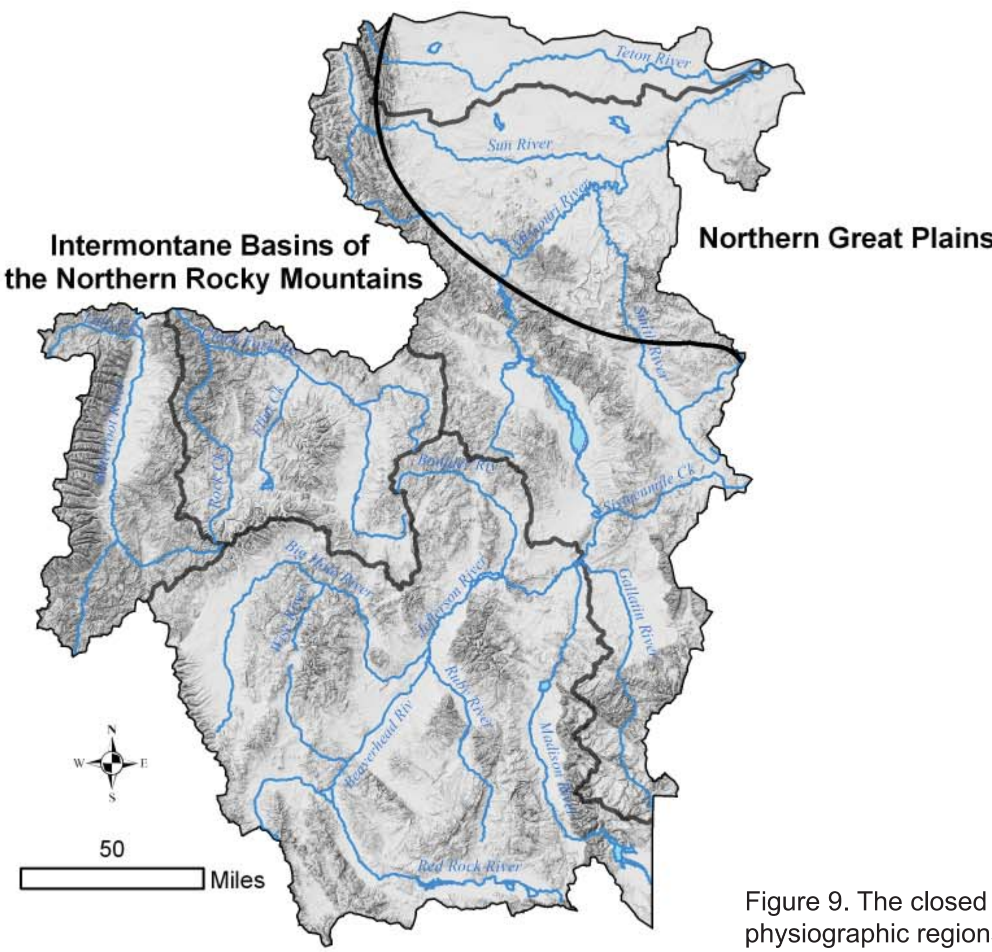


Figure 9. The closed basins occur in two different physiographic regions, the Northern Great Plains and the Northern Rocky Mountains.

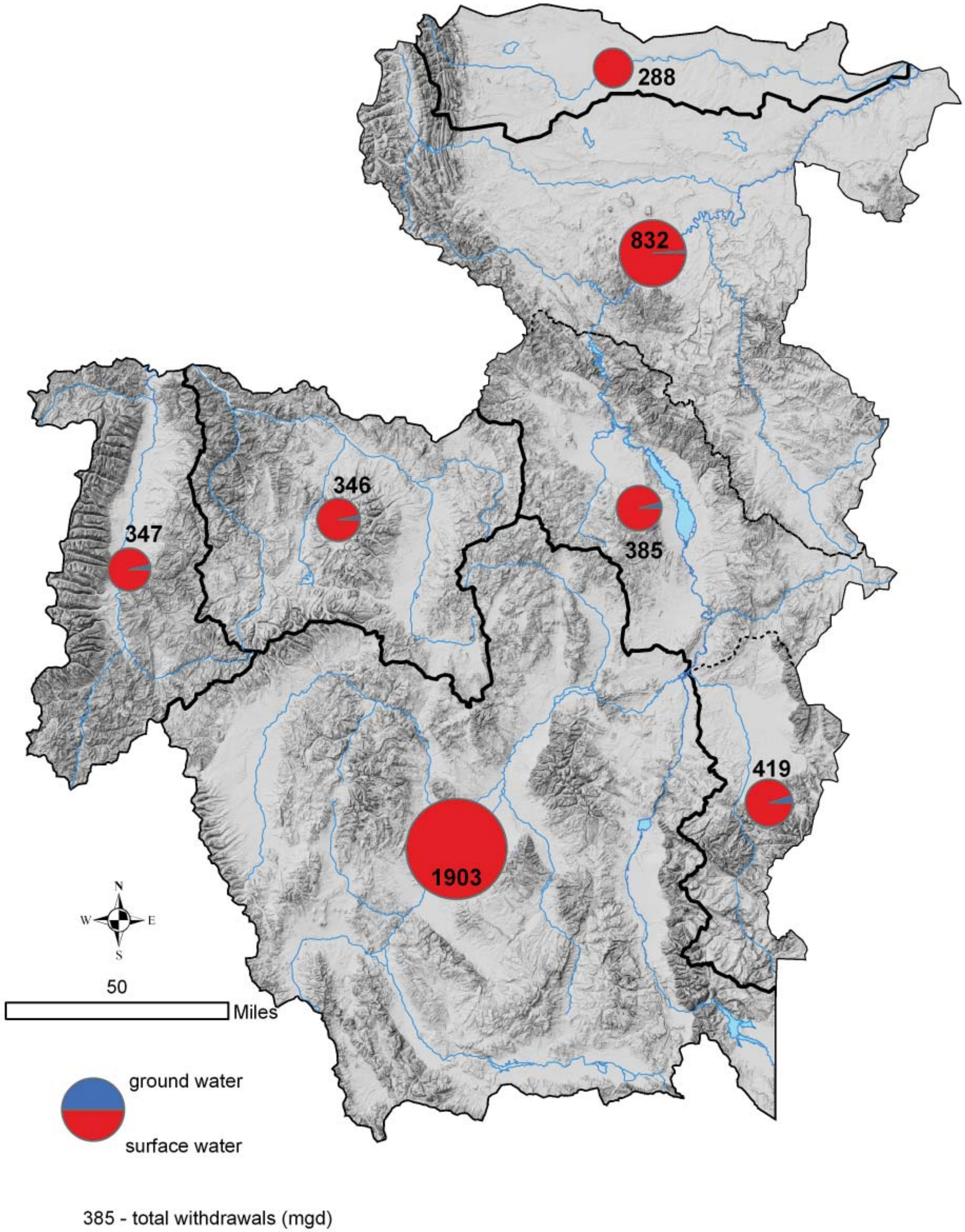


Figure 10. Water withdrawals in the closed basins are dominated by surface water. About 97 percent of water withdrawn for irrigation, industrial, municipal, domestic, and stock use is derived from surface-water resources (data from Cannon and Johnson, 2004).

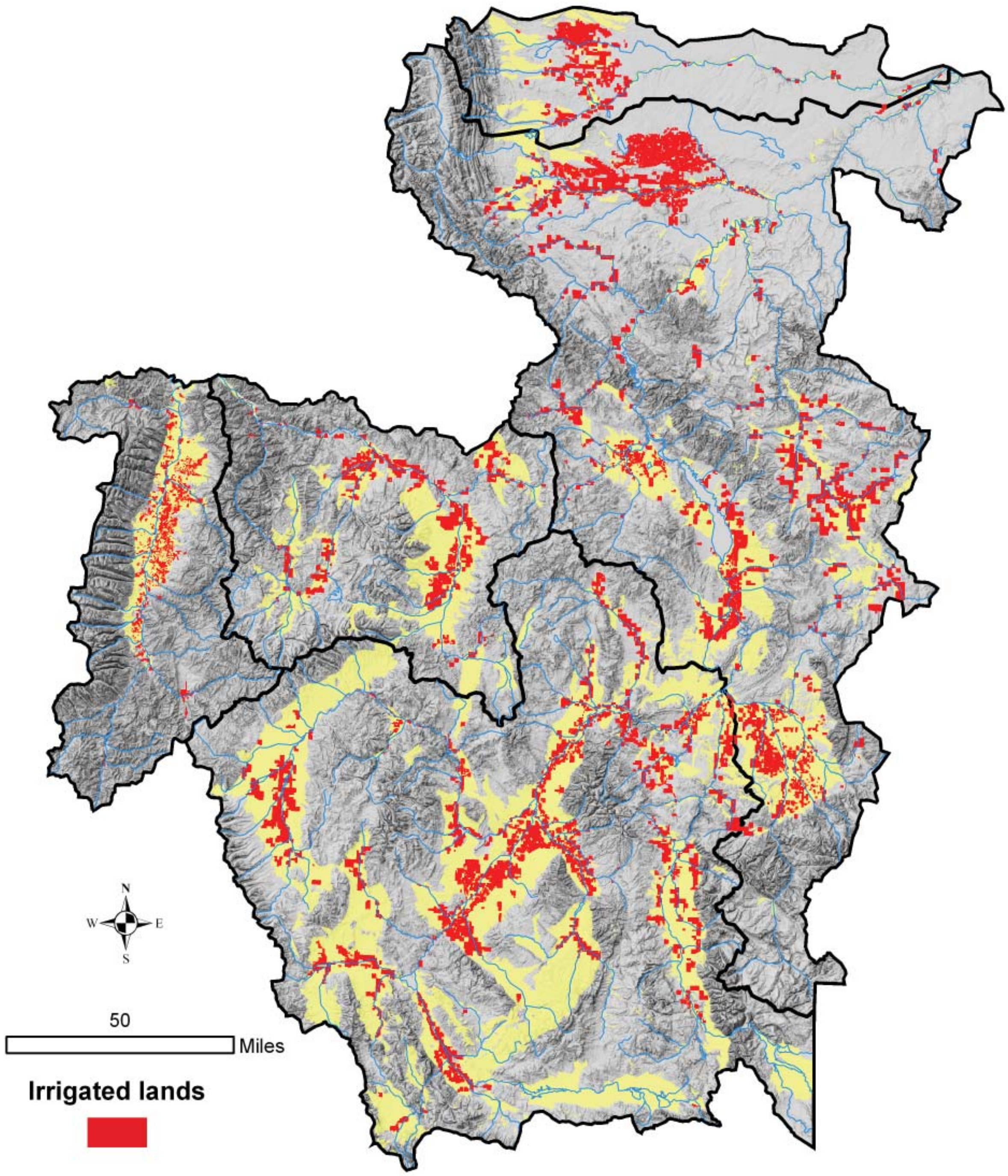


Figure 11. Irrigated lands generally occur in the valley bottoms near streams. Irrigated parcels obtained from the cadastral database; yellow represents areas of surficial sand and gravel.

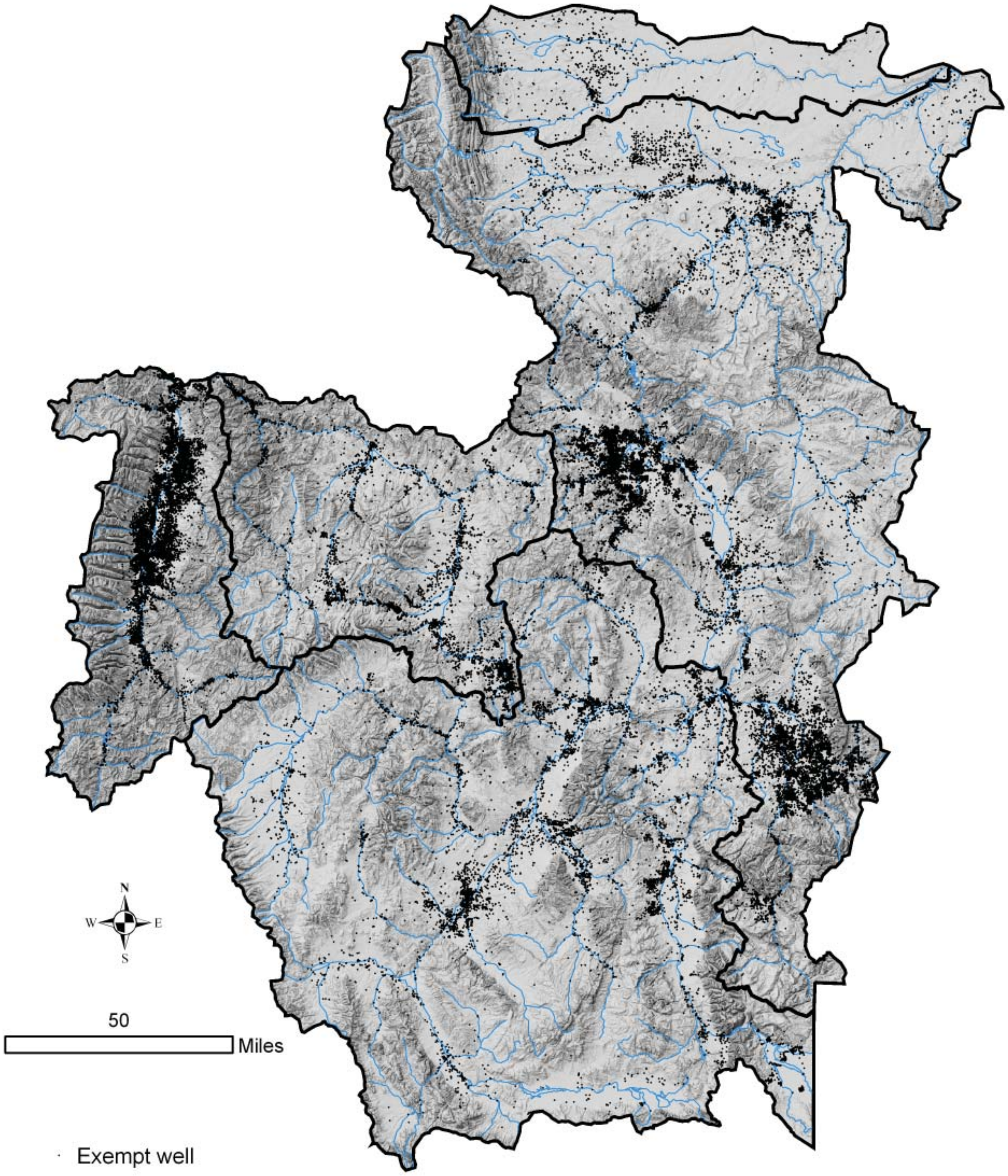


Figure 12. Locations of wells with a reported use of domestic or stock water. Data from the Montana Ground-Water Information Center.

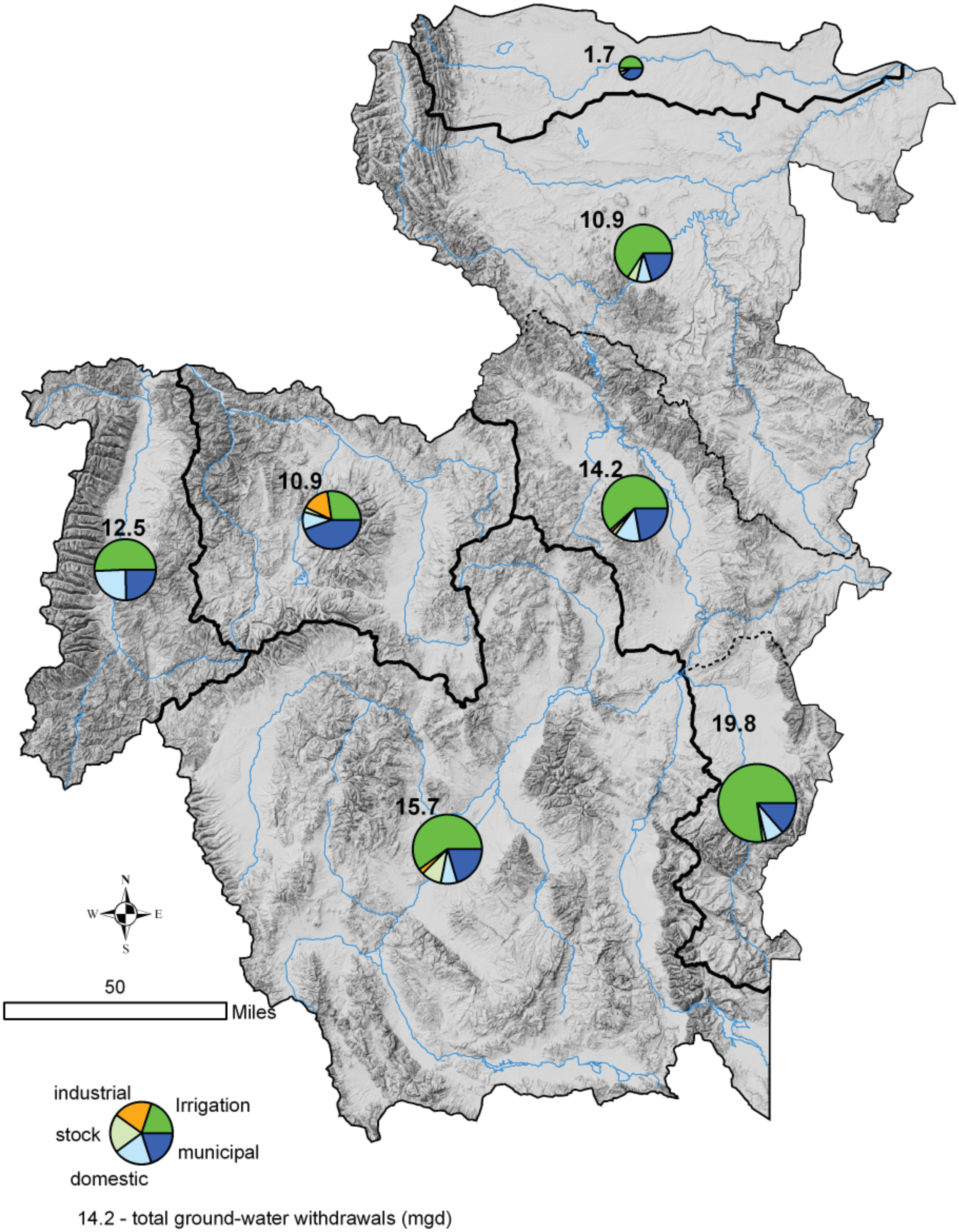


Figure 13. Most of the ground water withdrawn in the closed basins is used for irrigation and municipal water (data from Cannon and Johnson, 2004).

The climate of the study area is characterized by cold winters and mild summers, and most of the precipitation falls in winter and/or early spring. Across the closed basin area the average annual precipitation is 23 inches (University of Oregon PRISM data); however, the precipitation distribution (fig. 14) is closely related to altitude and topography. Valley bottom locations receive 10–14 inches annually, while adjacent mountain ranges receive in excess of 30 inches (fig. 14). Much of the mountain precipitation is in the form of snow and is stored into spring as snow pack. The spring snowmelt and subsequent runoff are important aspects of the water cycle; in many basins snowmelt may maintain stream flow well into the summer.

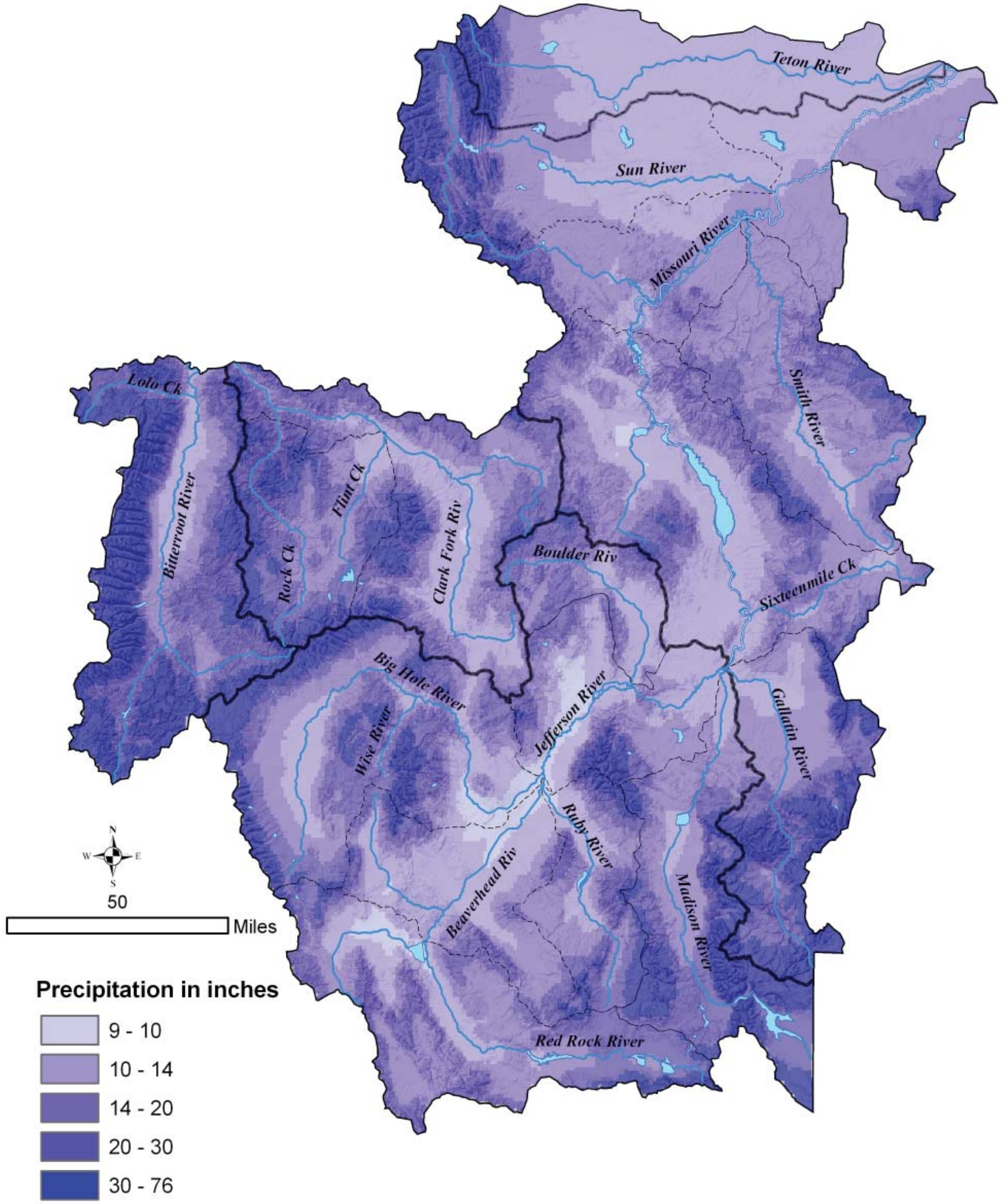
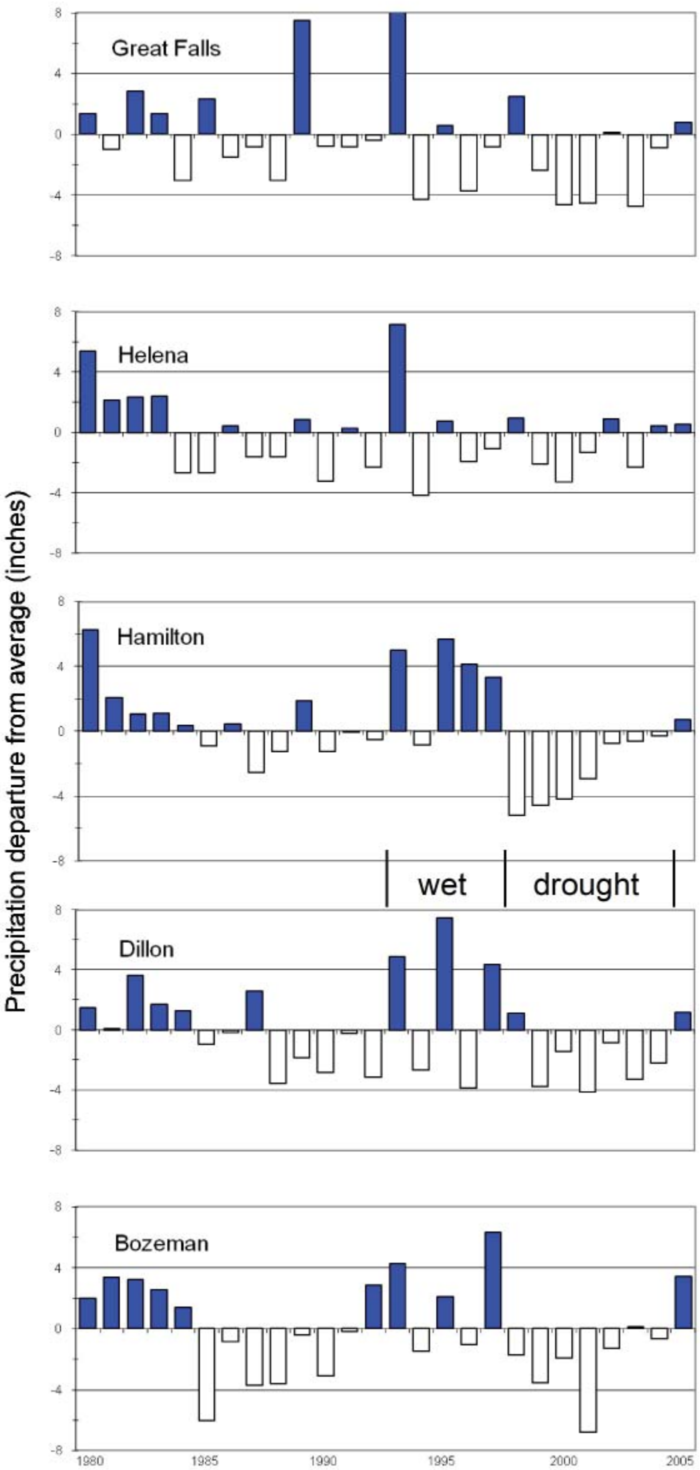


Figure 14. Average annual precipitation map of the closed basin area. Precipitation is strongly influenced by topography. Precipitation data from the Oregon Climate service.

The amount of precipitation that falls each year—and hence is available for stream flow and ground-water recharge—is variable. The departure from average annual precipitation for several stations across the area is shown in figure 15; departures of 15 to 20 percent of the long-term average are typical. Multiple years of above average or below average precipitation show wet periods and droughts. During multi-year wet or dry periods, adjustments in equilibrium conditions in ground-water systems will be reflected in changes in storage and discharge relative to increased or decreased recharge. Figure 15 shows a long-term aquifer response to such changes.

GEOLOGY

Because the rock type determines the water-bearing characteristics of aquifers, an understanding of the geology is essential to understanding the occurrence and distribution of ground water. The closed basins occur in two different physiographic provinces: (1) the intermontane basins of the northern Rocky Mountains and (2) the northern Great Plains (fig. 9). The topography in each physiographic province reflects a broad difference in geology and geologic history. This in turn results in very different hydrogeologic settings. Generally speaking, the geologic units vary in composition from unconsolidated



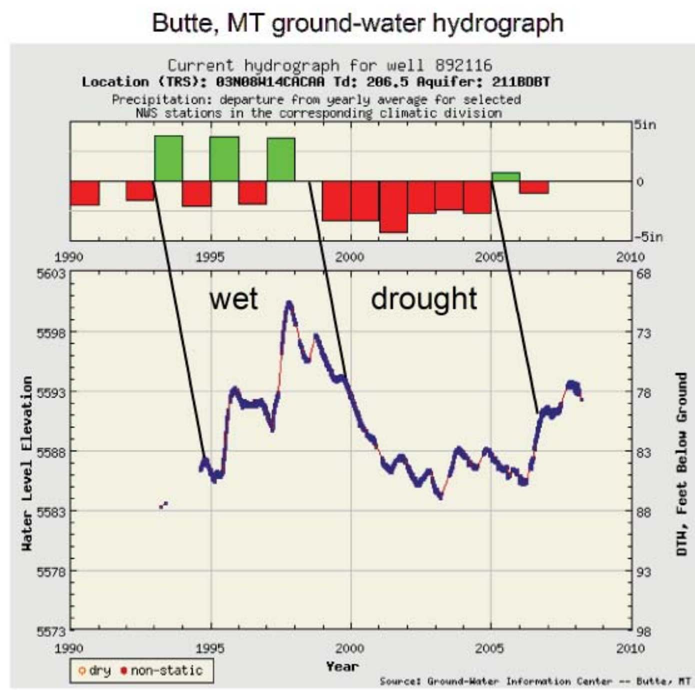


Figure 15. Average annual precipitation varies from year to year, with distinct wet and drought periods. The hydrographs from a well near Butte show the water table response to annual precipitation changes and longer-term wet and drought periods.

basin-fill deposits, to consolidated sedimentary bedrock, to metamorphic, igneous, and volcanic rocks (fig. 16).

The intermontane basins reflect a complex geologic history. The basins developed during an episode of extensional tectonics that started about 50 million years ago and continues today, as evidenced by current earthquake activity. The valleys opened up along faults, so at least one margin of each basin is fault-bounded. The geometry of a valley depends on the orientation of the fault zone along which it developed. The valleys are generally linear, and the dominant orientations are north–south, such as the upper Madison, Bitterroot, and Deer Lodge Valleys, and northeast–southwest, such as the Ruby, northern Beaverhead, and upper Jefferson Valleys. The lower Madison–Gallatin valley is anomalous in that it is bounded by faults with at least four orientations and is more equidimensional than the other valleys. Many smaller valleys follow the northwest–southeast structural grain inherited from structures in the oldest rocks in western Montana.

The basin-fill deposits in the different valleys consist of unconsolidated to semi-consolidated Tertiary-age sediments, overlain by younger, unconsolidated Quaternary-age sediments. The basin-fill composition may be highly variable within a valley and from one valley to another. For example, sediment may grade from clay-size on one side of a valley to bouldery gravel on the other side or may be coarse-grained at basin margins but fine-grained in the center. Permeable, coarse sediment often occurs in lenses enclosed by less permeable fine-grained sediment. During deposition of the basin fill, many plumes of volcanic ash spread over the area. Volcanic ash is common in the basin-fill deposits, both as beds and disseminated throughout the sediment. The ash has mostly altered to bentonite, an impermeable swelling clay, but elsewhere it consists of fine, glassy shards. Whether or not the ash has been altered to clay affects the permeability of the sediment, and where permeability is low helps form confining beds.

The depth of the valleys is also highly variable. A well in the northernmost part of the Gallatin Valley showed 245 ft of basin-fill deposits over bedrock; farther south near Churchill a well showed 826 ft of basin fill over bedrock, and 2 miles west of Four Corners a well showed about 580 ft of basin fill over bedrock (Hackett and others, 1960). In the upper Big Hole, a drillhole in the northern part of the basin penetrated basin-fill deposits to a depth of about 15,000 ft. A well drilled in the central part of the Deer Lodge Valley penetrated about 10,000 ft of Tertiary sedimentary beds before encountering volcanic rocks. In the Bitterroot Valley drillholes show that basin-fill deposits are, in places, 2,400 ft thick (Norbeck, 1980). The depth of the basin fill may vary significantly from one part of a valley to another; and in some basins bedrock protrudes through the valley fill.

Faults offset the deposits in most valleys and may juxtapose permeable and impermeable units. Many faults developed as the basins were forming and filling with sediment, an ongoing process. Some faults serve as ground-water conduits that produce springs at the surface. In addition to abundant faulting, the basins have complex histories that involve multiple episodes of filling with sediment, partial erosion, and subsequent refilling. The deposits below buried erosion surfaces may be significantly different than those above the surfaces, and the erosional surface that bounds them may be quite irregular. In some valleys, volcanic flows partially cover the erosional surfaces that developed between episodes of deposition.

The mountain ranges that separate the individual valleys are composed of many different rock types, including older Precambrian “basement” crystalline metamorphic rocks, younger Precambrian meta-sedimentary rocks (the Belt Supergroup), plutonic igneous rocks of Cretaceous age (Idaho and Boulder Batholiths), volcanic rocks of Cretaceous (Elkhorn volcanics) and Tertiary age (Lowland Creek volcanics), and Mesozoic and Paleozoic sedimentary rocks (fig. 16). Consolidated rocks that form the mountains generally have little primary porosity and low capacity for ground-water storage.

The geology of the closed basins in the northern Great Plains (Teton and the upper Missouri downstream from Holter dam), is dominated by older sedimentary rocks (figs. 8 and 16). The expansive plains that characterize most of this region are underlain by thick sequences of mostly Mesozoic shale with few sandstones. Paleozoic rocks are exposed along the Rocky Mountain Front and the Little Belt Mountains. The Madison Limestone is exposed along the northern flank of the Little Belt Mountains, and dips into the subsurface below the overlying Mesozoic units. Fracturing and dissolution have created significant porosity and permeability within the Madison. Relatively thin (less than 50 ft) Tertiary and Quaternary terrace gravels mantle the Mesozoic shale east of the Rocky Mountain front and form prominent “bench” features, for example the Fairfield and Burton benches. Quaternary alluvium occupies the Missouri, Sun, and Teton River valleys. Alluvium in the Missouri River valley and associated paleochannels near Great Falls may be as much as 200 ft thick. Much of the land surface in the Teton basin is mantled by till deposited by continental glaciation.

DISTRIBUTION OF AQUIFERS

In the intermontane basins, there are generally two types of aquifers: (1) shallow water table aquifers and (2) deeper confined to semi-confined aquifers. These aquifers contain large amounts of ground water and are highly productive and utilized. The extent of the basin-fill aquifers generally coincides with the extent of basin-fill deposits shown in figure 16.

Near-surface sand and gravel deposits (mostly

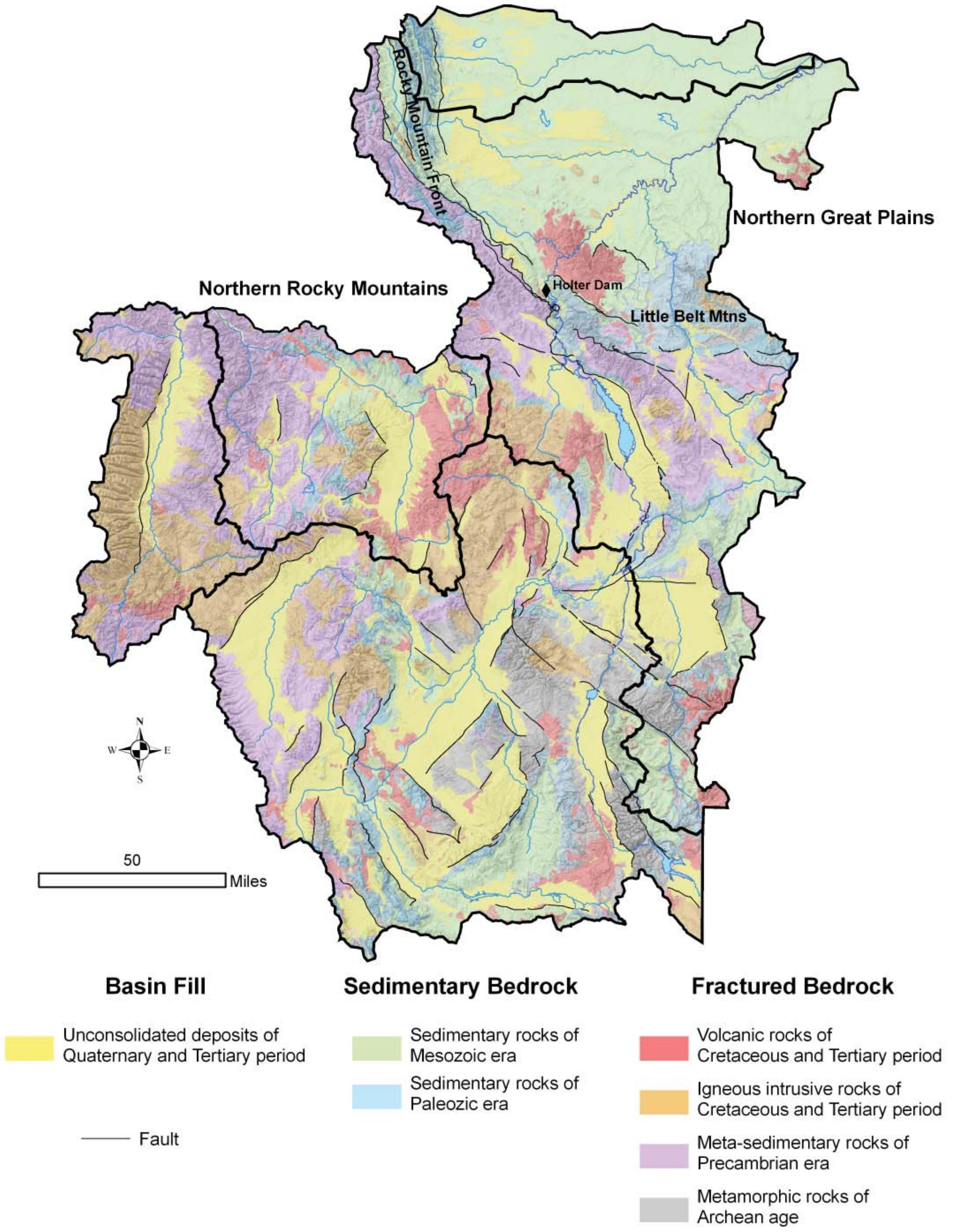


Figure 16. Generalized geology of the closed basin area.

Quaternary alluvium) coincident with the floodplains of main-stem streams contain very permeable aquifers that store and yield large volumes of water. These shallow alluvial aquifers are generally less than 50 ft thick (but may range up to 100+ ft) and are in hydraulic connection with the adjacent streams.

The deeper basin-fill aquifers occur in layers of sand and gravel separated by layers of silt and clay. Sand and gravel layers buried below silt and clay form confined to semi-confined aquifers in most, if not all, of the basins. Upper Tertiary sediments form most of the deep basin-fill aquifers. These aquifers in the upper Tertiary sediments can also be significant sources of water in the closed basins (Gallatin Valley, Bitterroot Valley). These aquifers consist mostly of unconsolidated to semiconsolidated deposits of sand and gravel, with interbeds of silt, clay, and volcanic ash. Some of these deposits are ancient alluvial fans that coalesced along the mountains bordering the basins. Some upper Tertiary aquifers were deposited by ancestral rivers that occupied the basins (Bitterroot). A generalized model of basin-fill aquifers is shown in figure 17.

Ground water is not as abundant in the northern Great Plains province. There are two major bedrock aquifers in the Paleozoic and Mesozoic rocks. The Madison Limestone is a high-yielding aquifer in upper Paleozoic rocks. The Madison aquifer is an important source of municipal, domestic, and agricultural water in the southern part of Cascade County. It is also the source of water that discharges from Giant Springs. The Mesozoic Kootenai Formation includes layers of medium- to coarse-grained sandstone that constitute an aquifer where it is exposed or where it is near the land surface—generally south of the Sun and Missouri Rivers. The terrace gravel benches are aquifers where there is sufficient recharge from irrigation water (Fairfield and Burton benches). The stream valley alluvium, especially along the Missouri River, also forms aquifers.

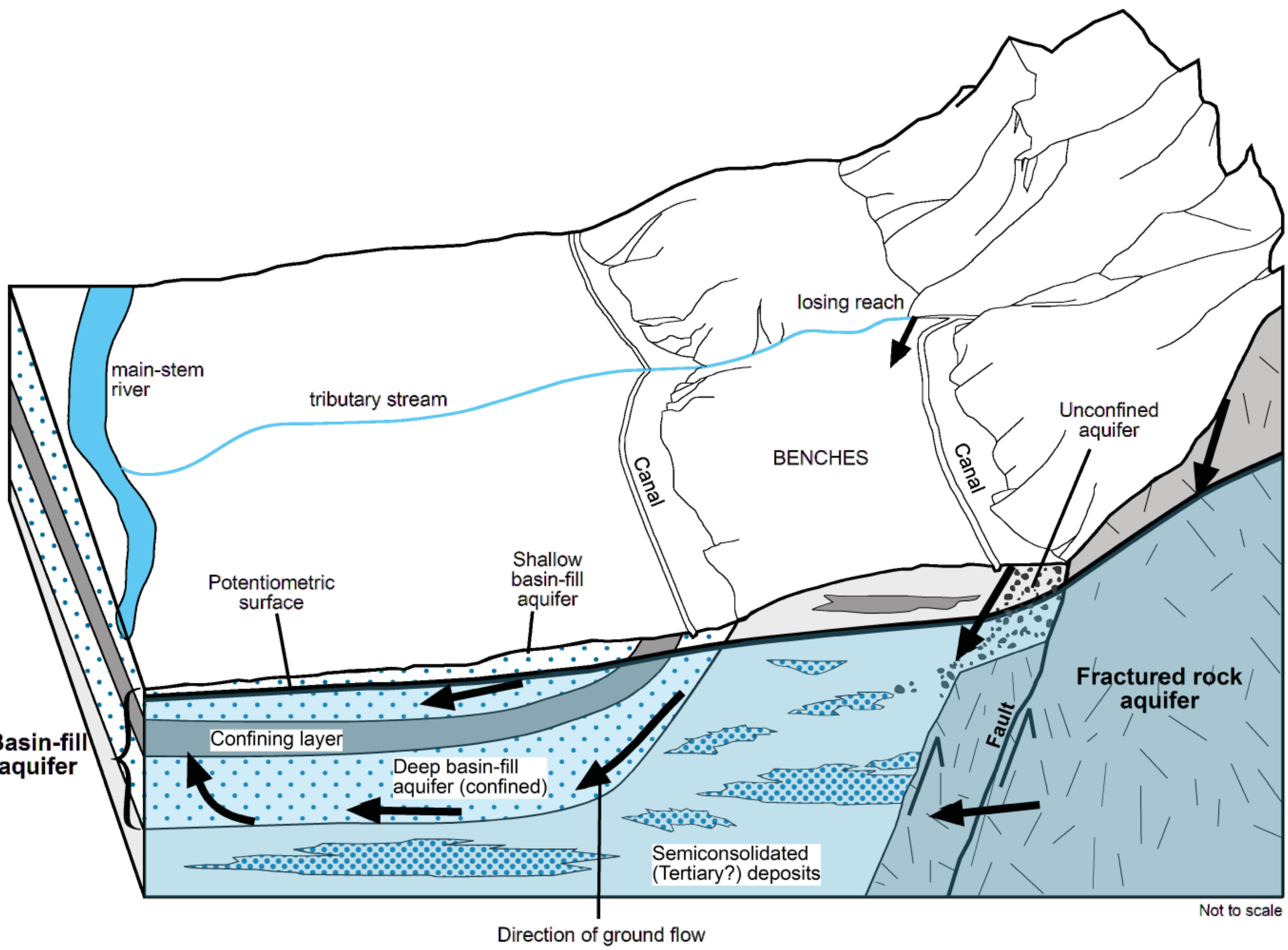


Figure 17. Schematic block diagram showing the distribution of basin-fill aquifers and ground-water flow in the intermontane basin part of the closed basin area.

GROUND-WATER OCCURRENCE AND MOVEMENT

Ground water in the closed basins is moving in response to gravity from recharge areas down the hydraulic gradient to discharge areas. The recharge is from precipitation that falls directly on the aquifers, from leakage through the beds of streams that cross the aquifers, and from irrigation water infiltration. Some buried aquifers receive recharge by downward leakage from overlying aquifers, or by lateral flow from adjacent aquifers. Lateral recharge from mountain ranges is particularly important for some of the basin-fill aquifers. The mountains receive recharge directly from precipitation, much of it in the form of snow; as water percolates into the fractured rock aquifers that compose the mountains, some moves laterally into the adjunct basin-fill aquifers. The water then moves through the basin-fill aquifers and discharges to surface-water bodies, such as streams, near the basin centers. Streams that drain mountain headwater areas may become losing near the base of the mountains where streambed material changes from bedrock to basin fill and where the stream gradient decreases (fig. 17).

Most aquifer discharge is to streams and springs, and to evapotranspiration where the water table is close to the surface, such as in riparian areas. A much smaller fraction of ground-water discharge is the result of well pumpage.

The aquifers in the closed basins yield variable amounts of water. Unconsolidated basin-fill and consolidated Paleozoic and Mesozoic sedimentary rocks are the most productive aquifers, whereas crystalline rocks generally are the least permeable. Reported yields in the basin-fill aquifers range from less than 10 to more than 1,000 gallons per minute (gpm). Well yields adequate to supply domestic and livestock watering needs (generally less than 20 gpm) can be obtained from most of the aquifers. Well yields from the Kootenai Formation are generally less than 50 gpm, with an average reported yield of about 20 gpm. Well yields from the Madison reportedly range up to 600 gpm with an average of about 50 gpm. The largest yields are from wells completed in shallow basin fill and upper Tertiary aquifers; some wells completed in these aquifers yield as much as 3,000 gpm.

SEASONAL WATER-LEVEL FLUCTUATIONS

Within the closed basins there are 306 wells that are part of the Statewide Monitoring Network (fig. 18); hydrographs from selected wells are presented in the appendix. Under natural conditions, ground-water levels generally are highest in the spring as a result of recharge from snowmelt and rainfall. Water levels decline rapidly during the summer when evapotranspiration rates are highest and discharge exceeds recharge. Water levels continue to decline slowly through the fall and winter months until recharge resumes in the spring to complete the annual cycle. The magnitude of the seasonal fluctuation usually is on the order of a few feet per year.

Many valley bottoms in the closed basins are laced with canals and irrigated with surface water (fig. 11). Losses from the canals and seepage from irrigated fields constitute a significant fraction of aquifer recharge. Ground-water levels in such areas typically begin to rise during April and May when surface water is first released to canals and fields, maintain a maximum from midsummer to the end of the irrigation season, and then decline steadily to an annual minimum just before the start of the next growing season.

Hydrographs from two wells completed in the Bitterroot Valley shallow basin-fill aquifer highlight the significance of irrigation recharge. Figure 19 shows that ground-water levels in the irrigated area, near Hamilton, rise quickly at the onset of irrigation. Continual application of water throughout the irrigation season supports the water table at a high level. After irrigation is shut off in the late summer or fall, water levels begin to drop. A well outside of the irrigated area, near Florence, shows a much different water-level response that is synchronized with Bitterroot River flow; water levels peak near the time that stream flow peaks and gradually fall back to a base level. On average the magnitude of water-level fluctuation in the Florence well is about 2 ft, whereas the average water-level fluctuation in the Hamilton well is on the order of 10 ft. Recharge from irrigation water accounts for the difference.

Changes in irrigation practices, such as the conversion from flood to sprinkler irrigation, can reduce the amount of recharge to aquifers, resulting in a decline of water levels. Lining canals can also reduce recharge to aquifers. Urban development and subdivisions that result in an increase in paved area can also reduce aquifer recharge.

GROUND-WATER QUALITY

The ground water in the closed basins is generally of very high quality (fig. 20). Most ground water in the closed basins has dissolved-solids concentrations less than the U.S. EPA secondary maximum contaminant level of 500 milligrams per liter (mg/L).

The concentration of dissolved solids in ground water provides a basis for categorizing the general chemical quality. Dissolved solids in ground water primarily result from chemical interaction between the water and the rocks or the unconsolidated deposits through which the water moves. Rocks or deposits that consist of readily dissolved minerals will usually contain water that has large dissolved-solids concentrations. The rate of movement of water through an aquifer also affects dissolved-solids concentrations; the longer the contact time, the more mineralized the water becomes. Thus,

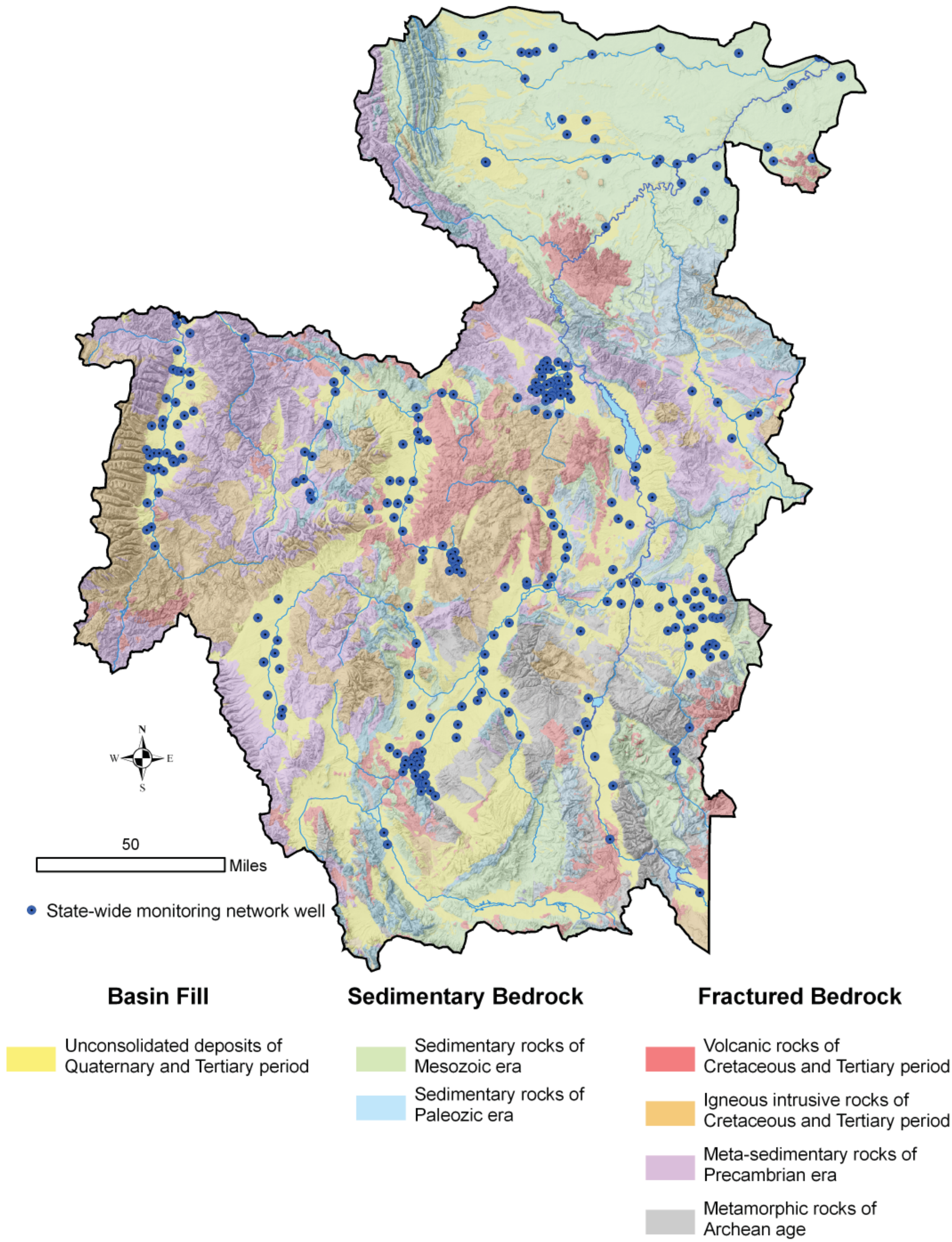


Figure 18. There are about 300 wells in different aquifers across the closed basin area that are monitored by the State-wide Monitoring Program at the Montana Bureau of Mines and Geology.

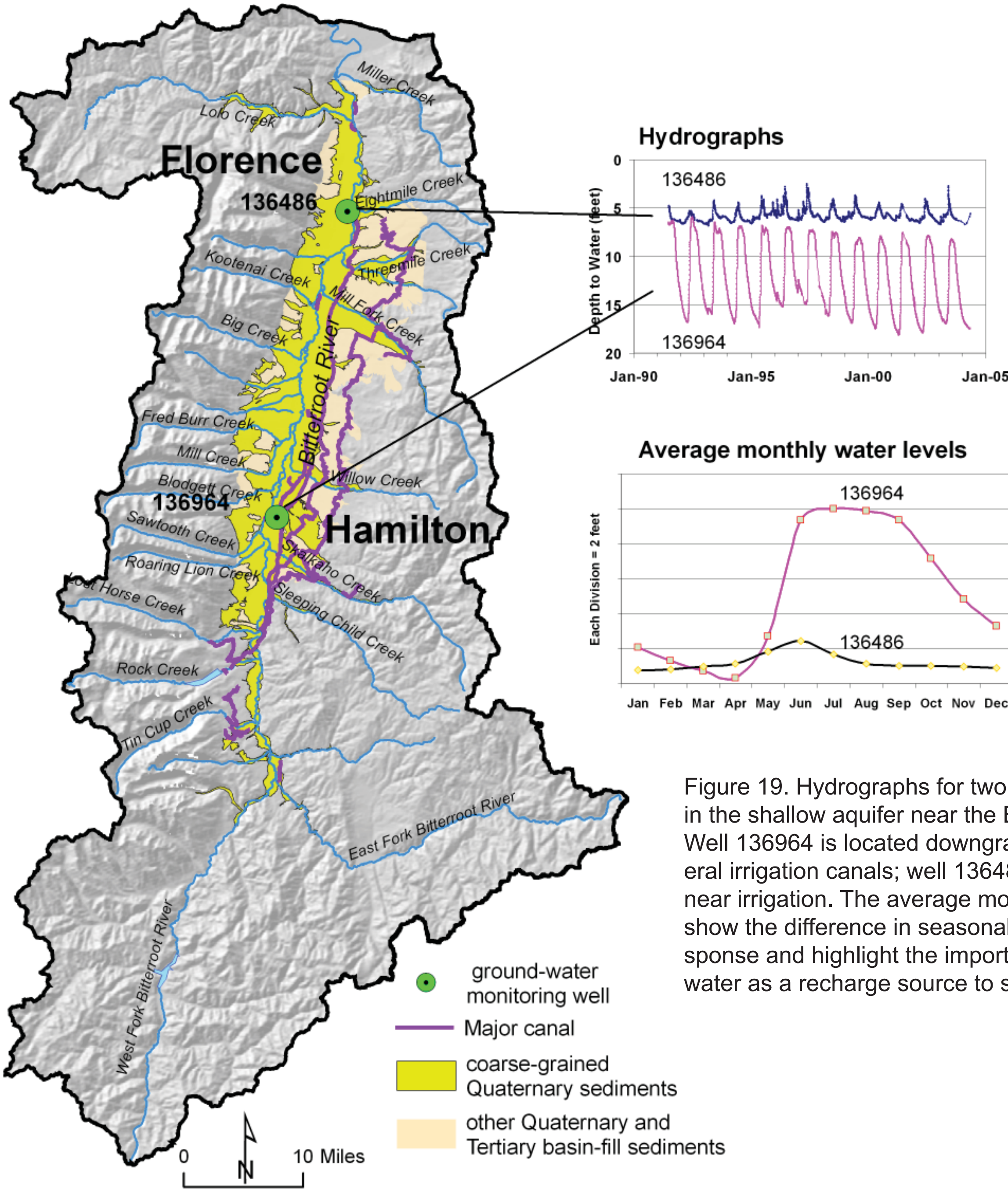


Figure 19. Hydrographs for two wells completed in the shallow aquifer near the Bitterroot River. Well 136964 is located downgradient from several irrigation canals; well 136486 is not located near irrigation. The average monthly water levels show the difference in seasonal water-level response and highlight the importance of irrigation water as a recharge source to shallow aquifers.

dissolved-solids concentrations in ground water generally are small in aquifer recharge areas and increase as the water moves downward into the deeper parts of the aquifers or near the ends of long ground-water flow paths.

Ground-water contamination that results from human activities can be categorized as being from either a point or a nonpoint source. A point source is a specific local site, such as a leaking underground storage tank that contains wastes or chemicals, a landfill, or a storage pond, pit, or lagoon. Nonpoint contamination sources are large scale and can extend over hundreds of acres, such as the application of fertilizer or pesticides to fields, urban areas with concentrations of septic tanks and cesspools or highly mineralized geothermal water, animal feedlots, mining operations, or salt from roadway deicing.

Shallow, unconfined aquifers are most susceptible to contamination from human activities because of the relative ease with which water can move from the land surface to the water table. Fractured rock and limestone aquifers are also particularly susceptible to contamination because they commonly contain large openings (solution cavities, joints, or fractures) that allow water to enter the aquifer almost instantaneously with little dilution (mixing) or dispersion. Confined aquifers are less susceptible to contamination than unconfined aquifers because they are buried and overlain by confining units that have minimal permeability and because the water within the aquifer is under pressure. Infiltration of contaminants into confined aquifers is slow and will only occur in recharge areas.

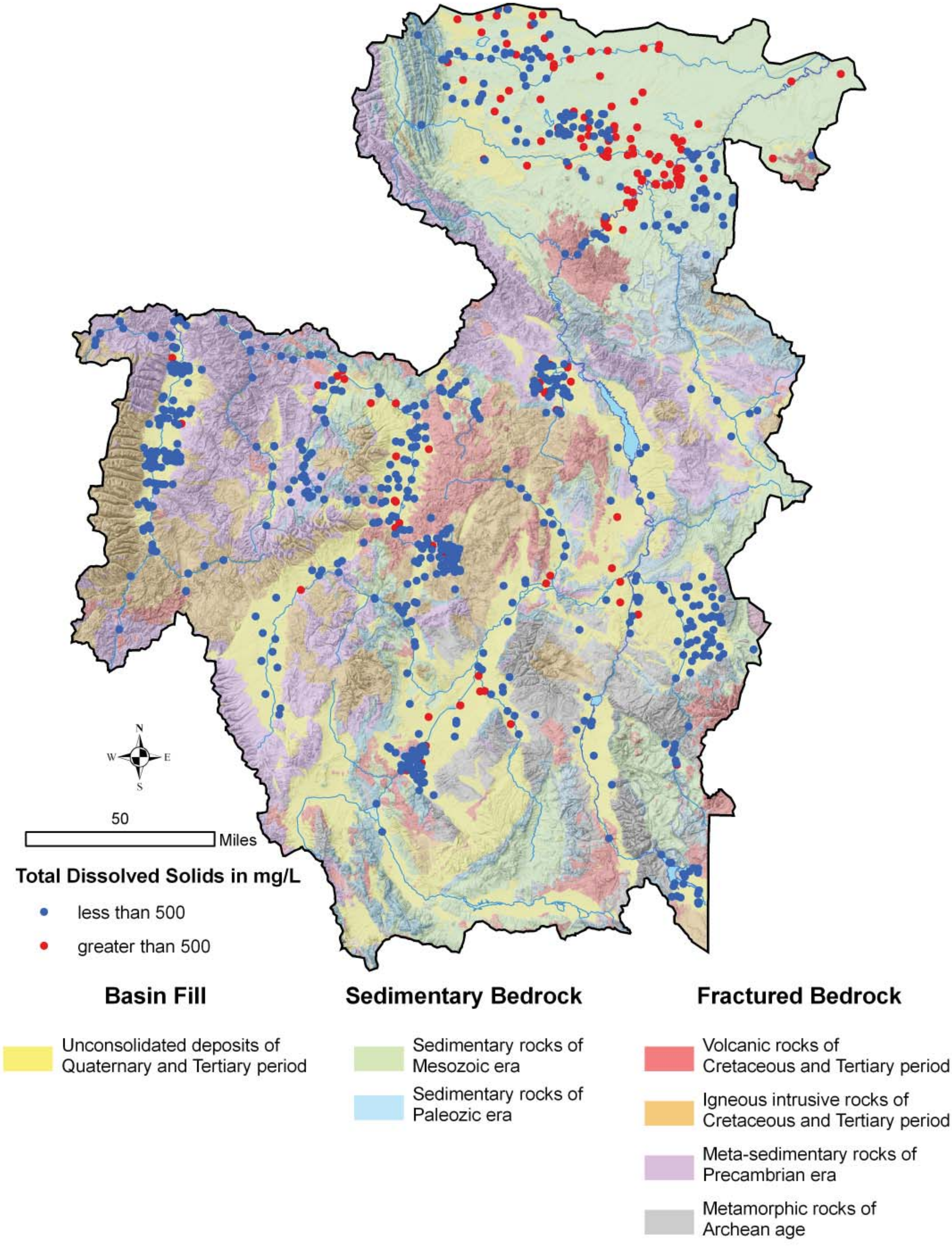


Figure 20. Ground water in the closed basin area is generally of high quality characterized by low total dissolved-solids concentrations.

SECTION 2: CASE STUDIES

Beaverhead:

**Ginette Abdo, Assistant Research Hydrogeologist,
Gary Icopini, Research Hydrogeologist, and
John Metesh, Senior Research Hydrogeologist**

Gallatin:

Kirk Waren, Hydrogeologist

Bitterroot:

John LaFave, Associate Research Hydrogeologist

LOWER BEAVERHEAD RIVER CASE STUDY

by

Ginette Abdo, Assistant Research Hydrogeologist,
Gary Icopini, Research Hydrogeologist, and
John Metesh, Senior Research Hydrogeologist

SUMMARY

The lower Beaverhead River sub-basin between Anderson Lane and Beaverhead Rock is one of three case studies conducted for HB 831. As many as eight ground-water permit applications for irrigation were in progress for this area. A review of the hydrogeologic assessments completed for those applications and others indicated that this sub-basin would provide a good setting to examine the range of hydrogeologic conditions common to the closed basins in Montana.

The objectives of this case study are many-fold:

1. Provide a detailed hydrogeologic evaluation and description of the lower Beaverhead River with an emphasis on ground-water and surface-water interaction;

2. Conduct an assessment of stream depletion by ground-water pumping under a range of hydrogeologic conditions and a range of pumping conditions; and

3. Evaluate the effectiveness of alternative strategies to offset stream depletion based on the stream depletion analysis.

As with all the watersheds in western Montana, the geology of the Beaverhead River basin defines the interaction between surface water and ground water. The floodplain is composed of Quaternary gravels, sands, silts, and clays washed into the valley from tributary streams and re-deposited by the river; these sediments are most often referred to as alluvial deposits. The floodplain deposits are generally 60 to 70 ft thick. Underlying the alluvial deposits of the floodplain are Tertiary layers of semi-consolidated conglomerates, sandstones, siltstones, and claystones. These rocks also make up the hills and benches south, north, and southeast of the floodplain. The hills north and northeast of the floodplain are also composed of Paleozoic limestones and sandstones as well as some Tertiary volcanic rocks. Faulting has caused a valley constriction at Beaverhead Rock where the Madison Limestone has been uplifted to form a partial ground-water dam at Beaverhead Rock. As a result, nearly all of the water exiting the basin at Beaverhead Rock is surface water, because the bedrock is so close to surface in that area.

A remarkable feature of the lower Beaverhead River geology is not evident at the surface. Beneath the sands and gravels of the uppermost alluvial deposit and on top of the bedrock is a layer of clay that is approximately 30 ft thick. A review of the wells in the floodplain strongly suggest that the clay is extensive and perhaps continuous from Dillon to upstream of Beaverhead Rock and potentially at least as wide as the floodplain. Locally, if not throughout the floodplain area, the clay layer can have significant effect on ground-water flow. Many wells completed in the deeper aquifer indicated a water level above the clay layer; in other words, the lower aquifer is confined and is separated from the Beaverhead River. At some point upstream of Beaverhead Rock and at the floodplain margins, well logs indicate that the clay layer either ends or merges with bedrock clay units. Aquifer tests conducted as part of this investigation confirm the separation of the lower aquifer from the alluvial aquifer and the river.

The source for all water in the valley is precipitation, primarily in the high mountains. The Clark Canyon Reservoir stores much of this water, which is then released to maintain stream flow and supply various ditches. For example, the East Bench Irrigation Canal that originates at Barrett's Diversion Dam provides irrigation to about 17,200 acres and is a significant source of ground-water recharge to the study area.

Ground water has been developed to some extent in all of the geologic units except the clay layer: the shallow, near-stream alluvium has been developed primarily for domestic and stock water use and the layered aquifer beneath the valley and on the south, north, and southwest flanks of the valley has been developed for irrigation. The bedrock on the north and northwest flanks of the valley has seen only moderate development for domestic use, but the volcanic rocks have demonstrated a good potential for high-capacity wells suitable for irrigation.

HB 831 addresses the concept of using mitigation or aquifer recharge as a way to manage stream depletion caused by pumping ground water. There are many methods available to calculate stream depletion, generally falling into two categories: analytical methods that provide a stream depletion value based on simple conditions and numerical methods, or modeling, that allow an examination of more complex situations. This report uses both to demonstrate their value in estimating stream depletion under several conditions.

The lower Beaverhead River provided a good opportunity to look at ground-water/surface-water interaction with an emphasis on evaluating depletion and methods to offset depletion. The objective was to evaluate the effects of ground-water withdrawal on stream flow under the range of hydrologeologic conditions found in the Lower Beaverhead River sub-basin as well as other sub-basins in western Montana. The analyses considered a test well and four irrigation wells (fig. 19 of the report) as representing the four hydrologic conditions under consideration: near-stream shallow (test well), near-stream deep (IR3), distal deep (IR1 and IR2), and basin margin deep (IR4).

Simulation of the near-stream, shallow conditions demonstrated the "immediate and direct" nature of pumping ground water next to a stream. The stream depletion rate reached the pump discharge rate quickly and continued to expand its influence on stream discharge after pumping stopped. In this particular case, the clay layer beneath the aquifer probably increased the rate at which the depletion occurred by limiting the volume of aquifer available to provide water to the well. The result is a greater need for water from the stream to supply the well.

Pumping ground water from the deep aquifer near the stream was evaluated with both the analytical and numerical methods. This gave an opportunity to compare the two methods as well as "calibrate" the numerical model. When both methods are used to calculate a stream depletion rate for a single well under simple conditions, they agree very well. For a single well pumping 850 gallons per minute (gpm) for 30 days, both methods calculate a stream depletion rate of about 0.32 cubic feet per second (cfs). A good comparison between the two methods lends confidence to using the model for more complex simulations. Two wells at roughly the same distance, about 1800 ft, from the stream were simulated with the model; pumping for 30 days at 850 gpm produced a stream depletion of 0.15 cfs for each well. A fourth well at a distance of about 20,000 ft from the river produced a stream depletion rate of about 0.13 cfs for the same pumping rate and pumping period. The results of each of these simulations of individual wells would compare favorably with the analytical method, but the analytical method cannot handle multiple wells. The model was set up to run all four wells at the same time; the result was a total stream depletion rate of 0.45 cfs. The results from the cumulative pumping simulation appears to underestimate the additive rate of each well, which would be 0.76 cfs. The reason lies in the relationship of time vs. distance for each well. The rate of depletion of the stream caused by each well hits its maximum rate at different times depending on its distance from the stream. This demonstrates that any simulation must be of sufficient length in time to evaluate cumulative effects of all wells, particularly those at greater distances from the stream.

The next series of simulations with the model evaluated stream depletion caused by more realistic conditions: pumping 850 gpm from each well for 90 days per year for several years.

As with the individual wells and the single pumping cycle, stream discharge decreased as pumping continued and recovered after pumping ended. The recovery, however, was not complete. A plot of stream discharge vs. time (fig. 29 of the report) reveals the trend of decreasing stream discharge resulting from a repeating cycle of pumping. It should be noted, however, that the trend is not linear. In other words, the cumulative effect of ground-water withdrawal is not simply a matter of adding up depletions caused by each well for each year. Stream depletion will reach a maximum value at some point in time and will be some fraction of the rate pumped from the well. The theoretical maximum fraction of depletion by a well is, of course, equal to 1.0 or 100 percent; in other words, the stream depletion is equal to the well discharge. However, the time it takes to reach the theoretical maximum must be considered.

The cumulative effects of four wells pumping 850 gpm for 90 days over a period of about 4 years is presented in figure 31 of the report. At the end of four pumping cycles, the stream depletion rate is about 1.5 cfs and the maximum fraction of the well discharge is about 70 percent; that is, the stream depletion rate at the end of the 4 years is 70 percent of the expected maximum rate. The stream depletion rate will reach a maximum value at some point in time and will be some fraction of the rate pumped from the well.

Stream depletion, expressed as a volume over time or rate, takes time to develop and reach a maximum. When and where along the stream the depletion occurs depends on the pumping rate and location of the well; multiple wells pumping in cycles will each affect the stream at different places and at different times. The preceding analyses demonstrated various means to estimate the timing, location, and volume/rate of depletion, but also demonstrated the needed to determine the rate of depletion with respect to time. Likewise, effective offset of stream depletion requires knowledge of when, where, how much, and how fast depletion will occur and knowledge of how a particular strategy that will offset stream depletion.

Methods to offset stream depletion from pumping near the Beaverhead River were simulated under two general conditions: replenishment of stream flow and replenishment of ground water. Figure 33 of the report presents the results of adding 2 cfs to the river well upstream of the depletion area during the same 90-day period as the pumping. This would be the effect of diverting water from some other source to the river or releasing stored water from a reservoir. The addition of the 2 cfs is easily recognizable by the spikes in stream discharge, but the longer term effect on stream discharge is inconsequential. The stream discharge never reaches its original value unless water is added. Immediately after the addition of the 2 cfs is stopped, stream discharge returns to near baseline discharge. Overall, stream discharge continues its trend downward and stream depletion is not mitigated. As a comparison, the same simulation was run, but instead of applying the 2 cfs to the stream, it was applied to ground water by using an irrigation ditch as a means of infiltration. Figure 34 of the report presents the results of replenishing ground water. Using infiltration, there is much more improvement in stream discharge compared to the surface-water approach, but it is nowhere near the original and the trend is still downward. For the short term, stream depletion has been offset but not eliminated.

The modeling shows that the goal of offsetting stream depletion has not been completely achieved by the ground-water approach, but these simulations serve to demonstrate the difference between replenishing surface water and replenishing ground water. The difference in response reflects the difference in residence time or storage time. In the model and in field conditions, the 2 cfs added to the stream will run off within days; the residence time or storage time is relatively short. Conversely, ground water has a residence time of weeks to months in shallow near-stream aquifers. Thus, a robust evaluation of stream depletion and offset will take these differences into account.

These analyses provide the background for meeting the objective of finding solutions specific to the lower Beaverhead River. Certainly, more simulations could be run for other conditions, including some that are in practice in other areas of the basin. For example, mitigation with the irrigation canal (but no flooding) over a longer period of time each year or at variable rates that take advantage of high stream discharge during spring runoff. Conveyance loss from irrigation canals can be considerable and may be a good component of offsetting depletion if the timing and location are favorable. If spring runoff waters could be diverted, the strategy becomes a form of aquifer storage recovery for an unconfined system; ice in the rivers and irrigation canals and frost may limit this approach in the Beaverhead River basin.

A major water consumer that is currently not being considered in the water management of the closed basins is the management of woody phreatophyte propagation (salt cedar, cottonwood, willow, etc.). Phreatophytes are defined as a type of plant that has a high rate of transpiration by virtue of a taproot extending to the water table. Native woody phreatophytes are generally promoted by State and Federal agencies, because these plants provide important habitat for a number of animal species. Phreatophyte water consumption rates are comparable to the annual application rate for sprinkler-irrigated alfalfa in southwestern Montana. However, due to their proximity to streams, these plants will have a direct impact on in-stream flows during the summer months when flows are lowest, and unlike irrigation this consumption cannot be limited or controlled once in-stream flows reach a critical point.

The effects of evapotranspiration from crops and phreatophytes were not considered in these evaluations, mainly to keep the differences between the various methods simple and clear. As noted, some have suggested substituting evapotranspiration or even lumping all consumptive use for the well discharge value. In the analytical methods, such as the Schroeder methods used here, the result is a much lower depletion rate. The same is true for the numeric modeling, but the limited value of the application becomes clearer. When and where the unconsumed water is returned to the system is every bit as important as the method(s) of offsetting stream depletion; it should not be treated as an artifact of a budget calculation. The simple example is perhaps the most common condition: pumping from a deep aquifer and recharging a shallow aquifer with the excess water. The deeper aquifer is not likely to be recharged by the shallow aquifer before the excess water is discharged to surface water. These analyses, along with exhaustive research presented in the literature, demonstrate that the timing, the rate, and the location of depletion are affected directly by the timing, location, and rate of pumping. The same is true for offsetting that depletion.

In the summer of 2008, additional surface-water and ground-water data will be collected during the irrigation season that will be used to "nail down" the relationships between pumping ground water, irrigation canal leakage (recharge), and stream discharge. The new data will also allow the model area to be expanded to include the area west and southwest.

The methods of analyzing stream depletion and offset of stream depletion for the lower Beaverhead River, while pertinent for local conditions, are easily applied in other parts of this and other watersheds. The area of investigation is of sufficient extent to develop useful models, but not so large as to take many years of study.

INTRODUCTION

The lower Beaverhead River sub-basin is one of three case studies conducted for HB 831. As many as eight permit applications for ground-water withdrawal for irrigation were in progress for this area. A review of the hydrogeologic assessments completed for those applications and others indicated that this sub-basin would provide a good setting to examine the range of hydrogeologic conditions common to the closed basins in Montana. In general, the Beaverhead River below Dillon, Montana is bounded by several types of bedrock on the valley margins, with alluvial deposits in the valley bottom.

The objectives of this case study are many-fold:

1. Provide a detailed hydrogeologic evaluation and description of the lower Beaverhead River with an emphasis on ground-water and surface-water interaction;

2. Conduct an assessment of stream depletion by ground-water pumping under a range of hydrogeologic conditions and pumping conditions; and

3. Evaluate the effectiveness of alternative ways to offset stream depletion based on the stream depletion analysis.

The MBMG gratefully acknowledges the cooperation of the landowners and agents of the Geoduck, Sitz, Open A, and Dallaserra properties.

PHYSIOGRAPHY

The Beaverhead River drainage drains an area of about 2,895 miles (fig. 1). The river originates about 23 miles southwest of Dillon, Montana at the Clark Canyon Reservoir and drains to the northeast. The Clark Canyon Reservoir receives water from Red Rock River and Horse Prairie Creek. The river flows northeast through the Beaverhead Canyon and into the Beaverhead River Valley for about 45 miles until its confluence with the Big Hole and Ruby Rivers near Twin Bridges to form the headwaters to the Jefferson River.

Water from the river is diverted for irrigation at Barrett's Diversion Dam, 11 miles below Clark Canyon Dam. The water is diverted to the East Bench Canal, which flows in the upland area on the east side of the Beaverhead River valley, and the Clark Canyon Canal, which provides irrigation to the valley bottom. The East Bench Canal and its lateral diversion serve about 17,200 acres (Bureau of Reclamation, 2008).

The basin is bounded by the Pioneer Mountains to the west, the Ruby Mountains to the east, and the Tendoy, Snowcrest, and Blacktail Ranges to the south (fig. 1). Major tributaries to the Beaverhead River are Grasshopper Creek, which flows towards the northeast, joining the Beaverhead River above Barrett's Diversion. Blacktail Creek flows to the northwest in a northwest–southeast-trending valley that is nearly at right angles to the Beaverhead River valley, joining the Beaverhead River near Dillon. Rattlesnake Creek flows towards the southwest and also joins the Beaverhead River near Dillon.

North of Dillon to Beaverhead Rock, a distance of about 16 miles, Stone Creek and Spring Creek flow into the Beaverhead River from the Ruby Mountains to the southwest. From Beaverhead Rock to Twin Bridges, the Ruby River flows into the Beaverhead River from the Ruby Mountains to the south.

The focus area of this study was from Anderson Lane, located about 8 miles northeast of Dillon, to Beaverhead Rock (fig. 2). In this area the floodplain portion of the valley varies in width, from about 3 miles near Anderson Lane to about 2 miles a few miles north of Anderson Lane. At Beaverhead Rock, the river is constricted by bedrock and the floodplain is less then a quarter-mile wide. The river bottom ranges in elevation from 5,100 ft in Dillon to about 4,800 ft near Beaverhead Rock.

Climate

Average annual precipitation in Dillon, Montana is 13.14 inches based on a 117-year period of record

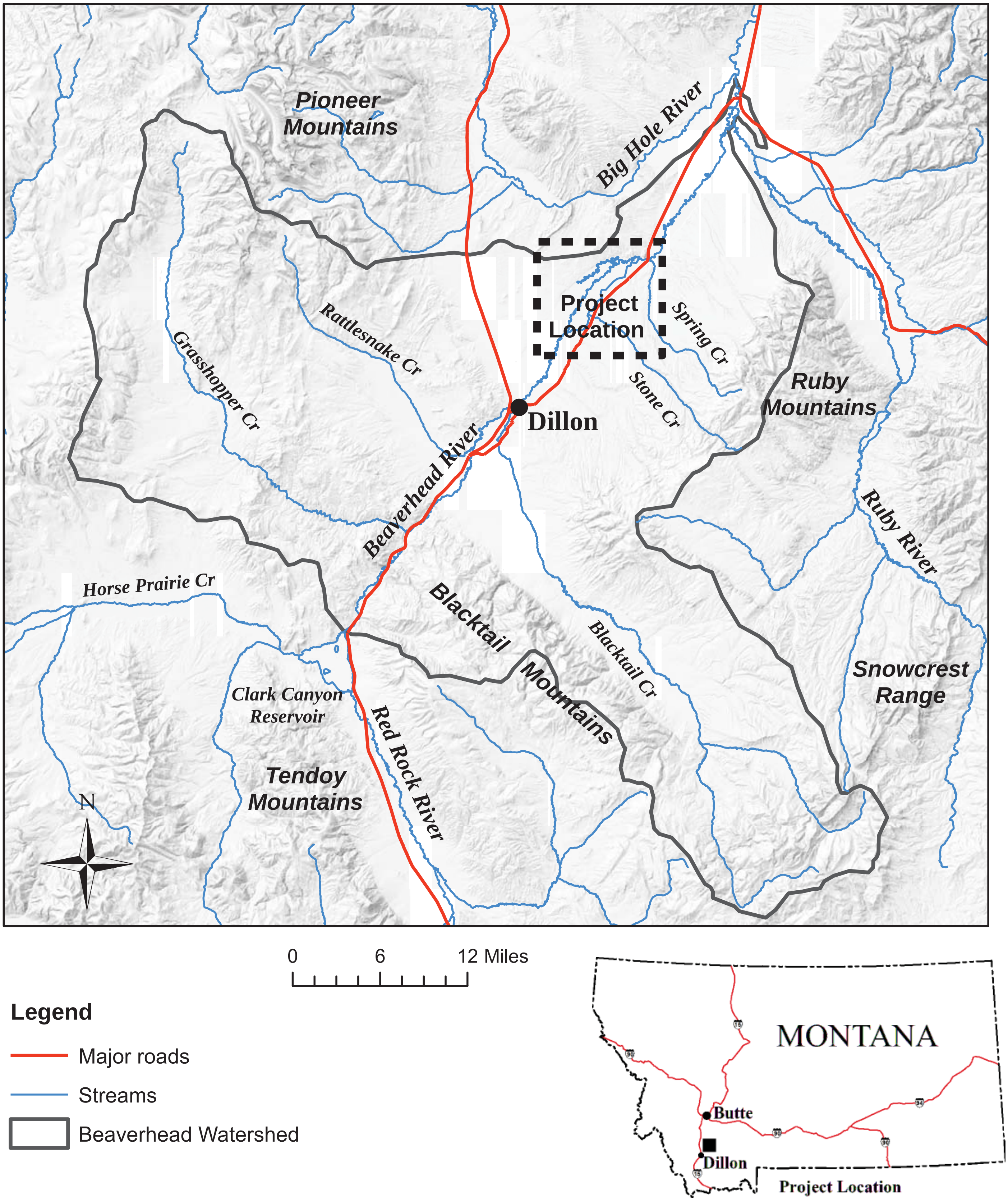


Figure 1. Map showing location of the Beaverhead River drainage and project area.

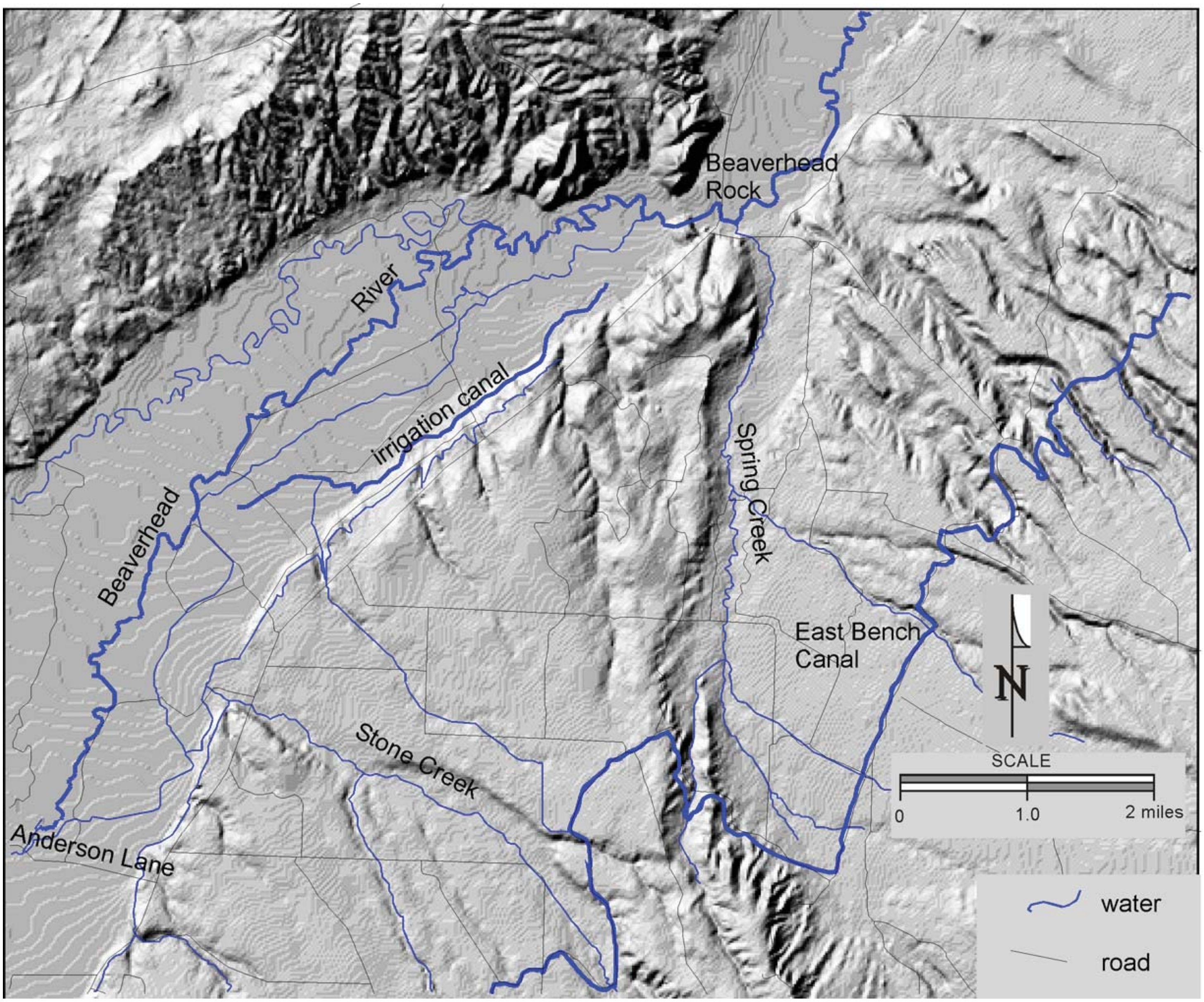


Figure 2. The focus area for this study was between Anderson Lane and Beaverhead Rock.

(Western Regional Climate Center, 2008). Figure 3 shows the departure from average annual precipitation from 1900 to 2007. In general, precipitation was above average from 1900 to 1930. Starting in the 1930s, during the Dust Bowl, there appears to be a general trend of below average precipitation years through 2007. There have been no years in the past decade with above average precipitation. The average maximum temperature over the period of record occurred in July (83.4°F), and the average minimum temperature in January (12.9°F). The annual average over the past 60 years of data is 11.91 inches; this would change the look of figure 3, but the general trend of below average precipitation would hold.

Geology

The Beaverhead valley geology is controlled by the northeast-trending Ruby Fault Zone along its southeast side (Ruppel, 1993), and in part by northeast-trending faults in the river valley (Ruppel and others, 1993). The Blacktail Deer Creek Valley is controlled by the northwest-trending Blacktail Fault Zone (Ruppel, 1993). The middle part of the Beaverhead River floodplain is constricted by the northwest-trending McCartney Fault Zone that bisects the basin and has brought bedrock to the surface at Beaverhead Rock northeast of Dillon. The July 25, 2005 Dillon earthquake and other recent seismic activity in the area are indications that some of the faults in the basin are active.

Gravity data (Hanna and others, 1993) suggest that the deepest part of the basin is in a northwest-trending fault zone that crosses and underlies the central part of the basin and is as much as 3,300 to 4,900 ft deep. South of the fault zone, the basin floor rises over a short distance to 1,690 ft, intercepted in a drillhole west of the Ruby Range (Ruppel, 1993). R. Thomas (oral commun., 2008, Professor of Geology, Western Montana College, Dillon, Montana) estimates that the valley-fill between Dillon and Beaverhead Rock may be about 1,000 ft thick.

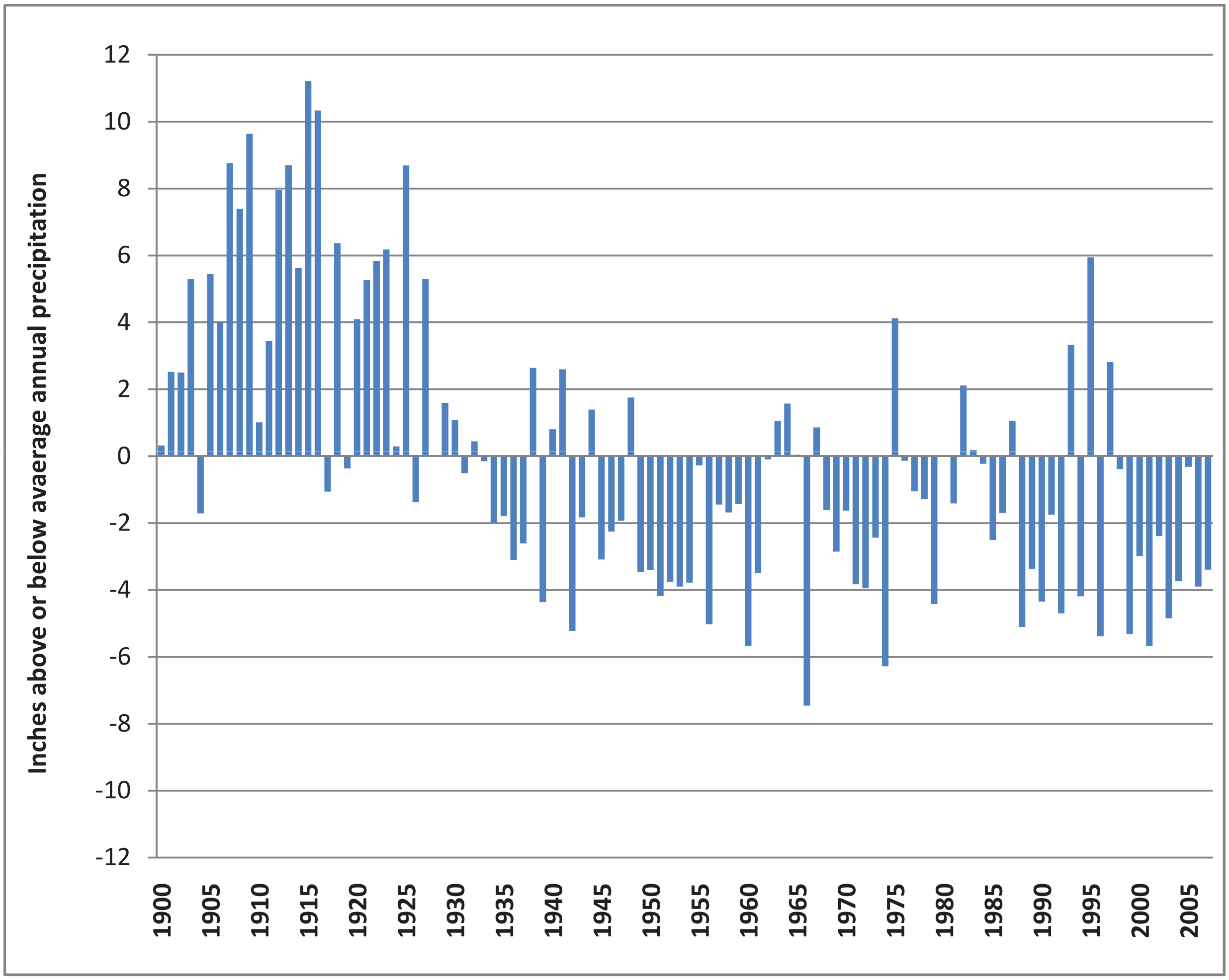


Figure 3. The departure from average annual precipitation from 1900 through 2007 at Western Montana College in Dillon, Montana. Note the below-average annual precipitation during the last two-thirds of the 1900s.

Figure 4 is a generalized geologic map between Dillon and Beaverhead Rock. Most of the bedrock associated with the mountainous areas that border the Beaverhead Basin is crystalline metamorphic rock. At Beaverhead Rock faulting has brought the Madison Limestone (Mississippian Age, deposited about 359 to 326 million years ago) to the surface. Triassic age rocks consisting of mudstone, siltstone, and limestone are also exposed in this area. The geologic map by Ruppel and others (1993) shows small exposures of Tertiary volcanic rocks exposed through the wedge of gravel outwash on the southwest side of the Beaverhead River.

Ground water has been developed mostly from the Tertiary rocks and sediments which form the upper benches and from the Quaternary deposits that underlie the Beaverhead River, the valley bottom, and tributaries. The Quaternary deposits consist mainly of clay, silts, sands, and gravels. Although this report does not attempt to differentiate between the Tertiary formations, a brief description summarized mainly from Frita and others (2007) is presented here to give the reader a sense of the lithology and depositional environments that constitute this deeper ground-water flow system.

The two main Tertiary units in southwestern Montana are the Renova and Six Mile Creek Formations. The Renova volcanic and volcaniclastic sequence was deposited during the early to middle Miocene in a broad continuous wedge across the basin over a low-relief flood-plain in a system overwhelmed with volcanically derived sediment, which typically is very fine-grained material that does not transmit water well. Non-volcanic facies include sandstone, carbonaceous shale, lignite, and limestone deposited in lakes and streams. Alt (1986) notes that in many places the Renova Formation is capped with a buried layer of red lateritic soil.

During middle Miocene the basin was segmented into several grabens by basin and range style faulting. Sequences of non-volcanic and volcanic sediments known as the Six-Mile Creek Formation filled the Beaverhead and other grabens in southwest Montana during the middle Miocene to late Pliocene. The Six-Mile Creek Formation is generally coarser-grained then the underlying Renova Formation and consists of mudstone, siltstone, conglomerate with local occurrences of limestone, volcanic fallout ash, pyroclastic ash flow tuffs, fallout tuffs, and basalt flows. The Six Mile Creek Formation is generally thickest near the axis of the valleys and thins as it overlaps the uplands.

PREVIOUS INVESTIGATIONS

Uthman and Beck (1998) performed a hydrogeological study in the upper Beaverhead Basin between Dillon and Barretts to study the effects of ground-water development on ground-water and surface-water availability. They defined three aquifers in their study area: a

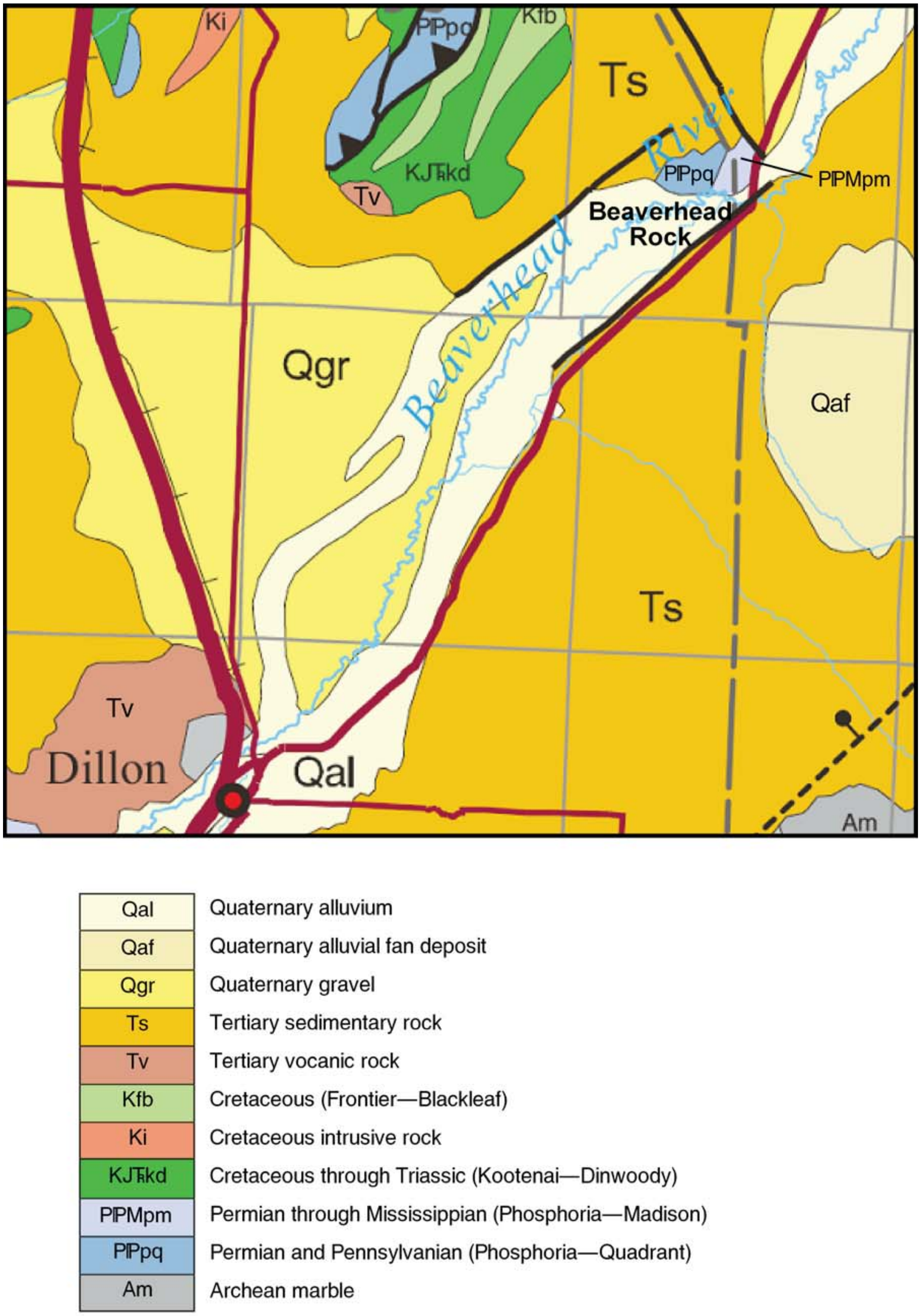


Figure 4. Generalized geologic between Dillon and Beaverhead Rock (Vuke and others, 2007).

bedrock aquifer which provided recharge to the valley-fill aquifers, a lower Tertiary aquifer which produced low water yields, and a coarser Quaternary/upper Tertiary aquifer. Ground-water monitoring revealed that water levels did not steadily decline from 1991 to 1996 as a result of ground-water development, but responded to seasonal recharge. They noted that in irrigated areas, drawdown occurred in the summer in response to pumping but rapidly recovered after irrigation ended. The Quaternary/upper Tertiary aquifer was capable of producing large amounts of water without causing adverse widespread drawdown impacts or adverse depletion of the surface-water system.

Ground-water flow models were used to assess the interaction between surface water and ground water. Two models were used to predict the impact on base flow in the Beaverhead River from ground-water development and drought conditions. They concluded that base flow varied slightly compared to the initial model and that withdrawing water from the system for irrigation did not substantially affect base flow accretions. A third model illustrated that the impact of a 3-year drought had less effect on base flow accretions then irrigation return flow. A fourth model showed that return flow from flood irrigation resulted in the largest accretions to the surface-water system. Their overall conclusion was that ground-water levels and surface-water availability had not been adversely affected by ground-water development from irrigation.

Uthman and Beck's study occurred upstream of Dillon, mainly encompassing the Blacktail Creek and Rattlesnake Creek drainages. Because of their comprehensive evaluation in the upper Beaverhead River, the HB 831 study focused downstream of Dillon mainly between Anderson Lane and Beaverhead Rock.

Weight and Snyder (2007) established a ground-water monitoring network of about 40 wells between Dillon, Montana and about 10 miles north of Dillon. The purpose of the study was to provide baseline information and to re-evaluate ground-water conditions in the Dillon area by comparing ground-water levels in 2005–2006 to those measured by Uthman and Beck (1998). They concluded that ground-water levels declined 2 to 5 ft as a result of a 7-year drought and the shutdown of the East Bench Canal from July 2003 to May 2005.

Based on a potentiometric surface map, Weight and Snyder concluded that the Beaverhead River loses water to the ground-water system until after its confluence with Stone Creek. North of Stone Creek to Beaverhead Rock, where the valley constricts, ground water is forced up to the surface, creating the wetlands near Beaverhead Rock. Ground-water hydrographs for wells monitored in the floodplain correlated with river stage, indicating the connection between the ground-water and surface-water systems.

Ground water in shallow wells monitored on the west side of the river showed a different response compared to two wells completed over 300 ft deep in Tertiary sediments. Weight and Snyder (2007) explained the difference in behavior by a clay layer that separated the two systems. The shallow ground water responded to recharge from the West Side Canal while the deeper flow system exhibited a more regional response from mountain precipitation and recharge. Aquifer testing of a well on the west side of the river yielded a transmissivity estimate of about 300 ft^2/day with a storativity of 0.0001, suggesting a confined aquifer. On the east side of the river Weight and Snyder (2007) report transmissivity estimates of 600 to 6,000 ft^2/day and storativity estimates of 0.008 to 0.01, indicating a range of very leaky to slightly confined conditions in Tertiary sediments.

Weight and Snyder included limited surface-water data in their report, with one graph of the surface-water flow based on the Beaverhead River flow at Dillon. Since GWIC identification numbers were not included in the report, it was not possible to correlate wells with lithologic and drilling information. Snyder was able to correlate some of the wells monitored during the study to GWIC numbers, and this information is currently being entered into GWIC (D. Snyder, written commun., 2008, Hydrogeologist, Montana Bureau of Mines and Geology).

Montana State University performed a surface-water study on the Beaverhead River between Barrett's Diversion and Twin Bridges, Montana (Sessoms and Bauder, 2005). Between 2004 and 2005 they developed a surface-water monitoring network that included 34 stations, 7 of which were on the river's main stem. A "water balance" approach was used to predict and estimate stream flows in the Beaverhead River. They divided the river into five segments: Barrett's Diversion to Dillon, Dillon to Anderson Lane, Anderson Lane to Beaverhead Rock, Beaverhead Rock to Giem Bridge, and Giem Bridge to Twin Bridges. Surface-water hydrographs were created based on the 'predicted water balance,' which was obtained by adding or subtracting any known inflows and outflows to the upstream monitoring location and comparing that to a graph generated from actual flow measured on the Beaverhead River at the downstream end of the segment. In each segment, the researchers identified multiple smaller diversions, tributaries, and sloughs that were not monitored and not accounted for in the predicted water balance. The difference between the water balance predicted value and the actual flow would represent the unaccounted sources and losses of water to the river.

Sessoms and Bauder's data indicated that depending on the time of year, the river may gain or lose water. Cumulative flows monitored within the five reaches showed that overall the river lost water from Dillon to Anderson Lane, Beaverhead Rock to Giem Bridge, and Giem Bridge to Twin Bridges. The river gained flow from Barretts to Dillon and Anderson Lane to Beaverhead Rock. The gain in flow between Anderson Lane and Beavehead Rock was consistent throughout the monitoring period (May–October 2005), and cumulative gains in flow during this period were 28,930 acre-ft.

This study provided insight on the dynamics of the Beaverhead River. However, the data did not account for all inflows and outflows within the five segments. Information was not provided to assess the importance of these missing data and the timing of irrigation was not stated in the report. Ground water was not monitored during the project; therefore, it was not possible to ascertain the role of ground water in the water balance. Although this information can provide a basis for further investigations, a more thorough approach is needed to assess the importance of all components that factor into the water balance.

In 2006, surface-water monitoring continued by Montana State University (Warne and others, 2006) on 14 irrigation diversions. In addition, surface water was monitored in Horse Prairie Creek, a tributary to Clark Canyon Reservoir, to aid in water management. To assess the efficiency of the East Bench Canal nine monitoring sites were established above and below check stations.

This study found that during 2006 more water was diverted than in previous years, citing that this might have been due to greater precipitation and availability of water. They noted discrepancies between irrigation company estimates of diverted water and those measured during their study, with four of the ditches each having cumulative differences of at least 800 acre-ft of water.

Although there were some difficulties with well placement in the East Bench Canal, they determined the greatest cumulative loss, 682 acre-ft per mile, occurred just south in the area of Anderson Lane to about 13 miles north of this area. The problem with this canal efficiency assessment is that since no estimates of evaporation were made and there were unaccounted-for withdrawals, it was not possible to determine if the loss was solely due to seepage.

DATA MANAGEMENT

The MBMG operates the Ground-Water Information Center (GWIC), where data for more than 220,000 water wells are stored and are accessible to Montana's citizenry. In addition to the well-log data, water-level, water-quality, aquifer-test, and other information are managed. Information from the database is accessible through the GWIC website at http://mbmg-gwic.mtech.edu.

Within GWIC, the MBMG groups data used in ground-water projects so that people interested in the project can easily access information. HB 831 Beaverhead Closed Basin case study data can be found on the website's "Ground-Water Projects Page" under the "House Bill 831 Project Data" heading.

For the HB 831 Beaverhead Closed Basin case study, the MBMG compiled well logs and added water-level and other data for 29 wells from hydrologic assessments provided by HB 831 applicants: Cottom Seed Company, Dallasera Ranch, Geoduck Land and Livestock, and the Sitz Angus Ranch. The well logs referred to in these reports were located, locations confirmed from the maps provided in the assessments, and geologic sources designated. More than 1,300 water-level measurements, mostly from the Geoduck Land and Livestock hydrologic assessment, were added to GWIC and online hydrographs are available through the web pages.

The MBMG drilled 13 wells or boreholes for the Beaverhead Closed Basin case study. The logs are available from GWIC and are also accessible under the "House Bill 831 Project Data" heading. Ancillary water-level, water-quality, and aquifer-test data generated by the MBMG form the basis for its surface-water/ground-water analysis. The Beaverhead Closed Basin case study includes three statewide monitoring network wells that provide water-level records beginning as early as October 1992. These records were attached to House Bill 831 project data in GWIC. Wells referred to in this report are denoted by the preface GWIC followed by a number (for example, GWIC 243511).

PROJECT DRILLING

The 13 wells drilled for this project were drilled during January and February 2008. Wells were drilled at three sites (fig. 5). Two of the three sites were chosen to perform aquifer tests near the Beaverhead River and in the Tertiary sediments on the East Bench (Spring Creek drainage site). Two wells were completed at a third location near a pre-existing irrigation well to monitor the effects of pumping an irrigation well on the ground-water system (referred to in site location in table 1 as 'irrigation well monitoring'). Table 1 provides a list of the wells drilled and completion information.

Most wells were drilled using a dual air rotary rig with driven 6-inch-diameter steel casing. Shallow monitoring wells 2C, 2D, and 2E were drilled with an auger rig. The three production wells (2B, 2F, and 3B) were completed by drilling to the desired depth, dropping in a 4-inch-diameter stainless steel screen with a packer at the top, and then pulling the steel casing back to expose the well screen. Except for the monitoring wells drilled at the Beaverhead River site, monitoring wells were completed by drilling past the target zone until water was lost (i.e., to the non-producing zone, usually a clay layer) and perforating the steel casing with four vertical rows of perforations in the water-producing zone. This approach left the bottom of the well open, which could interfere with the well performance by allowing sediment to heave into the well. Heaving was avoided by completing wells in clay layers, which are less likely to heave into the well.

Wells 2C, 2D, and 2E were completed with 2-inch-diameter polyvinylchloride (PVC) well casing

Table 1. Construction details of wells drilled as part of this project.

GWIC ID	Well Name	Site Location	Well Diameter	Completion Materials	Well Depth	Screen Interval	SU	Elevation (Top of casing)
242413	Well 1A	Irrigation well monitoring	2"	PVC	400	345.6-365.8	4.00	5031.33
242414	Well 1B	Irrigation well monitoring	6"	Steel	190	175-185	-	5031.87
242403	Well 2A	Near Beaverhead River	4"	PVC	95	74-94	3.95	4850.94
242404	Well 2B	Near Beaverhead River	6"	Steel	28.5	18.5-28.5	2.20	4848.55
242407	Well 2C	Near Beaverhead River	2"	PVC	24	13.5-23.5	3.00	4850.29
242415	Well 2D	Near Beaverhead River	2"	PVC	24	13.5-23.5	3.42	4850.51
242417	Well 2E	Near Beaverhead River	2"	PVC	24	13.5-23.5	3.42	4851.72
242406	Well 2F	Near Beaverhead River	6"	Steel	80	63-83	1.85	4849.11
242408	Well 3A	Spring Creek Drainage	6"	Steel	515	503-511	2.00	5183.92
242409	Well 3B	Spring Creek Drainage	6"	Steel	290	261.3-289.3	1.80	5183.55
242410	Well 3C	Spring Creek Drainage	6"	Steel	288	275-285	1.46	5185.30
242411	Well 3D	Spring Creek Drainage	6"	Steel	155	143-153	2.62	5186.43
242412	Well 3E	Spring Creek Drainage	6"	Steel	290	278-288	1.67	5193.21

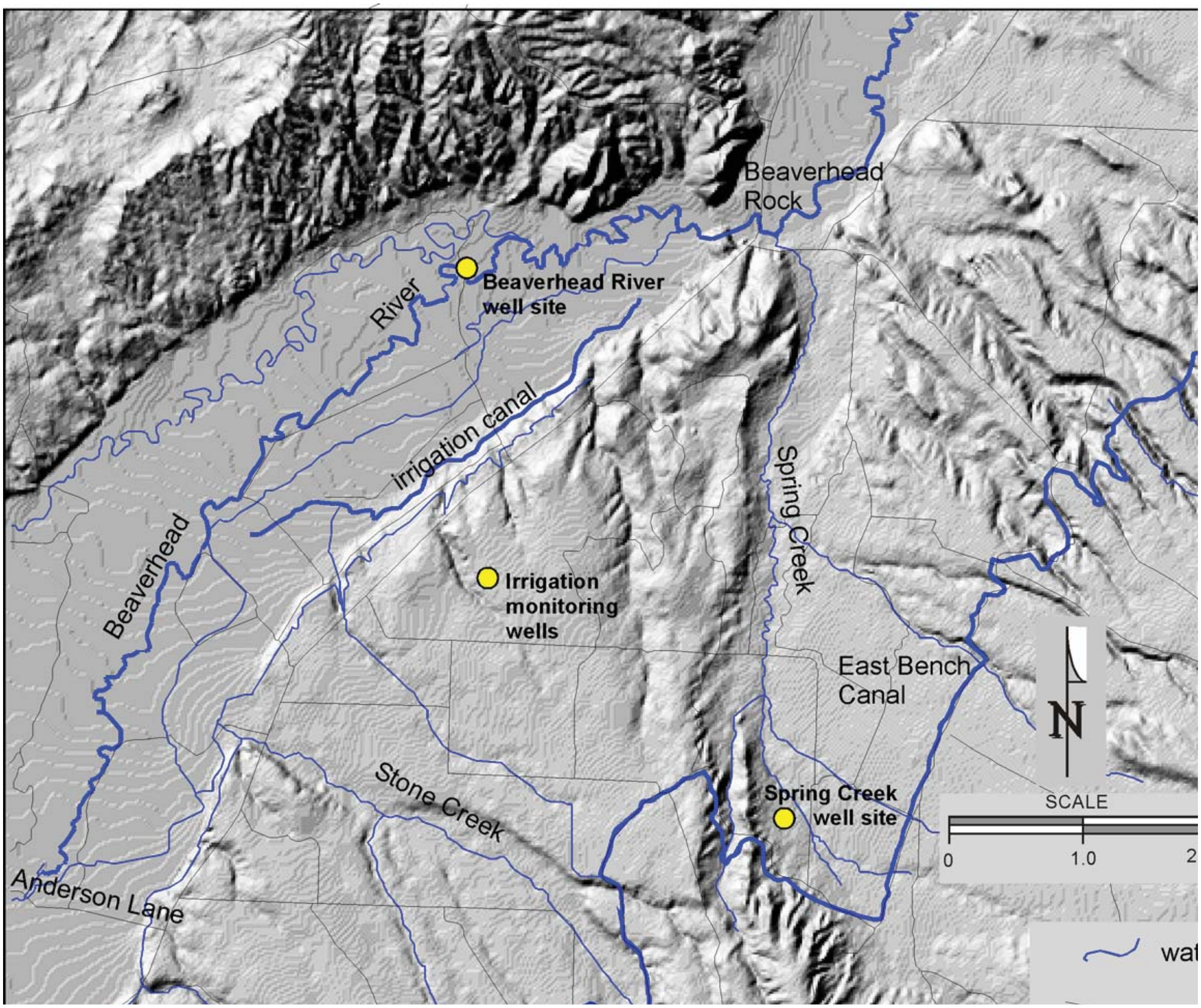


Figure 5. Drilling occurred in three locations in order to perform aquifer tests and to monitor the effect of an irrigation well on the ground-water system.

and 10-ft-long factory-slotted well screen. Well 2A was completed using 20 ft of 4-inch-diameter PVC with sawed slots instead of stainless steel well screen.

Wells Drilled near the Beaverhead River

The first well site was near the Beaverhead River and was designed to assess the impact of the river on ground-water availability, under the assumption that the drawdown cone would intercept the river. The goal was to drill a shallow production well (well 2B), a deep monitoring well (well 2A), and three shallow monitoring wells (wells 2C, 2D, and 2E) on the north side of the Beaverhead River (fig. 6).

Well 2A was drilled to a depth of 100 ft and screened from 74 to 94 ft. In this area, the uppermost 25 to 30 ft was composed of sands and gravels, typical of sediments deposited in an alluvial environment (fig. 7). Beneath this shallow alluvium a 30-ft-thick brownish-gray silty clay layer was encountered. Both the sand and gravel at the surface and the clay layer were interpreted to be Quaternary deposits. The clay layer was underlain by indurated siltstone and sandstone, interpreted as Tertiary deposits. The discovery of the clay layer led to drilling of a deeper production well (Well 2F) on the south side of the river in the Tertiary sediments at a depth of 83 ft. In this well, clay was noted at 25 ft below ground surface with clay intermixed with siltstone to about 50 ft below ground surface. Well 2F was drilled so that an aquifer test could be conducted on the deeper aquifer in addition to the aquifer test in the near-surface alluvial aquifer. Conducting these two aquifer tests would test the interconnectivity of these two aquifers.

Wells Drilled in the Spring Creek Drainage

The second drilling area was located on Tertiary sediments near the East Bench Canal in the Spring Creek drainage (fig. 8). This site is located ~ 0.5 miles from an irrigation well (IR4; GWIC 220025) that is 500 ft deep and screened through multiple aquifers. One of

Figure 6. Aerial photograph showing the locations of the wells used in the aquifer tests near the Beaverhead River.

The lithology in well 3A (GWIC 242408), drilled to a depth of 515 ft, was dominated by a brown to tan silty sandstone with less abundant layers of clay and gravel. Several zones of clean coarse sand, some of which contained mica, were encountered; the greatest thickness of these was about 15 ft. There were no significant clay layers noted during drilling, but the absence of water and cuttings indicated clay layers at 220–235, 305–335, 340–350, 383–395, and 422–450 ft below ground surface.

the goals for completing wells at this location was to perform an aquifer test to examine the differences in aquifer-test data between a multiple screened well (the irrigation well) and a well screened in a discrete aquifer unit. In addition to acquiring basic hydrogeologic information on the production aquifer, another goal of well placement at this location was to determine the interconnectivity between the production aquifer and aquifers located above and below the production zone. This location was also chosen to assess the long-term impacts of the East Bench Canal on the groundwater resources in this area.

During drilling three aquifers similar to those described in the well log for IR4 were encountered. The production well (well 3B), well 3C, and well 3E were completed in the middle (production) aquifer. Well 3A was completed in a deeper aquifer and well 3D was completed in a shallow aquifer.

Figure 7. The uppermost 25 to 30 ft of sediments in well 2A consisted mainly of sands and gravels.

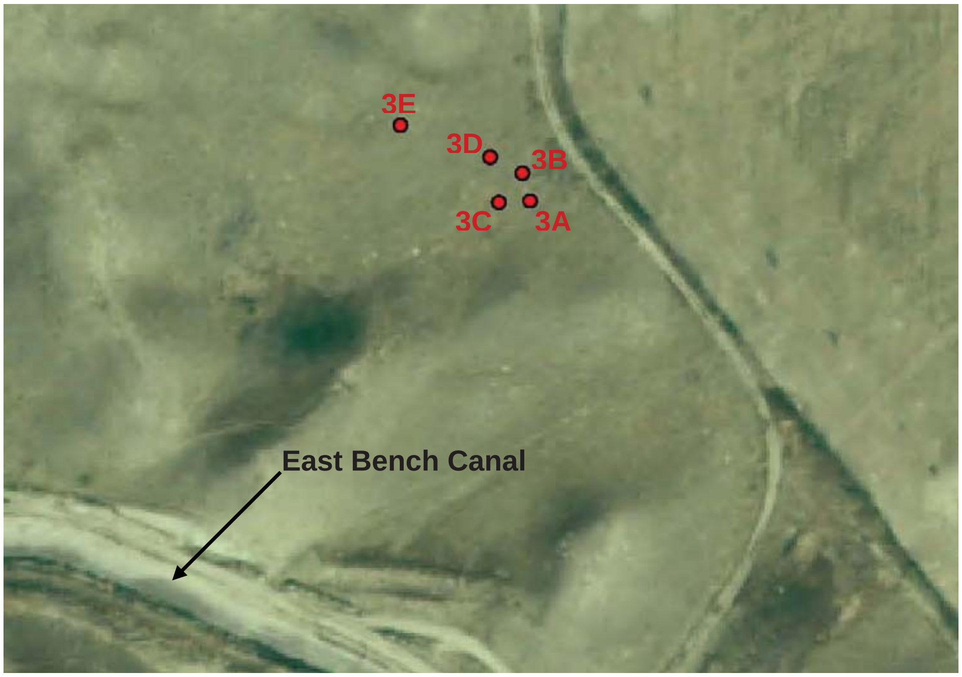


Figure 8. Aerial photograph showing the locations of the wells used in the aquifer test near the East Bench Canal.

Where some cuttings were obtained, these zones were associated with clay and/or fine gravel-size sandstone in a silty clay matrix.

Wells Drilled to Monitor the Effects of a Nearby Irrigation Well

Monitoring wells 1A and 1B were installed about 200 ft from IR1 (GWIC 204038), a pre-existing irrigation well (fig. 9). Well 1B was drilled to a depth of 190 ft and well 1A was drilled to a depth of 400 ft. These wells will monitor the effects on the ground-water system throughout the growing season from pumping a nearby irrigation well.

The lithology in well 1B consisted mainly of interbedded sand and gravel. Yellow-orange clay was prevalent in the upper 100 ft. Close examination of the sands at 150 ft showed that they were actually composed of sandstone fragments, indicating this unit was lithified and then fragmented during drilling. At 210 ft below ground surface lignite and rock fragments were noted. Below this a sandy marl was logged along with some igneous rock fragments. The marl was lithified and consisted of a mixture of calcium carbonate and silt with some quartz sand. Clean quartz sand was noted at 295–305 and 340–350 ft below ground surface. These types of sands are usually associated with alluvial deposition and yield abundant water. The well was completed at 400 ft below ground surface, at which depth a reddish-brown clay slurry was encountered.

AQUIFER TESTS

Methods

Constant-rate and step-drawdown pumping (aquifer) tests offer the most powerful tools for determining hydrogeologic properties of an aquifer (Driscoll, 1986). Analysis of aquifer-test data can yield estimates for transmissivity (T) and storativity (S), which are necessary for estimating aquifer properties and are important variables in algebraic and numerical models. These models can then be used to predict the impact of perturbations to the ground-water system. The data that is required for aquifer tests include:

- distances between the pumping well and monitoring wells;
- aquifer and confining layer thicknesses;
- well completion information;
- water levels in the monitoring wells before, during, and after pumping;
- water levels in the pumping well before, during, and after pumping;
- barometric pressure during the time of water-level measurement; and
- pumping rate.

All of the aquifer tests were completed during March 2008 while there was no irrigation occurring and prior to major runoff events.

The aquifer testing on the near-stream shallow (well 2B) and deep (well 2F) completed wells took advantage of two pre-existing wells in the area, which were also monitored during the aquifer tests (wells BR6 and BR7). The distances of the observation wells to each pumping well, and the well depths are listed in table 2. The distances of the observation wells to the pumping well, the well depths, and the screened/perforated intervals are listed in table 3 for the aquifer test performed at the Spring Creek drainage site.

Water Level Measurements

Water levels in the monitoring and production wells were monitored continuously using dataloggers (Solinst Leveloggers). Dataloggers in the production wells recorded water level at 15-second intervals throughout the aquifer test and recovery time periods. Dataloggers in the monitoring wells recorded water-level data at 1-minute intervals throughout the aquifer test and water-level recovery time periods. Dataloggers in the

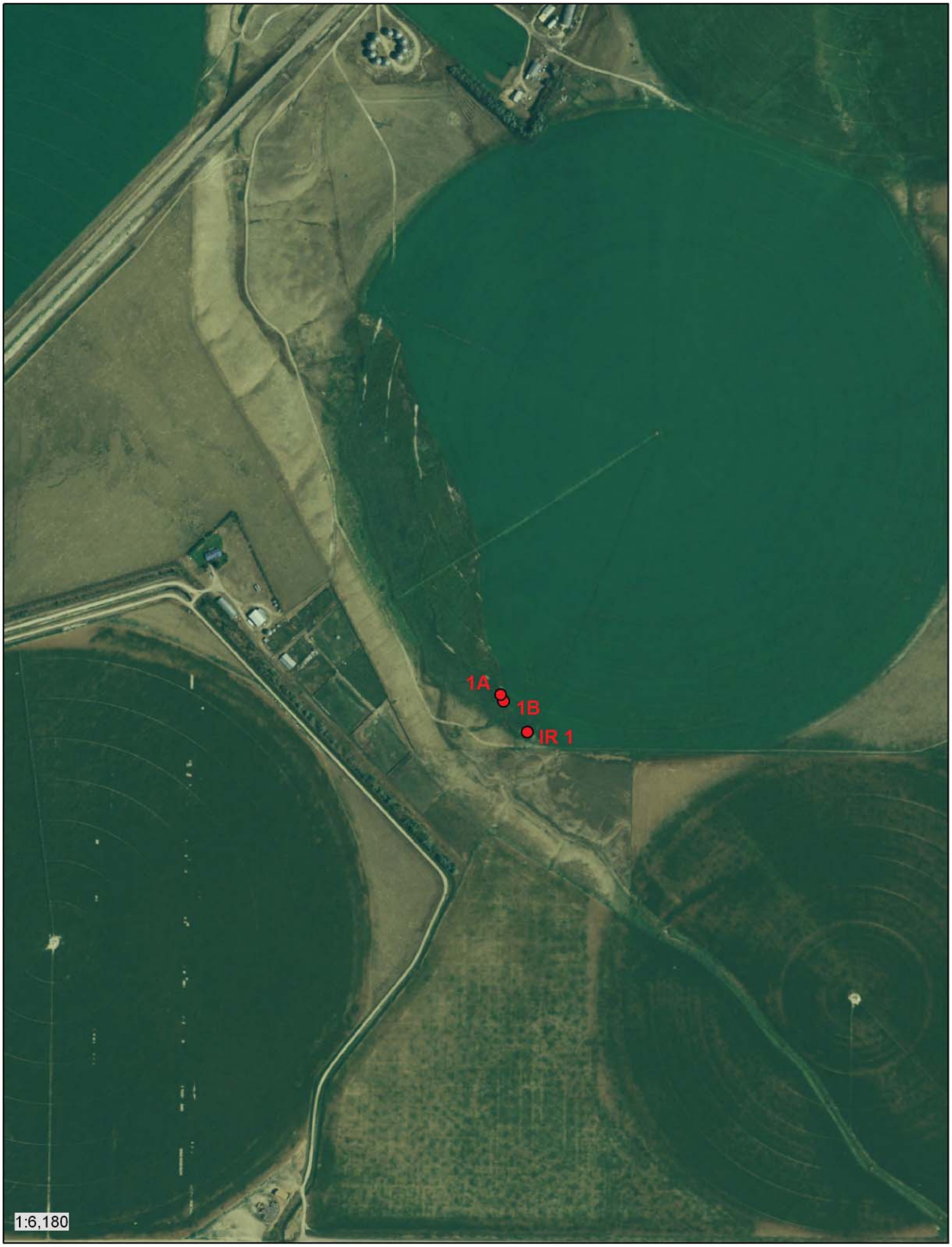


Figure 9. Aerial photograph showing the location of wells 1A and 1B drilled to monitor the effects of pumping from a nearby irrigation well (IR1).

Table 2. Summary of the well depths, distances of each monitoring well to the pumping wells 2B and 2F, and the depth of the screen/perforated intervals.

Observation Well	Well Depth (ft)	Distance from Well 2B (ft)	Distance from Well 2F (ft)	Screened/Perforated Interval (ft bgs)*
Well 2A	95	19.0	199.5	74–94
Well 2B	28.5	—	216.5	18.5–28.5
Well 2C	23.5	44.9	174.6	13.5–23.5
Well 2D	23.5	17.8	217.6	13.5–23.5
Well 2E	23.5	28.9	256.3	13.5–23.5
Well 2F	83	216.4	—	63–83
Well BR6**	18	177.1	46.9	13–18
Well BR7**	18	139.2	278.4	13–18

*bgs, below ground surface.
**Well depths and screened intervals estimated.

Table 3. Summary of the well depths, distances of each monitoring well to the pumping well 3B, and the depth of the screen/perforated intervals.

Observation Well	Well Depth (ft)	Distance from Well 3B (ft)	Screened/Perforated Interval (ft bgs)*
Well 3A	515	32.0	503-511
Well 3B	289.6	-	261-289
Well 3C	288.5	41.2	275-285
Well 3D	153	39.8	143-153
Well 3E	290	145.8	278-288

*bgs, below ground surface.

monitoring wells were installed at least 3 days prior to initiation of the aquifer tests to determine if there was a background trend to the ground-water levels. Water levels were also recorded after the pumps were shut off for a minimum of 72 hours to monitor water-level recovery. In addition to the water-level measurements, dataloggers (Solinst Barologgers) recorded the barometric pressure at 1-minute intervals (at each location) throughout the aquifer test and recovery time periods. The water-level measurements were corrected for changes in the barometric pressure during the time of measurement using software provided with the dataloggers.

Discharge Measurements

Discharge measurements were made using two different types of inline flow meters. Discharge measurements for the two aquifer tests at the river site were monitored using an ultrasonic flow meter that attached to the outside of the discharge pipe. At the site near the East Bench Canal the discharge was monitored using an impeller-type flow meter. Discharge measurements were taken periodically and recorded in a field notebook. After installing the pump in each of the production wells, step-drawdown aquifer tests were conducted to determine the maximum pumping rate for the constant-rate aquifer tests. Constant-rate aquifer tests were used to determine aquifer parameters. The aquifers were allowed to recover prior to initiating the constant-rate aquifer tests, which were conducted for a minimum of 72 hours. At the river site, a staff gauge was used to monitor the water level in the river, and staff gauge measurements were recorded at the same time as discharge measurements.

Aquifer Test Results

Although the goal of the aquifer tests was to obtain estimates of the hydrogeologic parameters of the associated aquifers, it is also instructive to visually inspect the drawdown data. The drawdown data for six monitoring wells during the shallow aquifer test at the river site are presented in figure 10. All four of the wells that were completed in the shallow aquifer showed immediate responses to pumping. This response was even recorded in BR6, which was across the river from the pumping well. In fact, neither the shallow observation-well data nor the pumping-well data (not presented) indicate that a constant head boundary (the river) was encountered by the drawdown cone even though almost all the wells are within 80 ft of the river and the drawdown cone extended under the river. If a constant head boundary were intercepted, the water levels in the wells would cease to decrease and the curve would become flat. The results of this aquifer test suggest that the river and the shallow aquifer are not hydrogeologically connected, which is highly unlikely. Sediments that underlie the riverbed typically consist of sands and gravels with clays intermixed. Depending on the riverbed morphology and variability in stream flow across a channel, zones of lower permeability material such as silts and clays can accumulate in lower energy sections of the river. The results of this aquifer test highlight the heterogeneity of these systems and the danger of depending too heavily on one aquifer test to define a system.

No response was observed in the two deep wells (2A and 2F) while pumping well 2B (fig. 10). Water-level data from these two deep wells plot on top of each other. This lack of a response to pumping suggests that the shallow aquifer is hydrogeologically isolated from

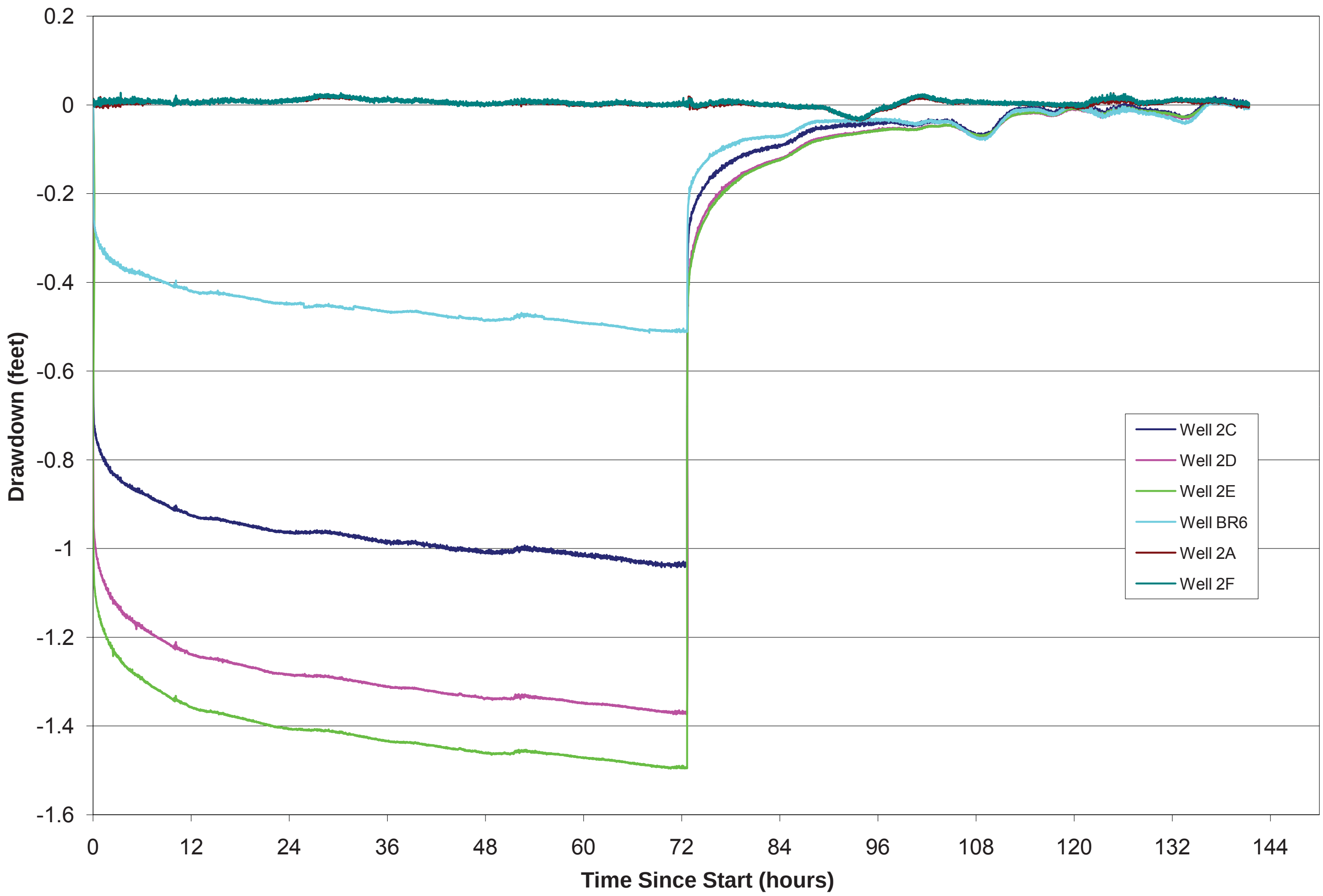


Figure 10. Drawdown data for the monitoring wells surrounding well 2B (pumping well) at the river site. Drawdown data is relative to the water level just prior to the start of pumping. Pumping rate was approximately 100 gpm. Maximum drawdown in the pumping well was approximately 4.9 ft. Static water level (pre-pumping) in the deep aquifer was approximately 4.5 ft lower than the static water level in the upper aquifer.

the deeper aquifer. The confined condition of the deeper aquifer is supported by the existence of the clay layer and the fact that the static water levels in the deep wells were approximately 4.5 ft lower than the static water levels in the shallow wells.

All of the shallow wells show similar decreases and increases in water levels throughout the period of record, which are most likely due to fluctuations in river stage. Data from BR7 showed a gradual downward trend that did not vary throughout the 30-day monitoring period, which indicates that the data were compromised; thus these data were not used.

The drawdown data for six observation wells during the deep aquifer test at the river site are shown in figure 11. Of the six observation wells, only the well completed below the clay layer (2A) showed a response to pumping. There is no indication that either the river or the shallow wells were affected by pumping the deep aquifer. Similar to the shallow aquifer test, these data indicate that the deep system is hydrogeologically disconnected from the shallow system.

All of the shallow wells had similar water-level fluctuations, but these appear to be caused by fluctuations in the river stage and not associated with pumping. Staff gauge measurements of the river were not collected at the same frequency as the water-level data in the wells, so the two data sets cannot be directly compared. However, the water levels in the river varied sporadically within about 0.08 ft, which is similar to the range of water-level fluctuations observed in the wells. Data from well BR7 were not presented for the reason stated above.

All of the observation wells for the aquifer test in the Spring Creek drainage showed a response to pumping (fig. 12). The five wells at this site were completed in three distinct water-producing zones within the Tertiary sediments (table 2). The fact that all three aquifers responded to pumping indicates that all three aquifers are hydrogeologically connected. However, the three wells completed within 32 to 41 ft away from the pumping well (wells 3A, 3C, and 3D) all responded differently, which indicates that there is significant variability in vertical hydraulic conductivity at this site. The Tertiary sediments consist of discontinuous layers of clay, silt, sand, and gravel, so a lower vertical hydraulic conductivity was expected.

Aquifer Test Analysis

The aquifer test data were analyzed using com-

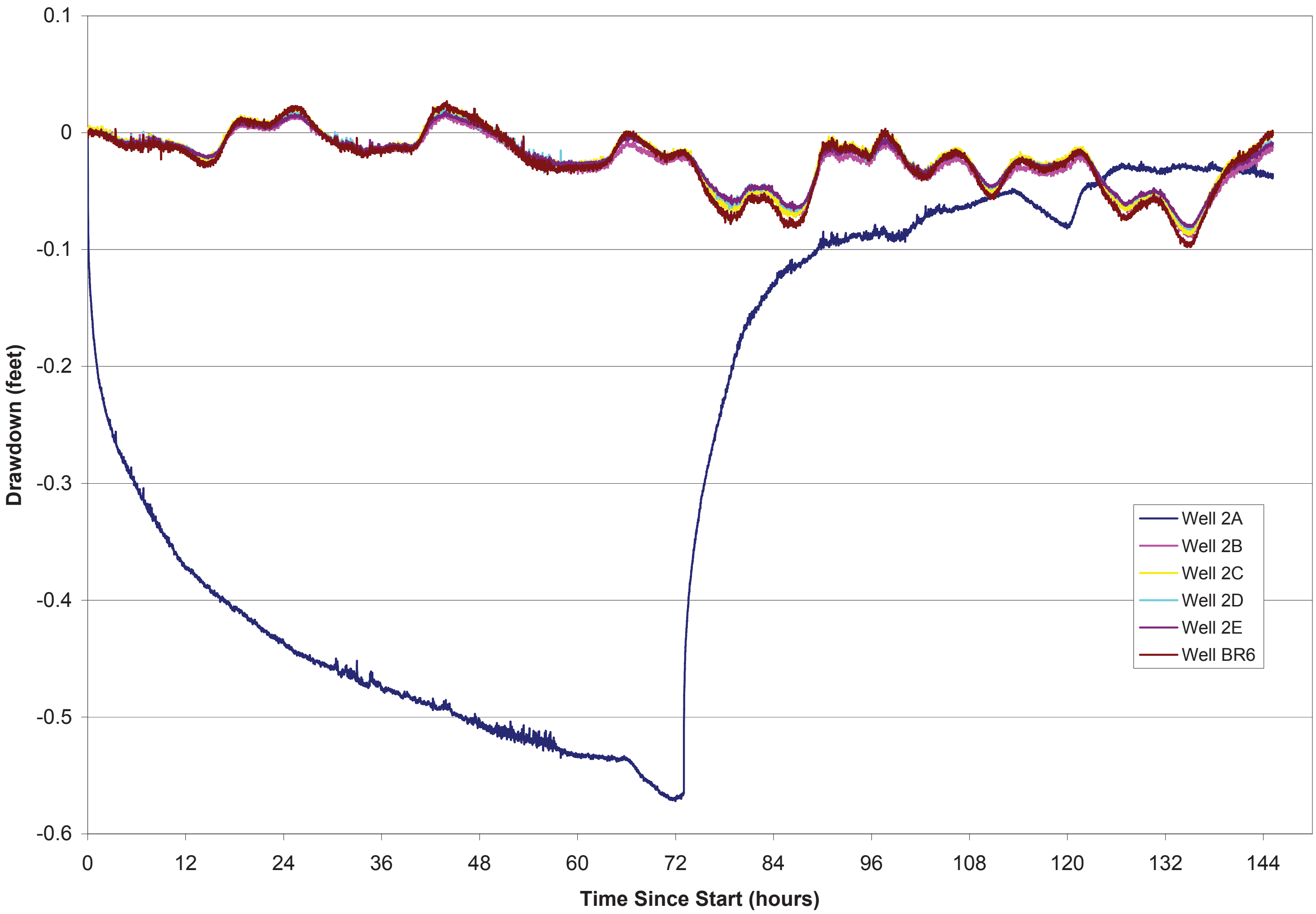


Figure 11. Drawdown data for the monitoring wells surrounding well 2F (pumping well) at the river site. Drawdown data is relative to the water level just prior to the start of pumping. Pumping rate was approximately 300 gpm. Maximum drawdown in the pumping well was approximately 8.6 ft. Static water (pre-pumping) level in the deep aquifer was approximately 4.5 ft lower than the static water level in the upper aquifer.

mercially available software (AQTESOLV). AQTESOLV allows the user to employ a variety of different solution methods to estimate aquifer parameters. Since the shallow aquifer at the river site represented a relatively straightforward unconfined aquifer, the most appropriate solution method for these data was the Theis method (Theis, 1935). AQTESOLV has made modifications to the Theis method, which allows for analysis of data from partially penetrating wells, as in this case. A summary of the transmissivity (T) and storativity (S) estimates using the Theis solution method for the shallow aquifer at the river site are presented in table 4. The estimates of transmissivity range from 4080 to 7420 m2/day. Transmissivity values are generally considered significantly different when they vary by an order of magnitude, so the estimated transmissivities from the different wells are not considered to be significantly different. A representative solution fit for the Theis solution is presented in figure 13 for well 2C.

The analysis of the aquifer test data is still on going for pumping wells 2F and 3B, which were completed in the Tertiary sediments. Several solution methods have been employed to analyze the data using AQTESOLV, but curves generated to date have not matched the data well. We are currently working to identify other software or matching techniques that better represent the data. Preliminary results indicate that the hydraulic conductivity of the Tertiary deposits is on the order of 1,000 to 5,000 m^2/day and the vertical hydraulic conductivity is approximately an order of magnitude lower than the horizontal hydraulic conductivity at the East Bench site.

Table 4. Transmissivity (T) and storativity (S) estimates for the shallow aquifer at the river site.

Well	T (m^2/day)	S
2C	5520	7.18E-7
2D	4210	3.24E-6
2E	4080	3.60E-7
BR6	7420	5.02E-5

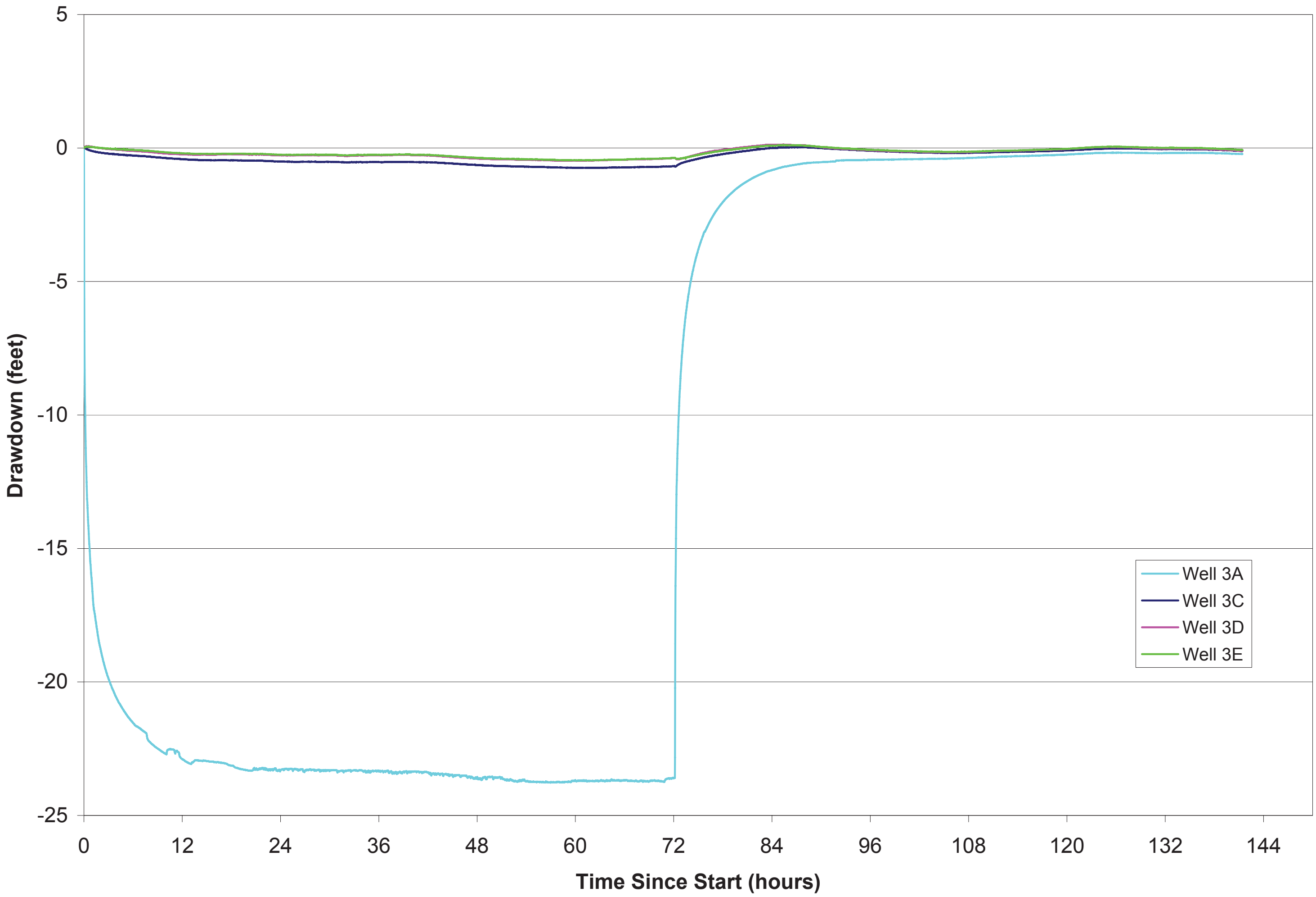


Figure 12. Drawdown data for the monitoring wells surrounding well 2A at the East Bench Canal site. Drawdown data is relative to the water level just prior to the start of pumping. Pumping rate was approximately 100 gpm. Maximum drawdown in the pumping well was approximately 75 ft.

WATER CHEMISTRY

Water-quality samples have been collected from three wells and three surface-water sites to date. Ground-water samples were collected from three MBMG wells during aquifer testing. A surface-water sample was collected of the Beaverhead River near the MBMG river test location (fig. 6). Another sample was collected from the Co-op ditch, which is a diversion from the Beaverhead River near Beaverhead Rock, but at the time of sampling the flow in the ditch was primarily from a spring that discharges to the ditch. Specific conductance, pH, and temperature were measured in the field and samples were collected to analyze for major cations and anions, trace metals, oxygen 18, deuterium, and tritium. Isotope analyses will provide information on the source and age of the ground water. Cation, anion, and trace metals are being analyzed by the MBMG laboratory and the results are not yet available. Isotope analysis is being performed by the University of Waterloo Environmental Isotope Laboratory. Table 5 summarizes the field parameters collected during sampling.

A pH of above 7.5 was noted in all the surface-water samples and in well 2F (242406), which was completed in Tertiary sediments underlying the Beaverhead River alluvium. The specific conductance of 692 microsiemens/cm measured in the Beaverhead River at Beaverhead Rock falls within the range of 374 to 921 microsiemens/cm reported by the USGS (n=98, data from 1998–2007; USGS, 2008). The pH at this location was 7.54, lower than the range of 7.8–8.6 reported by the USGS (n=20, data from 1998-2007; USGS, 2008).

The warmest temperature, 27.8°C, was found in the Co-op ditch. A temperature of 27°C was reported in this area by Sonderegger and Bergantino (1981) in their survey of the geothermal resources of Montana. The temperature of 15.7°C noted in well 3B (242409) was warmer than the wells sampled near the Beaverhead River (242404 and 242406).

Additional water-quality samples will be collected and the information will be used to determine the physical and chemical differences in the ground-water flow systems and ground-water age, and will help assess the interaction between surface water and ground water in this area.

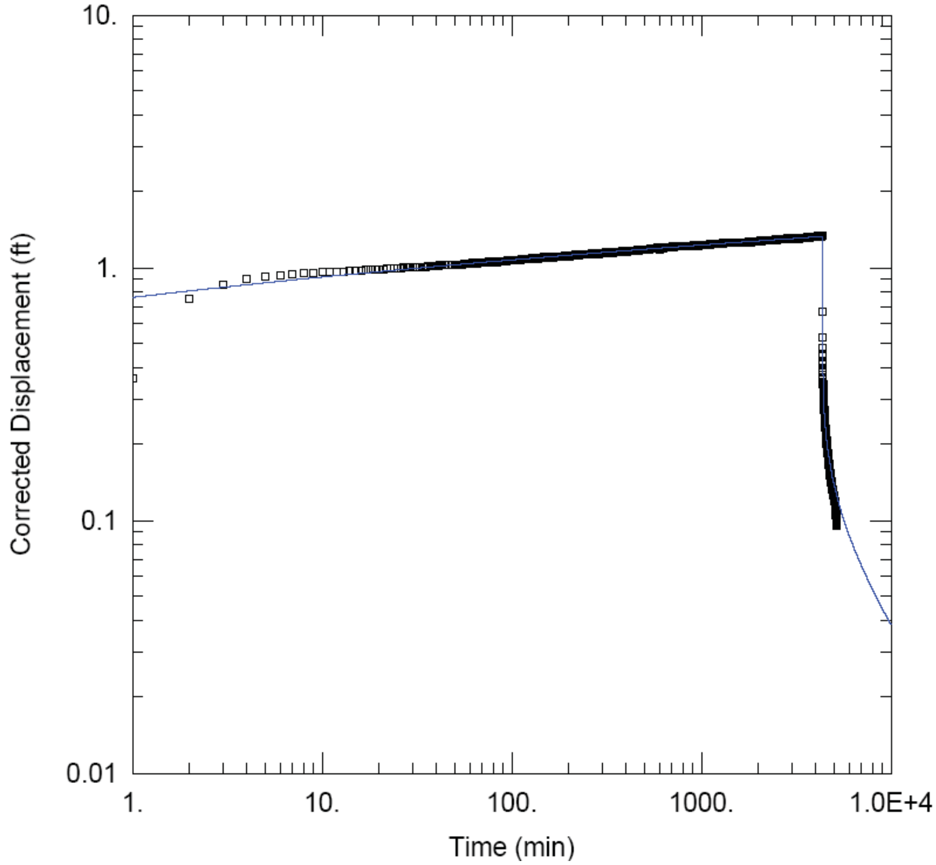


Figure 13. Unconfined Theis solution curve fit for the drawdown in well 2C during the shallow pumping test at the river site.

HYDROGEOLOGY

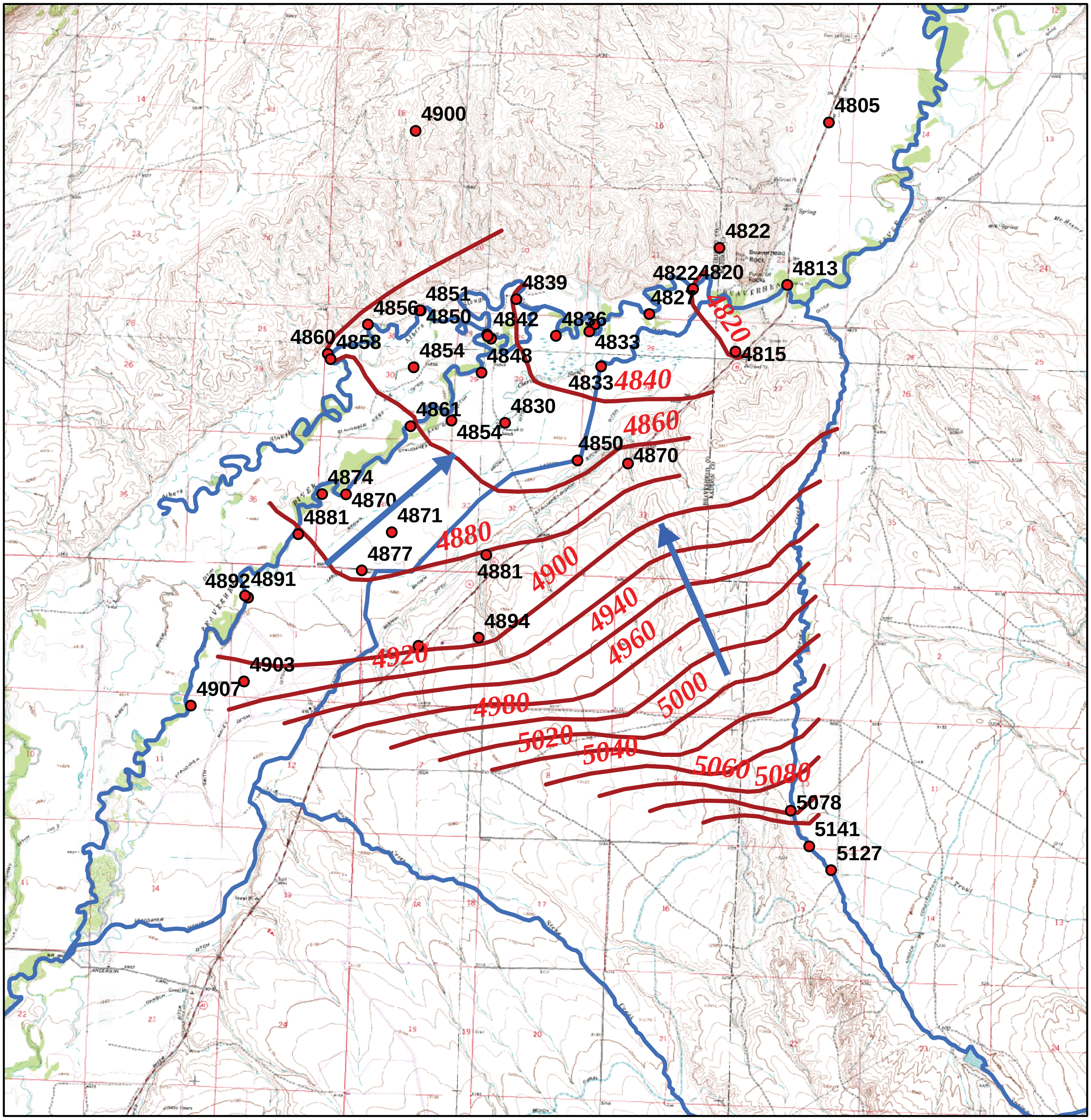
Ground-Water Flow Map

A composite potentiometric map was constructed for the northern part of the study area using water levels measured during August 2006 by Water Rights, Inc., and water-level information from drillers' logs (fig. 14). Wells in the valley bottoms were completed in the Quaternary alluvium, or in some cases the underlying Tertiary sediments. Wells on the East Bench and the area north of the river were completed in Tertiary sediments. The ground-water elevations are steepest in the upland areas, with a gradient of about 0.015 ft/ft. In the valley bottom the gradient flattens to about 0.004 ft/ft as noted by the wider spaced ground-water contour lines. This is a function of the transmissivity of the sediments. The gentler ground-water gradient in the valley indicates that the alluvium is more transmissive then the Tertiary sediments.

Ground-water flow in the upland areas is towards the valley, providing recharge to the alluvial system. In the valley, ground water flows either parallel or towards the stream, indicating that the river is gaining water from the ground-water system.

Table 5. Field data collected from water samples.

GWIC No.	Description	Date	pH	Specific Conductance (microsiemens/cm)	Temperature (°C)
242404	Well 2B	3/18/08	6.80	1162	7.3
242406	Well 2F	3/25/08	7.79	782	8.9
242409	Well 3B	3/25/08	6.81	331	15.7
Surface water	Beaverhead River near river test site	3/18/08	8.08	634	5.6
Surface water	Beaverhead River near Beaverhead Rock (USGS Gauge)	3/25/08	7.54	692	8.6
Surface water	Co-op Ditch	3/18/08	7.52	736	27.8



Legend

Ground-water elevation

Ground-water elevation contours

Streams

Direction of flow

0 1 2 Miles

N

Figure 14. Composite ground-water flow map showing the direction of ground-water flow in the valley bottom is towards or parallel to the river.

Hydrogeologic Cross Sections

Two hydrogeologic cross sections are provided in draft format to help explain the ground-water flow system and the nature of the geologic materials. Drillers' logs from the GWIC database, which were used to compile these cross sections, often contain inaccuracies with regard to location and misinterpretation of the geology that limits their usefulness in some cases. Even so, these cross sections provide interesting insight into the hydrogeologic system. The main geologic units of interest in this study were the valley-fill composed of Quaternary sands and gravels, the clay layer believed to be Quaternary age, and the Tertiary sediments.

The first cross section (A–A′) was composed as a longitudinal profile down the axis of the Beaverhead River valley from Dillon to Beaverhead Rock (fig. 15). The endpoints of the cross section were GWIC 863334, logged by Uthman and Beck (1998), and well 242403, logged by the MBMG as part of this investigation. From GWIC 242403 the cross-section line was bent towards the east to cross the valley in the Beaverhead Rock area. Survey grade location and elevation information were available for these two wells. The well logs showed sands and gravels interbedded with finer-grained units such as

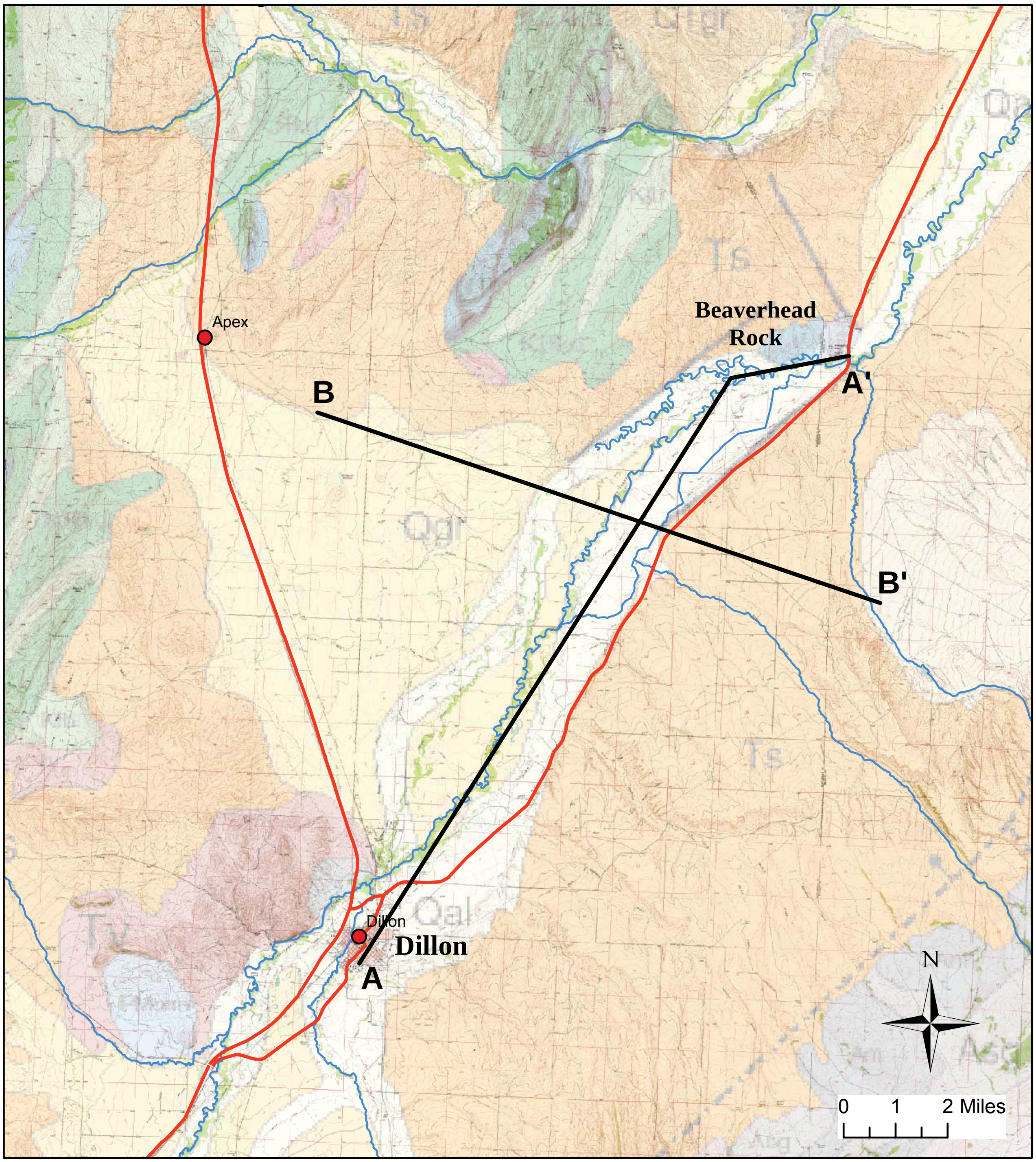


Figure 15. Location of cross section lines for cross section A – A′ and B – B′.

clays and silts. Several wells also encountered indurated siltstones and sandstones interpreted as Tertiary deposits. Because of the variability in the geologic descriptions (such as 'clay stone', 'hard tan', 'mudstone bedrock', and 'shale') and lack of deep wells in portions of the valley bottom, there was no attempt to approximate the depth of the Tertiary sediments below ground surface. A simplification was made to identify permeable and semi-impermeable units. The screened interval of the well and water level were also plotted on the cross section, if available. The preliminary cross section is shown in figure 16.

The cross section indicated that within the valley-fill there are permeable and semi-impermeable units. Consistent throughout the cross section were sands and gravels in the uppermost 30 to 40 ft, illustrated in the yellow. This aquifer is unconfined and is directly connected to the surface-water system. An unconfined aquifer is usually near the land surface, and there are no impermeable or confining geologic units overlying the aquifer. The water level in a well completed in an unconfined aquifer is the same as the water table outside the well (Fetter, 1994). Examples of wells completed in this aquifer are GWIC 109189, 108981, and 238709.

The drillers' log for the deeper wells showed semi-impermeable clays and silts beneath the alluvial aquifer that act as a confining or semi-confining layer(s). A confined aquifer is bounded by less permeable confining layers. Because of the pressure created in a confined aquifer, the water level in a well drilled into a confined aquifer will rise above the top of the aquifer. The cross section shows that in cases where the well was screened in permeable zones beneath these confining units, the water level rose above the permeable zone and/or confined/semi-confined aquifer. These wells completed in the deeper confined/semi-confined aquifer system are most likely not in direct communication with surface water. Examples of wells completed in the confined/semi-confined aquifer are GWIC 109439, 109361, 141233, 208123, and 242403.

Near Beaverhead Rock, the Madison Limestone is exposed on the surface and constricts the Beaverhead River Valley. Well logs in this area contained more clay, 'shale', and limestone. It is in this area that the clays/silts underlying the alluvium appear to merge into the clays weathered from the underlying bedrock. In any case, the valley constriction forces ground water to the surface. This is evident in increases in stream flow (Sessions and Bauder, 2005) and the wetland area just upstream from Beaverhead Rock.

A cross section that transects the valley (B–B′) was constructed in a northwest–southeast orientation. Although this cross section is in the early stage of construction, it illustrates the variability of the Tertiary sediments (fig. 17). Several wells on the northwest part of the cross-section line penetrated through volcanic rock (GWIC 199124, 192298, 204226, and 220080), which is also exposed on the surface (Ruppel and others, 1993). The drillers' information for GWIC 204226 and 220080 indicates that the volcanic rock is capable of producing large amounts of water. GWIC 220080 was pumped at 1500 gpm for 8 hours with only 3.5 ft of drawdown, and the water level in GWIC 204266 dropped by 10 ft after pumping the well at 750 gpm for 8 hours.

Clay appears to be more dominant in the drillers' logs west of the river. The clay does produce confining aquifer conditions as evidenced by wells 220058 and 108966, where water levels rose above the top of the lower aquifer where the wells were screened. Clay layers were noted over 100 ft thick in several of the wells.

Only two well logs were available on the valley bottom in this vicinity. Both wells were located west of the river and also show clay acting as a confining unit, as noted in cross section A–A′. The depositional history of the clay found underlying the Beaverhead River Valley is unknown. R. Thomas (oral commun., 2008, Professor of Geology, Western Montana College, Dillon, Montana) believes that the clay is Pleistocene in age, which encompasses the Quaternary period from 1.8 million to 11,550 years ago. To date, there is no evidence of ice in the valley that may have blocked the drainage and caused lake formation, which is usually associated with lower energy deposits such as clay. Thomas speculates that the clay may have been deposited in sag ponds on an outwash plain. A more thorough examination of the clay found in the valley bottom, which would include age dating and determining the spatial distribution, would help decipher its depositional history. If the clay layer was deposited at the same time as the Tertiary sediments of the benches, it would interfinger with the permeable units of the benches and form less of a barrier to ground-water flow among the benches, the alluvial aquifer, and the river. If the clay was deposited after the Tertiary sediments, the clay may be forming a cap on the older deposits that is restricting ground-water movement from the benches to the river. Ground water and surface water will be monitored in this area through the current growing season in an effort to elucidate the connection between surface and ground waters. We also hope to work with the Montana Tech Geophysical Department to complete a geophysical analysis of the area that will delineate the aerial extent of the clay layer.

The Tertiary deposits on the east side of the river consist of clays interbedded with sands. It is most likely that some of the sands are actually semi-consolidated sandstones as logged by the MBMG in GWIC 242408. Ground water in wells also indicates confined/semi-confined conditions where water levels rose above the top of deeper permeable zones such as in GWIC 242408 and 226863. Unfortunately no completion records for GWIC 201621 or water-level information for GWIC 201953 were available.

Figure 16. This cross section shows the permeable and less permeable units in a longitudinal profile from Dillon to Beaverhead Rock.The Beaverhead River alluvium, shown in yellow in the cross section, is about 30 to 40 ft thick and is underlain by a less permeable clay unit.

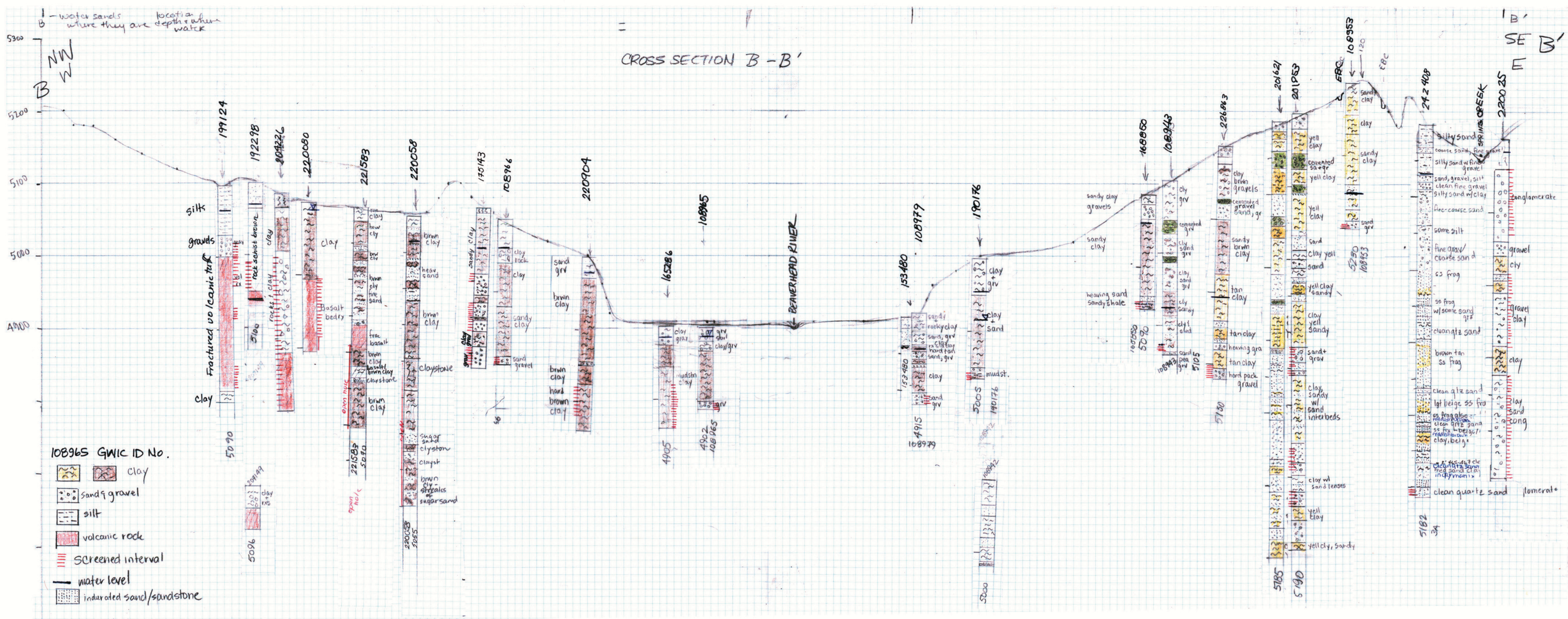


Figure 17. This preliminary sketch shows the variability in the Tertiary sediments. Ground water in wells completed in deeper aquifer units exhibit semi-confining to confining conditions.

STREAM DEPLETION

The connection between ground water and surface water has always been a fact in the physical system, but with greater demands on surface water throughout the western United States, ground-water extraction has become more highly regulated. The basic question for water managers has been the safe yield, sustained yield, or sustainable yield of the watershed; that is, the upper limit of ground-water development in a given watershed. HB 831 recognizes the need to address senior water-rights holders while allowing for the evaluation of ground-water development and offsetting stream depletion in closed basins. The goals of HB 831 are best achieved through a good understanding of the basic capacity or sustained yield of the watershed.

Spirited discussions regarding the effect of ground-water development on sustained yield are common in the literature. No author claims responsibility, but it has apparently become a working concept for some water policy makers that the sustained yield of an aquifer is equal to the recharge to the aquifer; one only needs to construct a water balance for a basin, establish the volume of recharge on an annual basis, and that value is the upper limit for ground-water development for that basin. Such thinking has drawn the ire of, and has been condemned as a myth by, most authors (Bredehoeft, 1997 is just one example). This concept is indeed an oversimplification of the water balance. Recharge equals discharge only under equilibrium; pumping from ground water disrupts that equilibrium and requires water from (a) induced recharge from the surface, (b) reduced discharge to the surface, or (c) both. There is general agreement that the rate of ground-water withdrawal need not exceed recharge to deplete a stream (Bredehoeft,1997; Sophocleous, 1997). There remain differences, however, as to the best way to determine sustained yield. Much of the discussion of late centers on the scale of observation/evaluation in determining the long-term yield of an aquifer. Bredehoeft and others (1982) argue that the water balance is irrelevant; capture, which is the reduction of discharge plus the induced recharge, need only be determined by long-standing methods of superposition. Superposition evaluates the changes caused by the new stress of pumping on pre-pumping conditions; aquifer properties and baseline water levels are the only data needed for these types of evaluations. The advantages of superposition evaluation have been defended in the literature; see, for example, Bredehoeft (1997), Bredehoeft (2002), Bredehoeft (2006), and Bredehoeft (2007). An alternate but not necessarily opposing view is that the evaluation should include as much of the hydrologic cycle as possible (Sophocleous,1997). For example, recharge is a fundamental component of the water balance; it should not be considered constant, nor should it be assumed that recharge is not affected (captured) by pumping (Sophocleous, 2004; Devlin and Sophocleous, 2005). It is well beyond the purpose of this report to sort out and resolve all these differences, mainly because both views are robust and applicable to evaluating depletion of surface-water discharge caused by ground-water development.

The Montana Water Use Act (Title 85, chapter 2, parts, 1–4, MCA) defines stream depletion as "the calculated volume, rate, timing and location of reductions to surface water resulting from a proposed groundwater appropriation that is not offset by the corresponding accretion to surface water that is not consumed and subsequently returns to the surface water".

In practice, the definition of stream depletion and its derivative, the stream depletion factor, varies with the author(s) of a given method. Moreover, benchmark methods such as those developed by Glover and Balmer (1954) and Jenkins (1968 and 1970) are sometimes used to calculate a stream depletion factor, but not always under constraints defined by the author (Miller and others, 2007). Although definitions vary with author and method, there is always agreement that response of the aquifer and the stream is non-linear and progressive (Glover and Blamer, 1954) as shown in figure 18; the non-linear nature of stream depletion is undisputed (Belleau, 1988). Similarly, the interaction between ground water and surface water is non-linear; that is, the rate of flow from the ground water to surface water is not the same as the rate of flow from the surface water to ground water under the same, but reversed, hydraulic gradient (Rushton and Tomlinsson, 2002). The shape of the curve in figure 18 depends on aquifer properties and the rate of ground-water withdrawal.

Methods of Determining Stream Depletion

Two categories of estimating stream depletion are commonly in use: analytical and numerical. Analytical methods employ a single governing equation derived under a specific set of assumptions. These methods include but are certainly not limited to those by Glover and Blamer (1954), Jenkins (1968 and 1970), Hunt (1999 and 2003), Collet and Zlotnik (2003), and Zlotnik (2004). Application of the analytical methods does not distinguish the two components of stream depletion: that ground water that is intercepted or captured by pumping and water withdrawn directly from the stream by ground-water pumping. Also, commonly used stream depletion models assume that all of the water withdrawn as ground water results in an equal amount of stream depletion. Other assumptions of these methods include:

- the stream "fully penetrates" the aquifer, not applicable in alluvial valleys where aquifers are hundreds or thousands of feet thick;
- the aquifer is homogeneous and isotropic; again, not applicable to valley-fill aquifers;
- the stream stage, stream discharge, and well discharge are constant; a changing stream flow due to runoff, natural depletion, are not considered; and

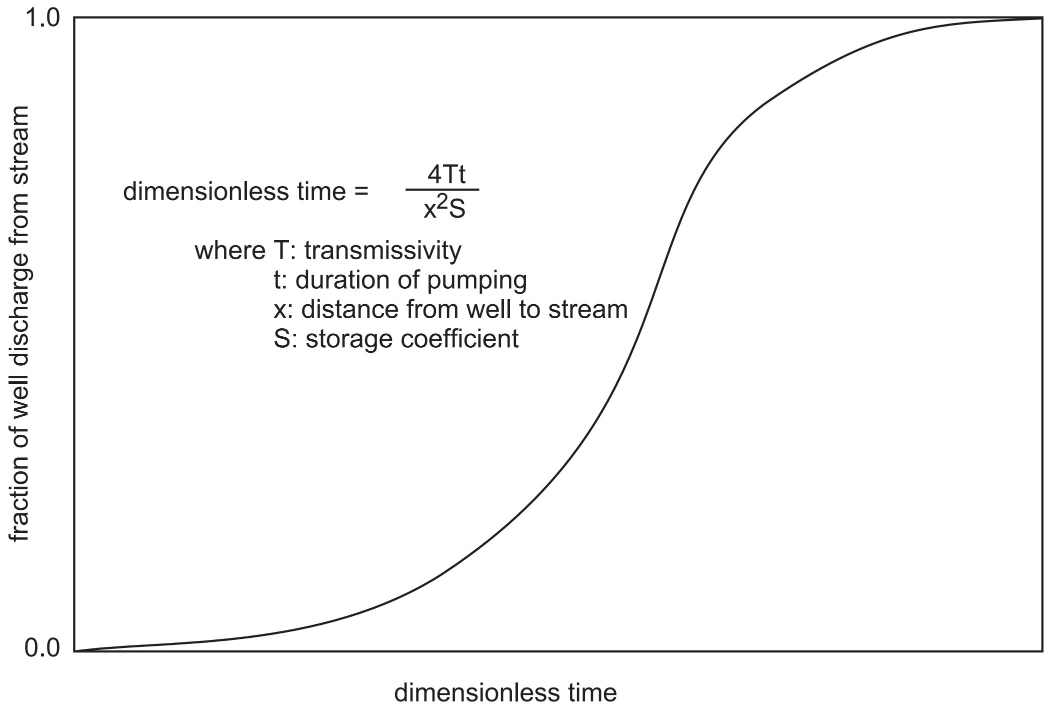


Figure 18. As pumping progresses the amount of water extracted from aquifer storage decreases and amount of water extracted from the surface increases, but not linearly (after Glover and Balmer, 1954).

- there is one well and one stream only; the cumulative effects of multiple wells with multiple discharge rates cannot be addressed.

Some argue that under heterogeneous hydrologic conditions, direct and immediate stream depletion can be much less than 100% of the volume pumped from the well (Zlotnik, 2004). There have been numerous attempts to eliminate the more restrictive assumptions such as a stream width or a partially penetrating stream (Zlotnik and others, 1999; Darama, 2004), a gaining stream (Di Matteo and Dragoni, 2005), and layered aquifers (Butler and others, 2001). There are also methods that directly evaluate aquifer-test data for estimating stream depletion (see Lough and Hunt, 2006) and testing the streambed hydraulic conductivity for use in stream depletion equations. It is also important to note that analytical methods for determining stream depletion only consider the impacts of the gross withdrawal of ground water and that the relationships between consumptive and non-consumptive uses cannot be evaluated directly. However, it has become practice for some entities to substitute consumptive use for well discharge (IDS, 2005). Using the net withdrawal (consumptive minus non-consumptive uses) results in a lower value for well discharge and a lower estimate of stream depletion as a way of demonstrating that stream depletion is equal to the water consumed. However, the timing and location of return flows for the water that is withdrawn but not consumed is not addressed. Consumptive use as a substitute for the actual volume withdrawn and returned is only valid in a sufficiently large area such as a watershed over a sufficiently long period of time such as many decades.

Numerical methods employ computer-generated approximations common to the hydrogeologic assessment of watersheds and aquifers. MODFLOW (McDonald and Harbaugh, 1988), BRANCH (Schaffranek, 1987), and HSPF (Bucknell and others, 1997) are just a few numerical models that have been used to estimate stream depletion. The STRMDEPL model (Zarriello, 2001) was developed as an interface for HSPF to directly estimate stream depletion. Most recently, Markstrom and others (2008) combined the Precipitation Runoff Modeling System (PRMS) (Leavesley and others, 1983) and MODFLOW codes in GSFLOW to facilitate basin-scale modeling of stream depletion. Many of the limitations of the analytical methods are overcome by the numerical models; however, the data requirements for a robust simulation are often beyond what is available for some watersheds. Many authors have evaluated various analytical methods and numerical simulations for calculating stream depletion at the field scale (for example, Sophocleous and others, 1995; Sphocleous, 2002; Miller and others, 2007); the numerical methods are generally favored. The discussion of superposition methods vs. evaluating the entire hydrologic cycle can also be carried out with numerical models. On the one hand, the impact of a well on stream discharge is modeled and superimposed on background water levels; only aquifer properties and boundary conditions are needed (Bredehoeft, 2004). On the other hand, a basin-scale model that includes all aspects of the water cycle can be conducted. The basin-scale model that incorporates the entire water cycle generally provides the most robust evaluations of stream depletion (Devlin and Sophocleous, 2005). A practical evaluation of stream depletion lies somewhere between the large basin-size model and the spreadsheet calculation for a single well. The compilation of data sufficient to completely evaluate stream depletion in any of the closed watersheds in Montana would take a decade or more, but a thorough evaluation for a sub-basin of that watershed may only take a few years.

EVALUATION OF STREAM DEPLETION IN THE LOWER BEAVERHEAD RIVER

MODFLOW (McDonald and Harbaugh,1988) was used to simulate ground-water and surface-water conditions in the lower Beaverhead River basin; Ground Water Vistas (version 5.0, Rumbaugh and Rumbaugh, 2007) was used for pre- and post-processing. The model area was bounded by Anderson Lane on the southwest, downstream of Beaverhead Rock (BR) on the northeast, and the East Bench Irrigation Canal near the head of Spring Creek and Stone Creek on the southeast (figure 19). The focus area of the model was the area between Anderson Lane and Beaverhead Rock. The objective of the model was to evaluate the effects of ground-water withdrawal on stream flow under the hydrologic conditions found in the Lower Beaverhead River sub-basin as well as other sub-basins in western Montana. The simulations herein are intended to evaluate changes in ground-water levels and stream discharge under pumping and depletion offset scenarios that show the greatest response and are not meant to reflect the actions of the landowners.

Numerical Model Dimensions

The model consisted of 127 rows, 153 columns, and 3 layers for a total of 58293 cells (16953 no flow, inactive cells); model cell-spacing was held uniform at 328 ft in both x and y directions. Aquifer parameters used in the model were based on those reported in permit applications and aquifer tests conducted as part of this study (table 6). The top layer included the near-stream alluvium and shallow bedrock aquifers; boundary conditions in the top layer include the river (stream package; 274 segments/cells), irrigation drains as applicable (drain package), basin margin recharge (general head boundary; 123 cells), and shallow test wells (well package). The elevation of the top layer was based on a 7.5-minute scale digital elevation model with 30-meter spacing. The second layer was intended to simulate the layer of clay, where present, as well as its interaction with the other hydrologic units. In the floodplain and partially into the floodplain margins, layer 2 was held at 30 ft thick; elsewhere layer 2 was varied, but always much thicker than 30 ft to accommodate the expected range of water levels. The presence or absence of clay in layer 2 was controlled by assigning the appropriate value of hydraulic conductivity to the model cell. No boundary conditions were applied in layer 2. Layer 3, the bottom layer of the model, was used to simulate the Tertiary-age sediments/bedrock and Mesozoic-age bedrock aquifers in the areas outside the floodplain, beneath the floodplain deposits, and in the Beaverhead Rock area. Boundary conditions in the bottom consisted only of production/test wells for the Tertiary aquifer (well package).

Data were generally good for overall hydraulic conductivity but limited for horizontal anisotropy and lacking for vertical hydraulic conductivity. Horizontal hydraulic conductivity was assumed to be isotropic (Kx = Ky) for the model. The lithology of the clay layer and Tertiary-age bedrock suggest a much lower vertical hydraulic conductivity (Kv) and were assumed to be 1% of the value used for the horizontal hydraulic conductivity. The vertical hydraulic conductivity of the alluvium was assumed to be 10% of the horizontal hydraulic conductivity. There are two types of simulations used in this and most modeling efforts: steady-state and transient. The results from steady-state simulations reflect the final conditions when no changes are occurring; for example, pumping wells have reached their ultimate, maximum drawdown. Steady-state simulations do not take time into account; for example, the time it takes to reach maximum drawdown in a pumping well is not known. Transient simulations do take time into account. For example, a well can be pumped for a specified number of days and the simulation can produce drawdown vs. time.

Numerical Model Calibration

Calibration most often involves comparing modeled head values to those observed in the field under similar conditions. Under steady-state conditions, adjustments were made to hydraulic conductivity, stream stage/discharge, and other parameters prior to simulating pumping, in order to achieve a good comparison between the model and hydrologic data presented in the earlier sections of this report. The steady-state simulation was also calibrated to maximum drawdown at test wells throughout the area in the near-stream alluvium and Tertiary aquifer. Calibration of a transient model requires longer term data. Stream flow, water-level records, and aquifer test data were used for a partial calibration; however, the final version of the model will require calibration with the detailed irrigation season data yet to be collected in

Table 6. Modeled hydrostratigraphic units are represented by three modeled layers.

Layer	Lithology	Horizontal hydraulic conductivity (ft/day)	Average thickness (ft)	Storage coefficient
1	alluvium	75	30	1E-01
	Tertiary-age bedrock	4	30+	1E-5 to 5E-4
	Mesozoic-age bedrock	1-2	30+	5E-4
2	clay (where present)	0.01	30	1E-06
	Tertiary-age bedrock	4	30+	1E-5 to 5E-4
3	Tertiary-age bedrock	4	200	5E-04

2008. The existing data available through cooperative land owners and their consultants proved invaluable for constructing this draft model. In addition to providing some level of certainty, if not calibration, the steady state simulation was used as the basis for drawdown calculations in subsequent transient simulations. Although there was good comparison between those few wells and the steady state simulation, lack of water level data for the majority of the model area precludes the calculation of calibration statistics for this draft. As will be discussed in the following section, calibration of the transient simulation also included a comparison between the model and an analytical solution for estimating stream depletion.

Steady-State Modeling

The steady-state simulation was used to establish a baseline from which to evaluate changes in water levels and stream flow as a result of changes in pumping or other stresses. The conditions in the baseline simulation should be considered the low end of the range of flow. That is, stream flow, as well as ground-water flow and gradients, are at their lowest so as to enhance the differences between the baseline simulation and simulations of various changes. The transient simulation was used to evaluate the change in stream flow (depletion/accretion) caused by various changes. Under real conditions, the amount of surface water and ground water flowing into the area varies considerably. The evaluation of changes, however, is best accomplished by holding surface-water and ground-water "recharge" constant.

Stream Depletion Analyses

As noted, stream depletion is the reduction in stream flow resulting from the withdrawal or capture of ground water by pumping and is usually expressed as a volume per time rate [e.g., gallons per minute (gpm) or cubic feet per second (cfs)]. In addition to rate, stream depletion must be evaluated with respect to location (the portion of the stream that will be depleted) and timing (when the depletion starts and ends). It follows that methods to offset stream depletion should be evaluated in the same manner. The analyses described herein considered the test well and the four irrigation wells in figure 19 as representing the four hydrologic conditions under consideration: near-stream shallow (test well 2B), near-stream deep (IR3), distal deep (IR1 and IR2), and basin margin deep (IR4). It is important to note that even the greatest stream depletion rate calculated falls within the error of measurement of stream discharge. It is commonly accepted that the error in measuring stream discharge is at least 10%; in a stream that is discharging 100 cfs, the error would be at least ±10 cfs.

Near-Stream, Shallow Conditions

The near-surface alluvium deposited by the Beaverhead River is rarely used for ground-water withdrawal; however, it is included in this analysis to serve as a basis for potential future development in this watershed or for present development in other watersheds under similar conditions. The model was run with a single well (test well on fig. 19) pumping 850 gpm or about 1.9 cfs. The first simulation was steady-state; that is, the well was pumped until maximum drawdown and stream depletion were reached. In figure 20, the red line and left axis indicate pre-pumping (baseline) stream discharge with distance downstream (each segment is about 328 ft); the green line indicates stream discharge under steady-state pumping conditions. As shown, the stream discharge is about 100 cfs at Anderson Lane and increases to about 103 cfs at the Beaverhead Rock (BR) area and about 104 cfs about 1 mile downstream. The gray shaded area indicates the maximum reduction of stream discharge or depletion in each segment; at steady-state, depletion affects the stream starting at about segment 30 and extends downstream The stream depletion is a maximum of about 1.8 cfs; in other words, nearly all of the water being pumped from the well is water that has been captured from discharging to the stream. This is what would be expected under these conditions. Steady-state simulations do not provide the timing of depletion, only the ultimate maximum rate and location of depletion; a transient simulation addresses the timing.

The results of a transient simulation of the same near-stream condition are presented in figure 21. As before, the red line indicates baseline; however, in this case, the green line indicates stream flow after 30 days of pumping, not steady-state. Also, the blue line indicates conditions after 150 days, or 120 days after pumping stopped. The gray portion of the graph indicates stream depletion for each segment just after 30 days of pumping. Note that the uppermost location of depletion, at segment 90, is downstream of that indicated in the steady-state simulation.

A comparison of the steady-state and transient simulations suggests that stream depletion starts and maximizes soon after pumping starts; both show a similar maximum depletion of about 1.8 cfs which is nearly equal to the well discharge of 1.9 cfs. The upstream extent of depletion is at about segment 100 in the transient simulation compared to about segment 75 in the steady-state simulation. In other words, the extent of stream depletion in the transient simulation has not yet reached the full extent although pumping has been stopped. This is a subtle demonstration that the effects of pumping continue to expand even after pumping is stopped.

Near-Stream, Deep Conditions

As discussed earlier, there is a considerable body of work comparing analytical solutions to numeric modeling results. As a means of calibrating the model directly to its intended use, such a comparison was made for this model. Schroeder (1987) provides a convenient means of applying the Jenkins (1970) method of estimating stream depletion with a single well pumping

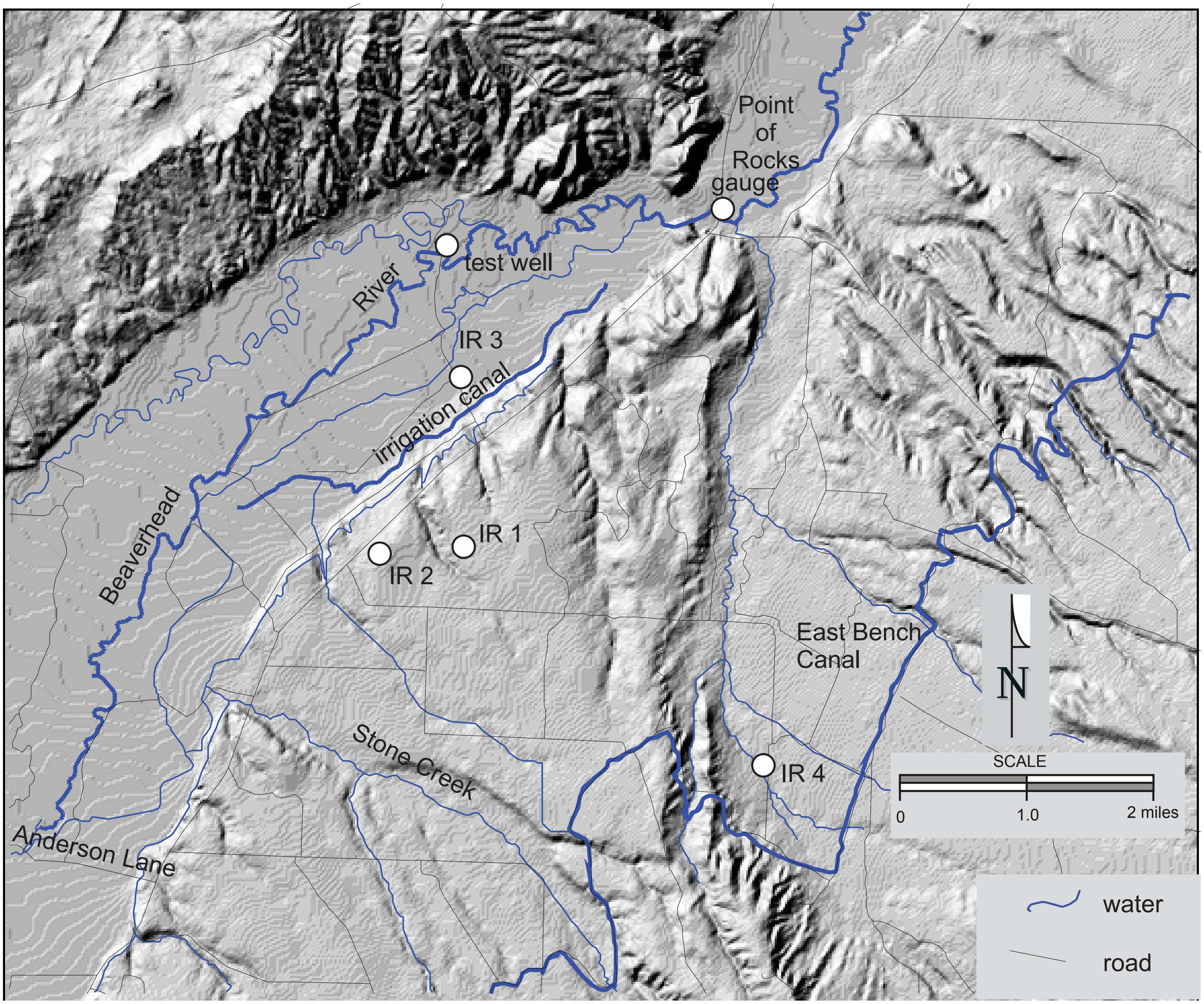


Figure 19. The grid for the ground-water flow model included the entire area shown. The focus was the area between Anderson Lane and Point of Rocks.

at a known rate, at a known distance from a stream given the transmissivity and storage coefficient of the aquifer. Given that the irrigation wells of the lower Beaverhead River nearly all utilize the deeper Tertiary aquifer, data from a deeper well at a distance most likely to show significant depletion was used for the comparison. Well IR3 (fig. 19) was "completed" in the Tertiary sediments simulated here as layer 3 of the model at a distance of 1,800 ft from the river. In the simulation, the well was pumped at 850 gpm for 30 days using the same aquifer parameters for both the numerical simulation and the analytical solution.

Figure 22 presents the results of the numeric simulation ("model"); again the stream discharge with distance downstream is presented for baseline (red), end of pumping (green), and recovery (blue), which in this case is about 150 days (120 days after pumping ended). The maximum stream depletion is about 0.32 cfs. Figure 23 shows the results from the Schroeder analysis; the maximum stream depletion using this method is about 0.33 cfs. The two methods show very good agreement.

Distal, Deep Conditions

Wells IR1 and IR2 (fig. 19) are completed in the deep Tertiary aquifer approximately 1.5 miles from the Beaverhead River. Neither alluvial material nor clay associated with the floodplain was reported in the driller=s log. Both wells are outside the influence of recharge from shallow aquifers and are good representatives for a deep aquifer distal from the stream. The model was run as before with only IR1 or IR2 pumping for 30 days followed by 120 days of no pumping. Figures 24 and 25 present the results for simulating wells IR1 and IR2 individually. As expected, the maximum rate of depletion for the period of simulation, about 0.15 cfs, is nearly equal between the two wells and less than that of a well closer to the river (e.g., IR3).

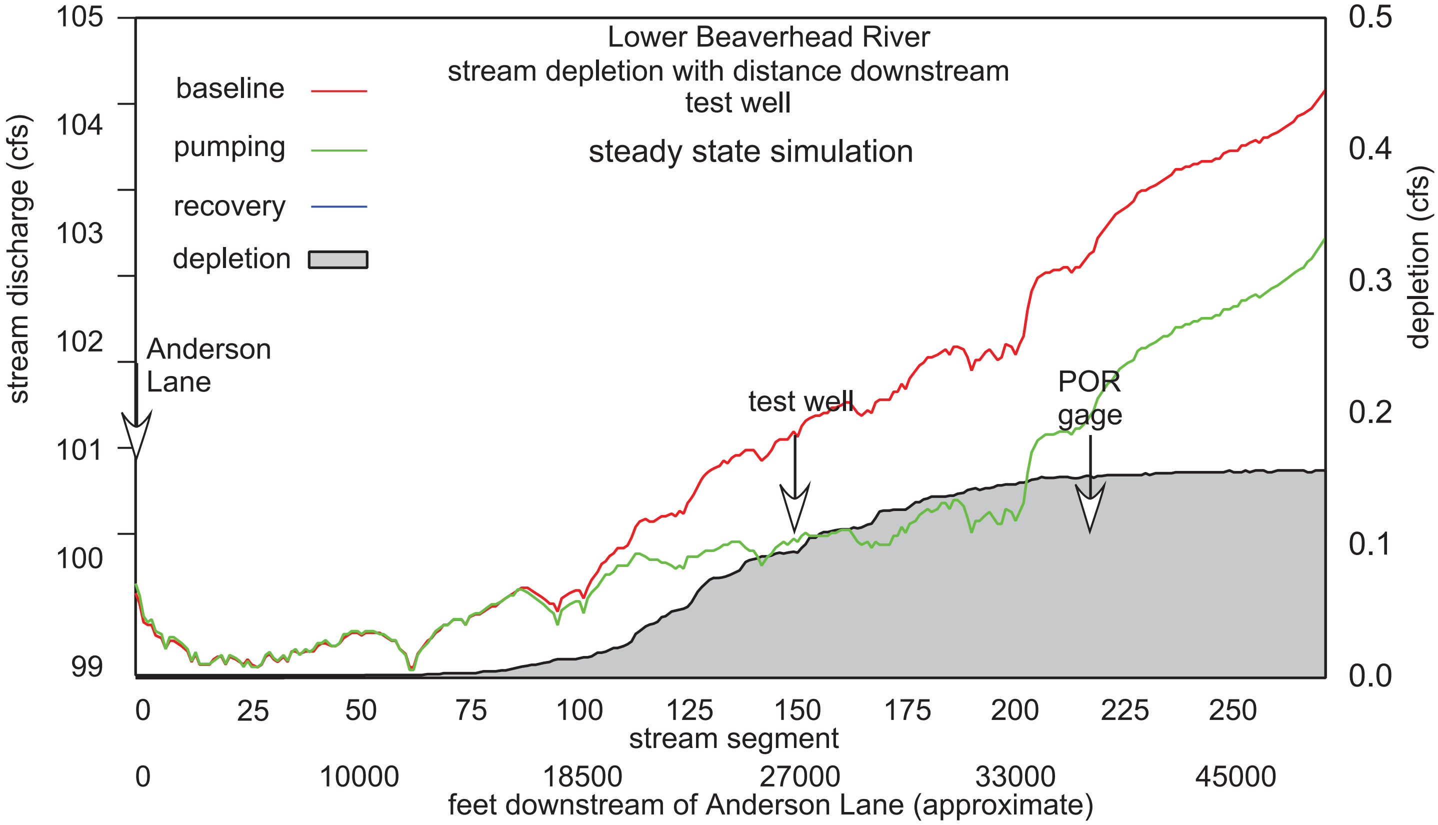


Figure 20. The steady-state simulation of pumping the alluvium near the river at segment 150 shows that depletion affects the river upstream as far as segment 50 and has a maximum depletion of about 1.8 cfs (gray area and right axis).

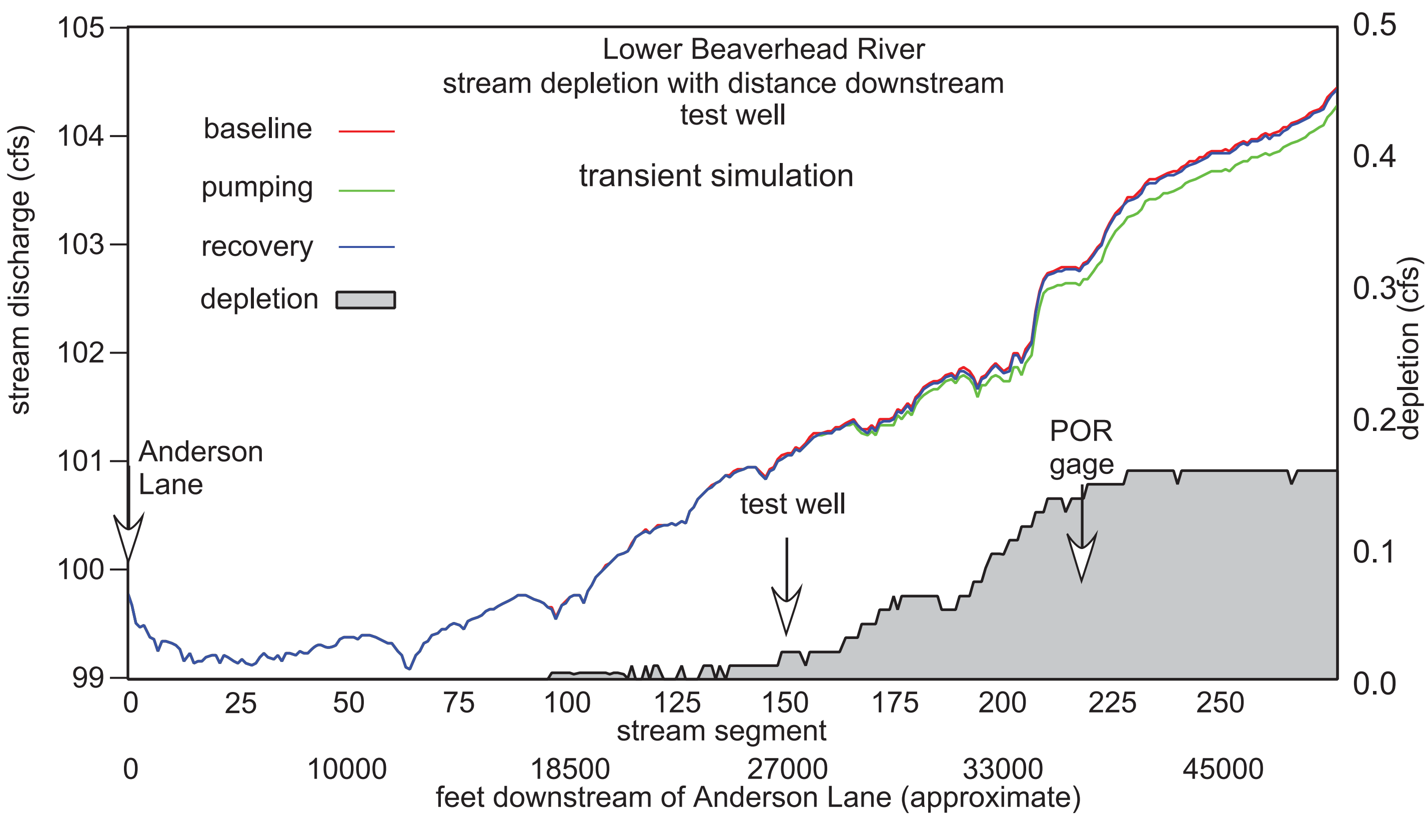


Figure 21. The transient-state simulation of pumping the alluvium near the river at segment 150 shows depletion in the river after 30 days of pumping at 850 gpm (1.9 cfs). The maximum depletion is about 1.8 cfs (gray area and right axis).

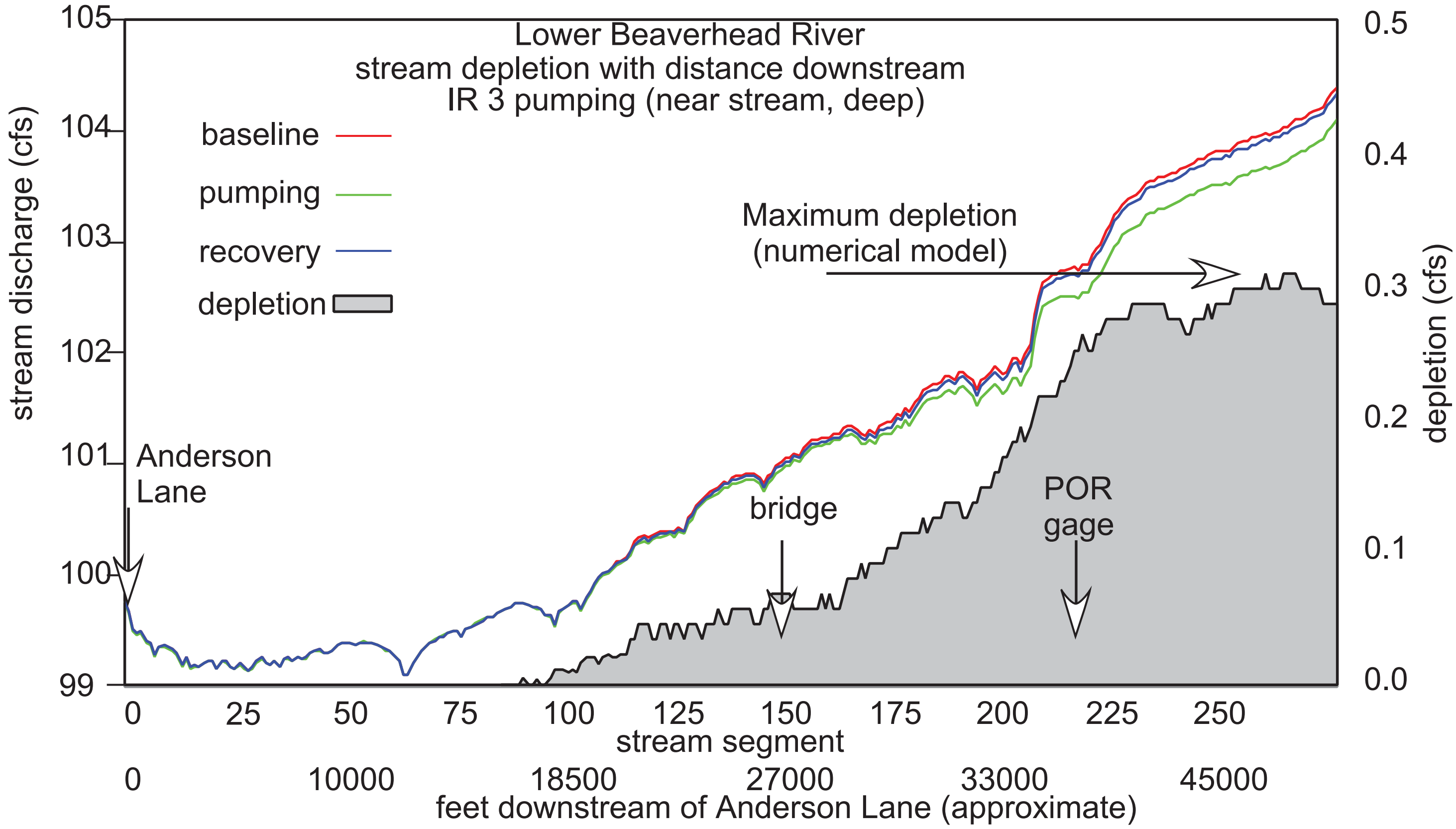


Figure 22. Well IR3 was pumped for 30 days at a rate of 850 gpm in the numerical model simulation; this produced a maximum depletion rate of about 0.32 cfs (gray area and right axis).

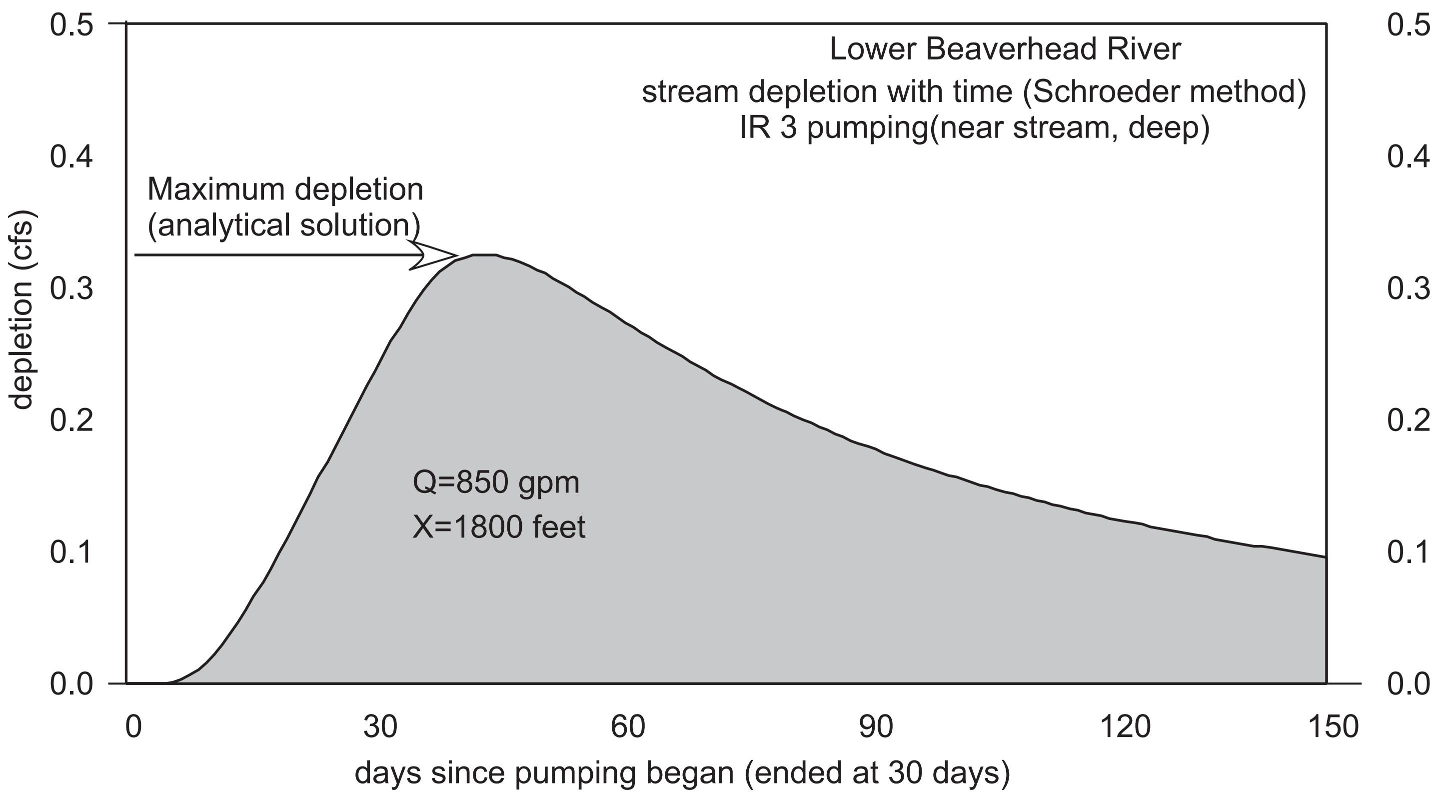


Figure 23. A well pumped for 30 days at a rate of 850 gpm at a distance of 1800 ft was used to solve for a depletion rate of about 0.35 cfs. Compare this to the modeling results in figure 22.

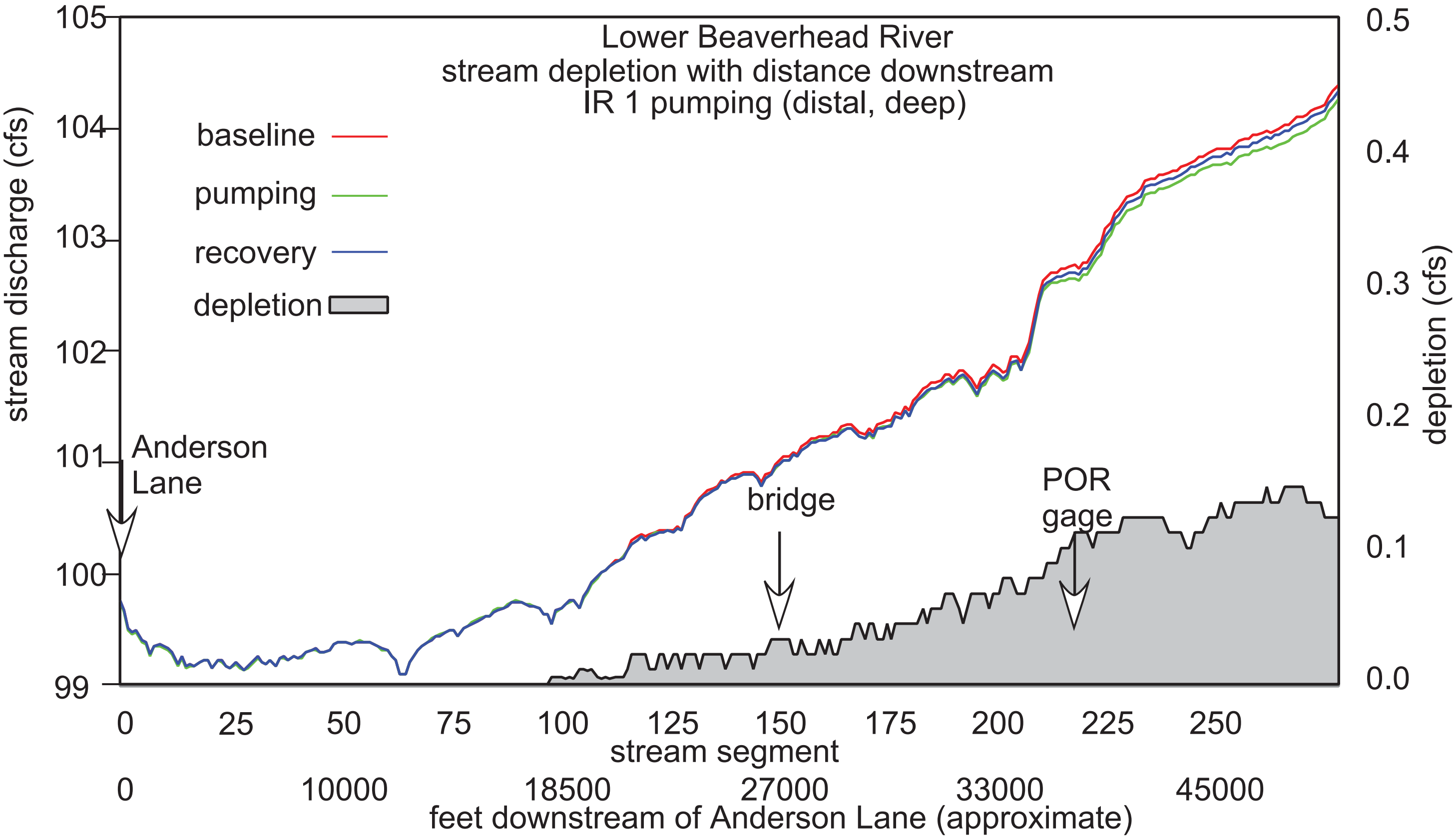


Figure 24. Well IR1 was pumped for 30 days at 850 gpm to produce a maximum depletion of about 0.15 cfs (gray area and right axis).

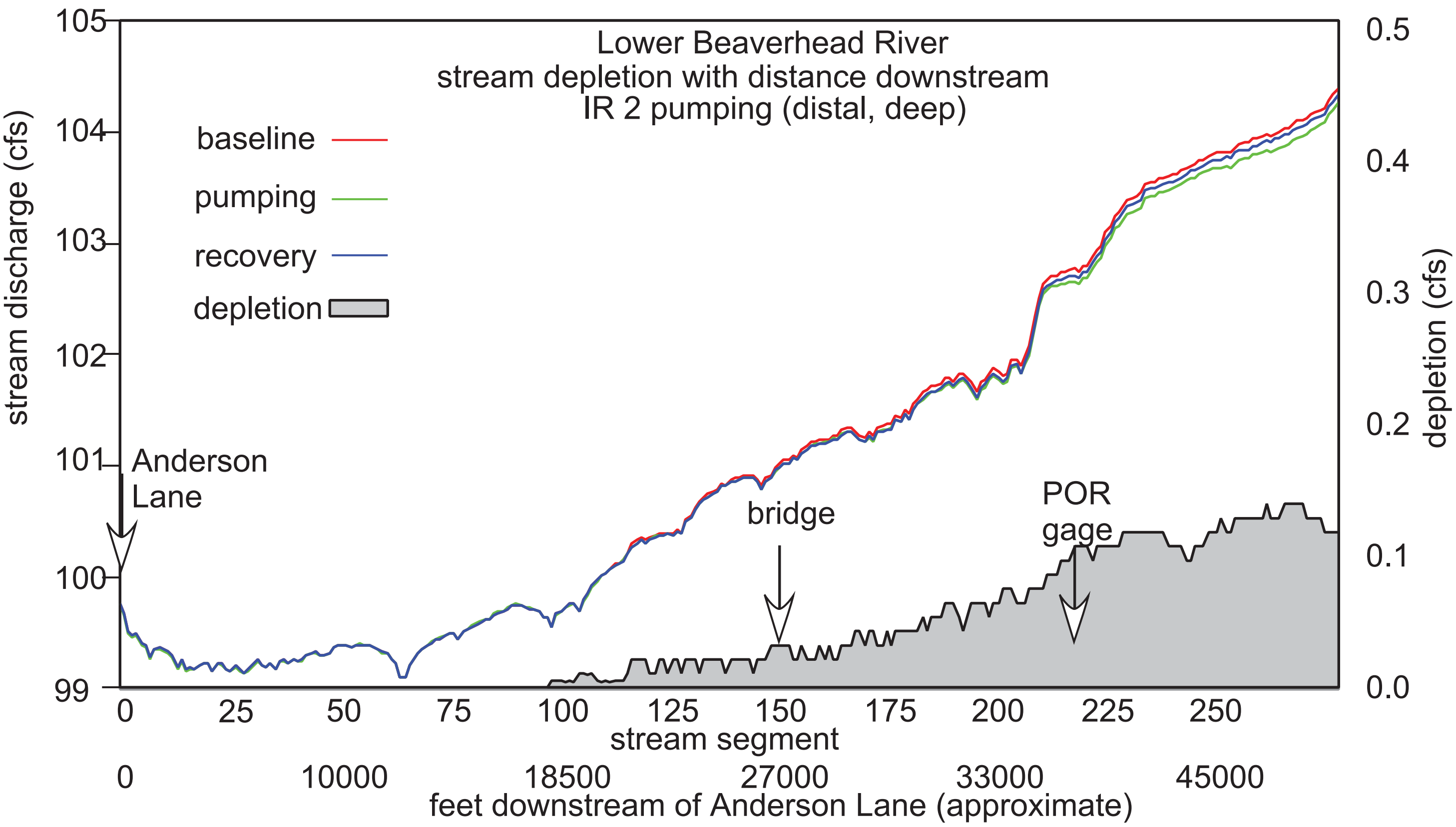


Figure 25. Well IR2, at about the same distance from the river as IR1, was pumped for 30 days at 850 gpm and produced essentially the same results as those for IR1.

Basin Margin, Deep Conditions

Well IR4 (fig. 19) is about 20,000 ft from the river; the well is 500 ft deep and completed in the same Tertiary-age aquifer as the other IR wells. The model was run with only IR4 pumping; the maximum stream depletion for the period of the simulation was about 0.13 cfs (fig. 26).

These simulations of individual wells demonstrate that the stream depletion decreased with distance from the stream. As will be discussed later, the time to reach the maximum stream depletion can be considerable and quite variable. Overlapping cones of depression and position with respect to the stream can greatly influence what reach of stream is affected as well as the rate of depletion

Cumulative Effects

The results of the model compare well with those obtained using the analytical method when a single well is considered. The analytical method, of course, is restricted to a single well in a single aquifer under simple boundary conditions. The numerical model, however, can consider multiple wells as well as multiple cycles of pumping. Figure 27 presents the results of all four irrigation wells (IR1, 2, 3, and 4), each pumping 850 gpm (about 1.9 cfs) for 30 days; the maximum stream depletion rate for the period of simulation is about 0.45 cfs. The results of the simulation of cumulative pumping indicate a much smaller rate (0.45 cfs) than the additive rate (0.76 cfs) of each well (0.35, 0.14, 0.14, and 0.13 cfs, respectively) that one would expect. The explanation for the difference lies in the timing and location of the depletion. Each well contributes to the depletion, but the distance from the stream determines when its maximum rate will be achieved. It is important, then, that any simulation be of sufficient length in time to evaluate cumulative effects of all wells, particularly those at greater distances from the stream.

Transient Simulation of Cyclic Pumping

The foregoing analyses used a single pumping and recovery cycle; in these cases, full recovery of the depletion is inevitable. In practice, however, recovery may not be complete before another cycle of pumping and depletion begins; a review of previous figures shows that the recovery line (blue) never quite returns to the baseline (red). In another simulation, all four of the wells were pumped at a rate of 850 gpm for 90 days. The system was allowed to recover for the remainder of the year, 275 days, and the cycle repeated for 3 more years (table 7). A drawdown curve for well IR3 typifies the response in each well (fig. 28). The first 120 days of all the simulations have no activity of any kind, to demonstrate that no background changes are occurring within the model prior to adding new activities.

As with the individual wells and the single pumping cycle, stream discharge decreased as pumping continued and recovered after pumping ended. The recovery, however, was not complete. A plot of stream discharge vs. time (fig. 29) reveals the trend of decreasing stream discharge resulting from a repeating cycle of pumping. It should be noted, however, that the trend is not linear. The cumulative effect of ground-water withdrawal is not simply a matter of adding up depletions caused by each well for each year. Stream depletion will reach a maximum value at some point in time and will be some fraction of the rate pumped from the well. The theoretical maximum fraction of depletion by a well is, of course, equal to 1.0; the stream depletion is equal to the well discharge. However, the time it takes to reach the theoretical maximum must be considered.

Returning again to the simple case of a single well in a single aquifer, the analytical method by Schroeder (1987) provides a convenient estimate of the fraction of the well discharge (equal to the ratio of the stream depletion rate to the well discharge rate).

Table 7. Stress period set up for transient simulations.

Stress Period	Period Length (days)	Cumulative Time (days)	Time steps	Activity
1	120	120	60	none, used to continue steady-state conditions
2	90	210	30	first pumping cycle
3	275	485	75	no pumping
4	90	575	30	second pumping cycle
5	275	850	75	no pumping
6	90	940	30	third pumping cycle
7	275	1215	75	no pumping
8	90	1305	30	fourth pumping cycle
9	275	1580	75	no pumping

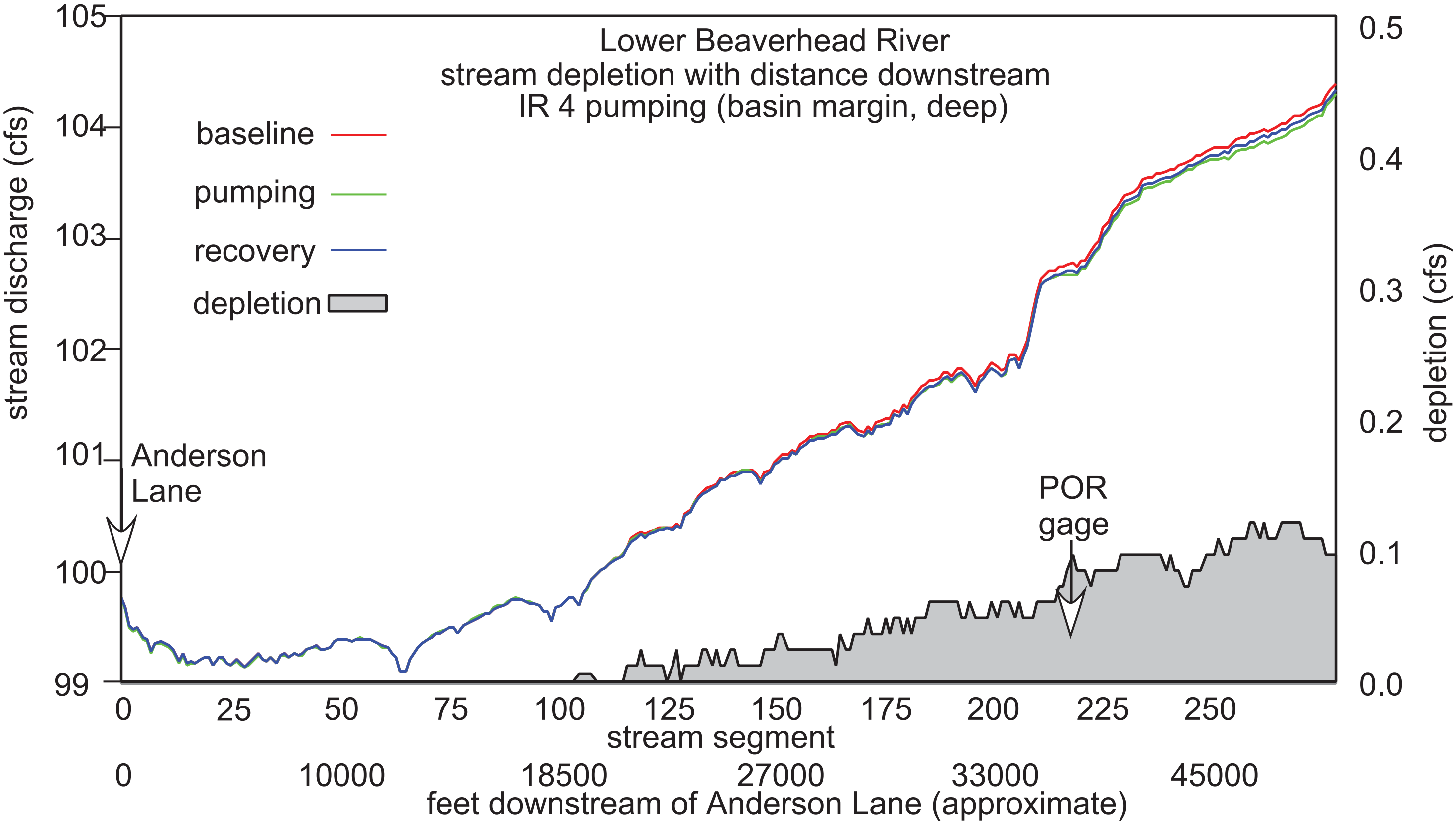


Figure 26. Well IR4, about 20,000 ft from the river, was pumped for 30 days at 850 gpm to produce a maximum depletion of about 0.13 cfs (gray area and right axis).

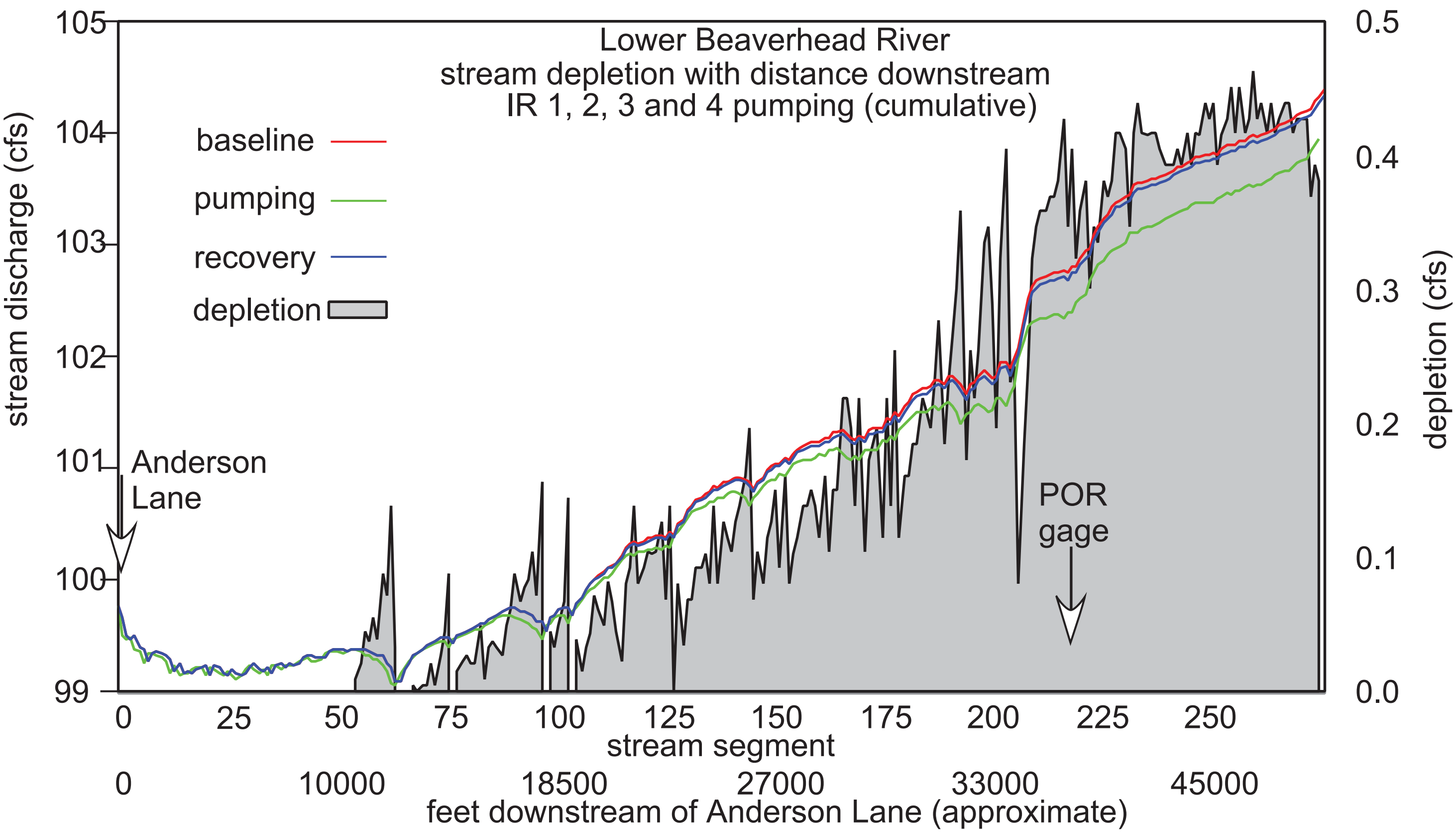


Figure 27. All four wells were pumped at 850 gpm each for 30 days. The cumulative maximum depletion was about 0.45 cfs.

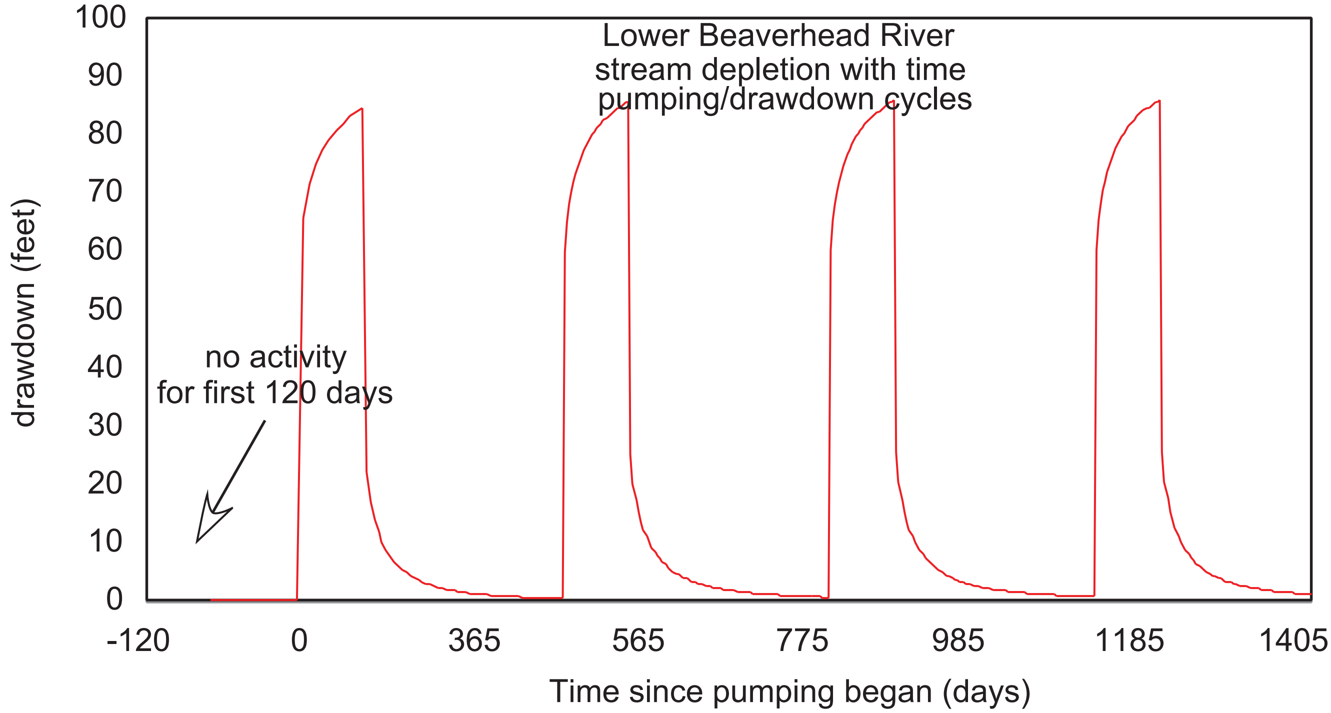


Figure 28. Each of the four wells were pumped at 850 gpm for 90 days and allowed to recover for 275 days. The cycle was repeated four times. Drawdown varied for each well, but was typically about 90 ft.

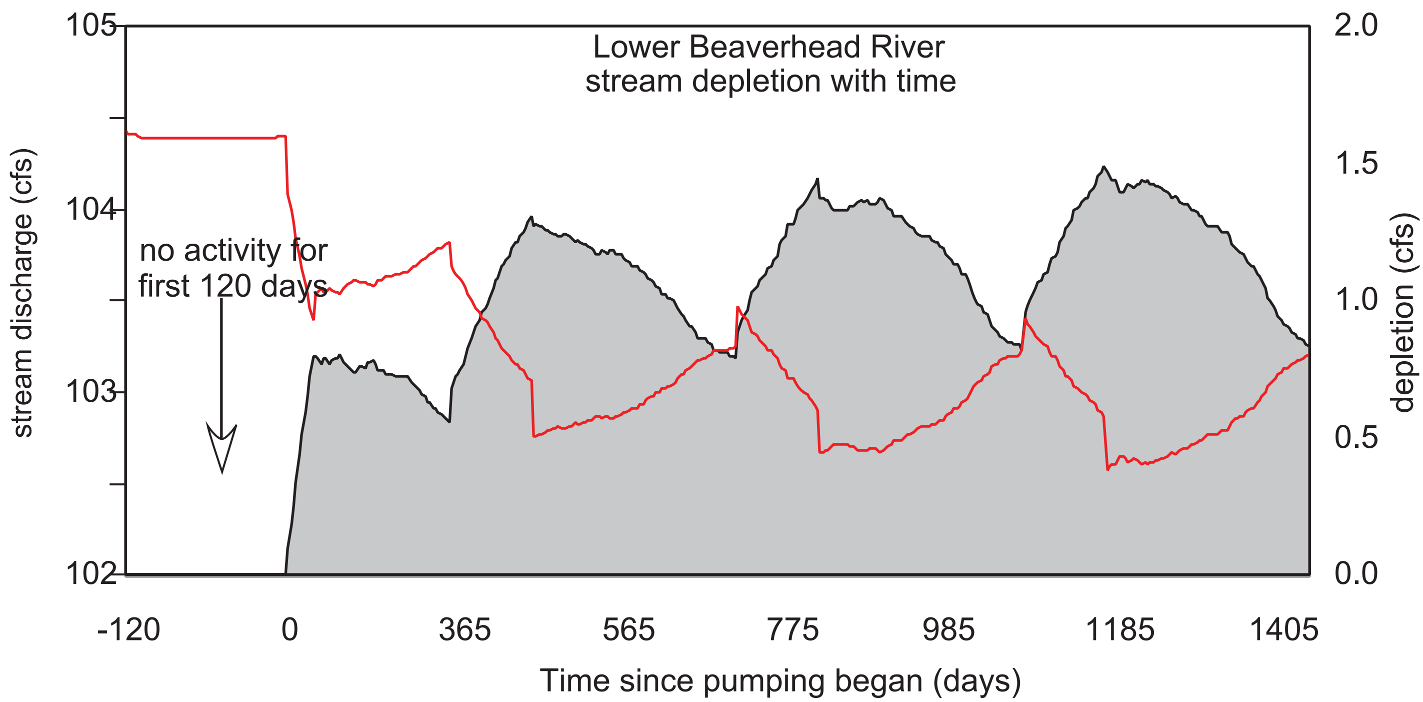


Figure 29. Stream discharge (red line) declined and recovered each year, but only partially; the long-term trend was downward.Similarly, the stream depletion (gray area and right axis) increased. Note that neither trend is linear.

Figure 30 presents the stream depletion analysis (using the Schroeder method) discussed earlier with the addition of the estimate of pump discharge fraction. For the case of a single well pumping 850 gpm for 30 days at a distance of 1800 ft from the stream, the fraction of well discharge reaches a maximum of about 70 percent. Note that the maximum occurs sometime after pumping stopped at day 30.

The cumulative effects of four wells pumping 850 gpm for 90 days over a period of about 4 years is presented in figure 31; these results are from the numeric model. The red line shows the increasing rate of stream depletion; the gray open circles and trend lines show the average and maximum fraction of well discharge. At the end of four cycles, the stream depletion rate is about 1.5 cfs and approaching 2 cfs, and the maximum fraction of the well discharge approaches 70 percent. In a similar manner, the contribution of each well with respect to depletion and discharge fraction can be generated. Although the analytical and numeric methods tend to agree, both analyses indicate that the stream depletion during the period of simulation fall short of the expected maximum of 2 cfs.

Analysis of Methods to Offset Stream Depletion

Effective mitigation of stream depletion requires knowledge of when, where, how much, and how fast depletion will occur. The preceding analyses have demonstrated various means to estimate the timing, location, and volume of depletion, but also demonstrated the needed to determine the rate of depletion with respect to time. Using the Schroeder method in the simple example, if a single well is pumped at 850 gpm without cycling, the stream depletion will be 83% of the pump discharge in 5 years, 92% in 20 years, and 96% in 100 years. Cycling at 90 days per year, the Schroeder method estimates a maximum depletion of about 1 cfs (460 gpm) or a fraction of about 40% of the pump discharge after 100 years (fig. 32). To summarize, stream depletion does not have a linear relationship with ground-water withdrawal; the rate of change in depletion is much higher in the early period of pumping.

Methods to offset stream depletion from pumping near the Beaverhead River were simulated under two general conditions: replenishment of stream flow and replenishment of ground water. From a mass balance point of view, the maximum stream depletion of four wells pumping 850 gpm (about 2 cfs each) for a quarter of the year is a total of about 2 cfs (fig. 31). The depletion then could logically be offset by the addition of 2 cfs to the stream of the same 90-day duration as the pumping. Figure 33 presents the results of adding 2 cfs to the river upstream of the depletion area during the same 90-day period as the pumping. This would be the effect of diverting water from some other source to the river or releasing stored water from a reservoir. The red stream hydrograph is the stream discharge with time without offset and the blue line is stream discharge with offset. The addition of the 2 cfs is easily recognizable by the spikes in stream discharge. Also notable, however, is that the net effect on stream discharge is inconsequential. Only when water is added to the stream does discharge reach its original value (dashed line). Immediately after the addition of the 2 cfs is stopped, stream discharge returns to near baseline/pre-offset discharge. Likewise, stream depletion (gray area graph) shows no improvement and the trend continues upward between applications of the 2 cfs. Overall, stream discharge continues its trend downward and stream depletion is not mitigated.

Figure 34 presents the results of the alternative option of replenishing ground water. In this case, the 2 cfs was added as leakage from an existing canal on the margin of the floodplain near well IR3 (fig. 19). In order to precisely control the rate of recharge for comparison with the previous simulation, 47 wells were used to inject the 2 cfs along the canal; in "real life" the canal would simply be activated, wells would not be necessary. As before, the red lines show the baseline, pre-offset discharge with a downward trend reflecting more stream depletion as the pump cycle continues. The blue lines are the result of diverting 2 cfs from the stream and using it to replenish ground water. Rather than spikes of increased flow when water is added to the stream, the blue hydrograph shows the temporary drop in stream flow due to the diversion. With regard to stream depletion, the blue hydrograph shows there is much more improvement in stream discharge compared to the surface-water approach, but it is nowhere near the original and the trend is still downward. For the short term, stream depletion has been offset (gray line graph and right axis) but not eliminated; as reflected in the stream discharge graph, the net depletion continues to increase.

The modeling shows that the goal of offsetting stream depletion has not been completely achieved by the ground-water approach, but these simulations serve to demonstrate the difference between replenishing surface water and replenishing ground water as methods to offset stream depletion. The difference in response reflects the difference in residence time or storage time. In the model and in field conditions, the 2 cfs added to the stream will run off within days; the residence time or storage time is relatively short. Conversely, ground water has a residence time of weeks to months in shallow near-stream aquifers. Thus, a robust evaluation of stream depletion and methods to offset that depletion will take these differences into account.

Stream depletion offset based on the volumetric accounting of the stream depletion provides an alternative strategy. Simply put, the volume of water discharged from four wells pumping 850 gpm each is about 59 million cubic ft. If that is to be mitigated over the same period of time as the pumping, the offset value is about 7.5 cfs in 90 days, not 2 cfs as determined from the Schroeder method. Even when using a well discharge of 3400 gpm (850 gpm x 4 wells), the depletion rate at 100

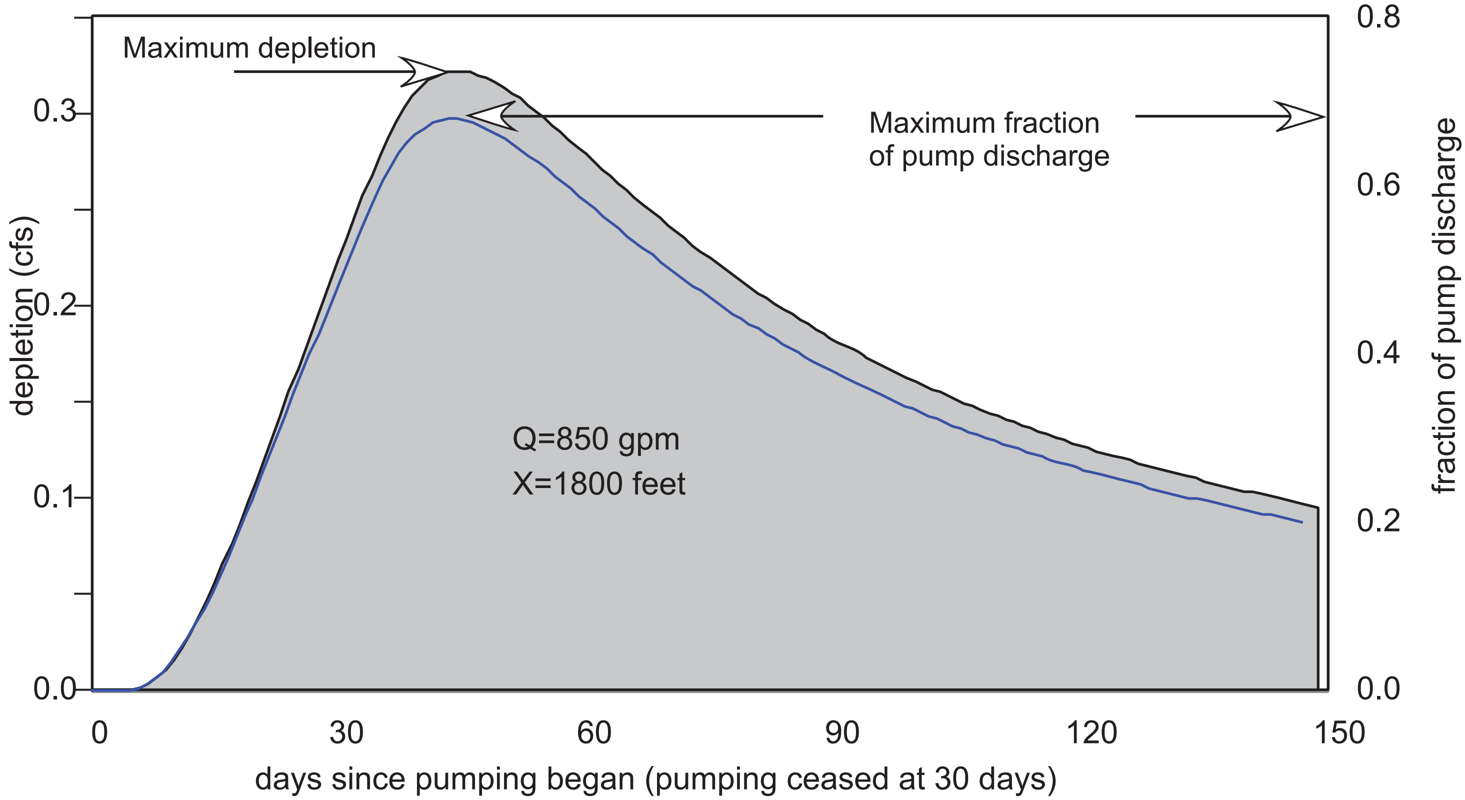


Figure 30. The analytical method by Schroeder (1987) produces a value of about 70 percent for the fraction of pump discharge originating as surface water.

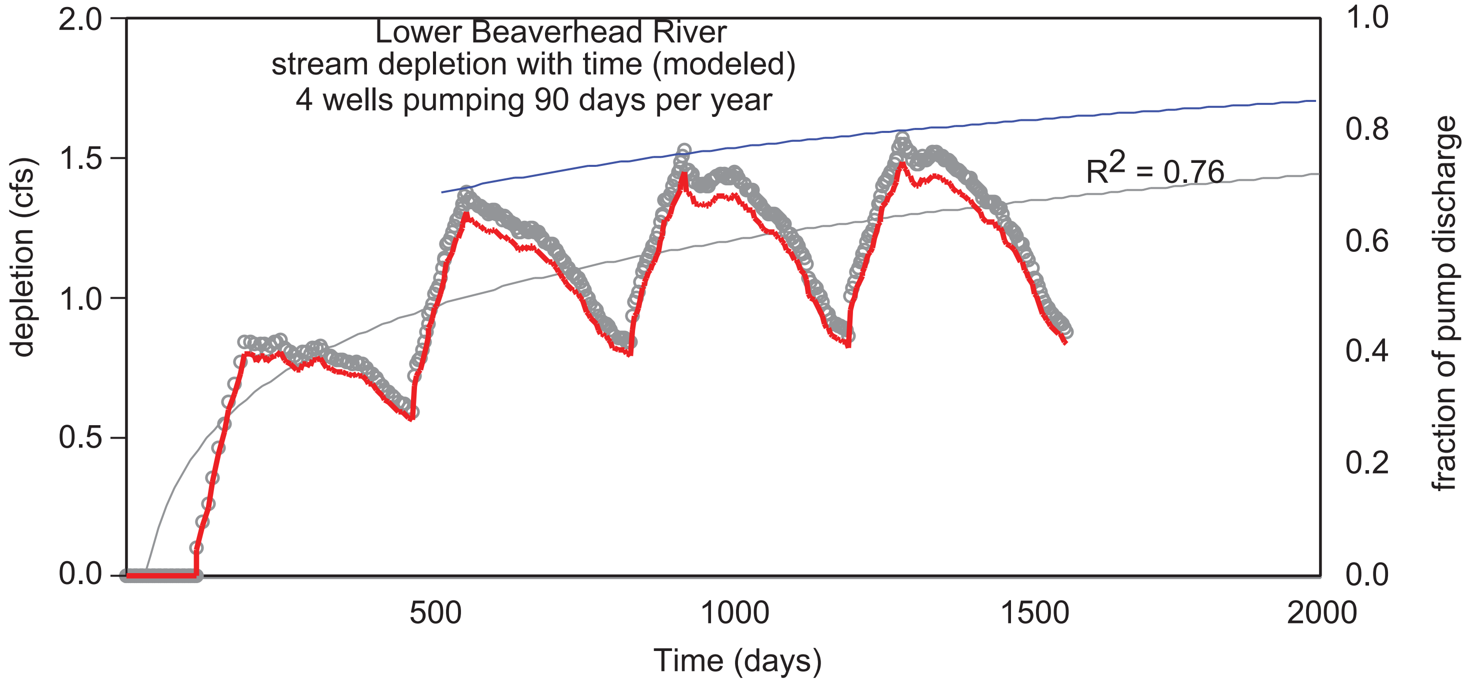


Figure 31. As stream depletion increases (red line), the fraction of pump discharge originating as surface water (gray dots) also increases. The trend continues upward as indicated by a best-fit line of the average fraction (gray) and maximum fraction (blue).

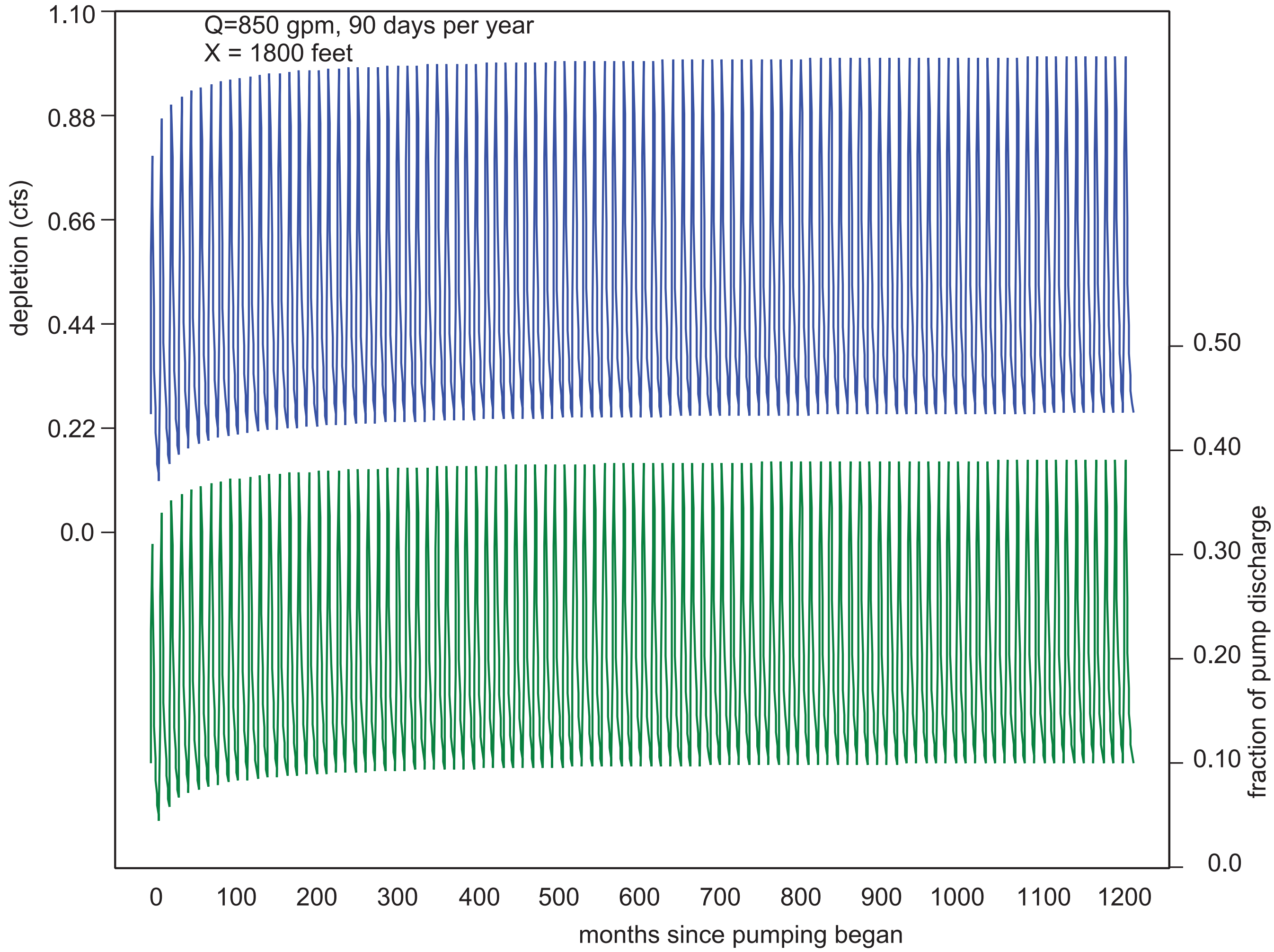


Figure 32. The Schroeder method estimated a maximum stream depletion of about 1 cfs (blue and left axis) and a maximum fraction of about 40 percent (green and right axis) in 100 years of cyclic pumping. Aquifer parameters reflect those found in the deeper aquifer of lower Beaverhead River.

years by the Schroeder method is 1500 gpm or about 3 cfs.

Figure 35 presents the results of using 8 cfs diverted from the river as recharge via the irrigation canal. As before, the red hydrograph is stream discharge with the four wells pumping and no offset. The blue hydrograph shows the rapid 8-cfs drop in stream discharge, but also reflects the effects of recharge from the irrigation canal. In fact, stream discharge exceeds the baseline (dashed line) by almost 1 cfs. The depletion graph (gray area graph) shows true accretion (below the line) as a result. It also shows a slight increase in accretion with time, indicating that stream discharge will not only be greater than the original but will increase.

Summary of Depletion Analyses

These analyses provide the background for meeting the objective of finding solutions specific to the lower Beaverhead River. Certainly, more simulations could be run for other conditions, including some that are in practice in other areas of the basin. For example, offset with the irrigation canal (but no flooding) over a longer period of time each year or at variable rates that take advantage of high stream discharge during spring runoff. Conveyance loss from irrigation canals can be considerable and may be a good component of stream depletion offset if the timing and location are favorable. If spring runoff waters could be diverted, the strategy becomes a form of aquifer storage recovery (ASR) for an unconfined system; however, ice in the rivers and irrigation canals and frost probably limits this approach in the Beaverhead River basin.

A major water consumer that is currently not being considered in the water management of the closed basins is [management of?] woody phreatophyte propagation (salt cedar, cottonwood, willow, etc.). Phreatophytes are defined as a type of plant that has a high rate of transpiration by virtue of a taproot extending to the water table (Fetter, 1988). These phreatophytes typically occur as riparian vegetation along waterways. Programs are in place or have been proposed to control or eliminate non-native species, such as salt cedar. On the other hand, native woody phreatophytes are generally promoted by State and Federal agencies, because these plants provide important habitat for a number of animal species. However, the natural controls on the

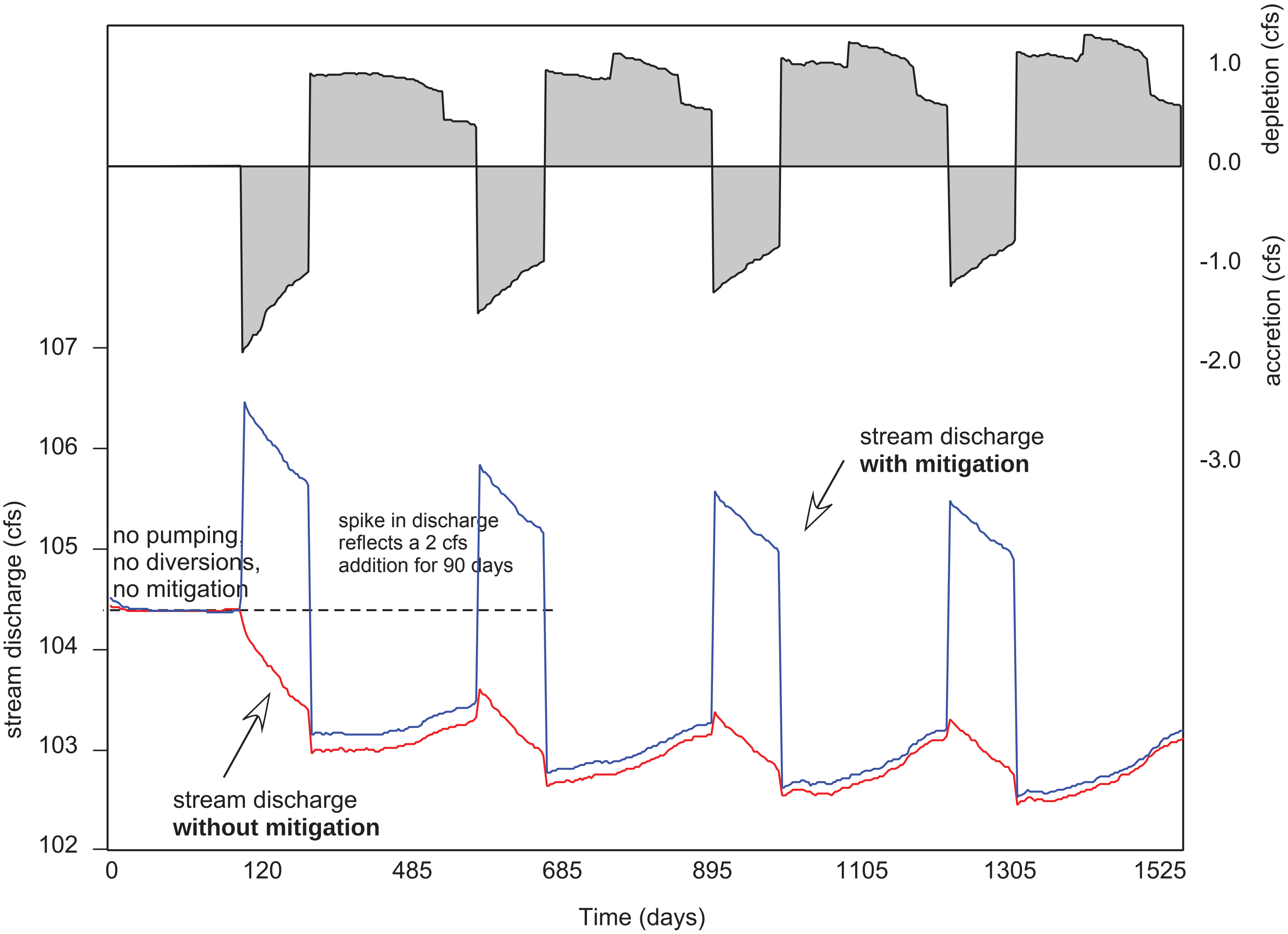


Figure 33. Stream discharge without offset (red line and left axis) shows a declining trend. Stream discharge with offset roughly equal to the predicted stream depletion applied to surface water shows a very similar downward trend (blue line and left axis) and discharge remains well below the original (dashed line). Similarly, a comparison of stream discharge before and after the attempted offset (gray area graph and right axis) shows no improvement.

propagation of these phreatophytes have effectively been removed from most riparian systems in Montana.

Prior to the settlement of Montana by Europeans, beaver, fire, and large ungulates (bison, moose, and elk) naturally controlled the abundance of these plant species. Settlement of the bottom land along the rivers and streams has effectively eliminated the natural controls on phreatophyte propagation. In addition, recent conservation measures promoting riparian vegetation has greatly limited man-made controls on phreatophyte propagation in many areas. The removal of natural and man-made controls on phreatophyte propagation has led to unnatural abundances in many areas compared with pre-settlement densities (Lesica and Cooper, 1997). For example, when William Clark came down the Yellowstone River in 1806 he could not find any trees large enough to build canoes between present-day Livingston and Park City, Montana. At Park City Clark's party lashed two cottonwood trees together to make canoes 16 to 24 inches wide (MacDonald, 1950). William Clark would have little difficulty finding large cottonwoods along the Yellowstone today.

The abundance of woody phreatophytes would not be an issue if it were not for their intensive water consumption. Hackett and others (1960) reported a 2 acre-ft/acre annual water consumption rate for a cottonwood grove in the Gallatin Valley, which is likely to be a typical consumption rate for phreatophytes in the closed basins of southwestern Montana. This consumption rate is comparable to the annual application rate for sprinkler-irrigated alfalfa in southwestern Montana. However, due to their proximity to streams, these plants have a direct negative impact on in-stream flows during the summer months when flows are lowest, and unlike irrigation this consumption cannot be limited or controlled once in-stream flows reach a critical point. Because these are largely artificial ecosystems, it is important to balance the habitat enhancement derived from phreatophyte propagation with the water consumption. Active management of phreatophyte abundance is needed to avoid negatively impacting the availability of water resources in the closed basins.

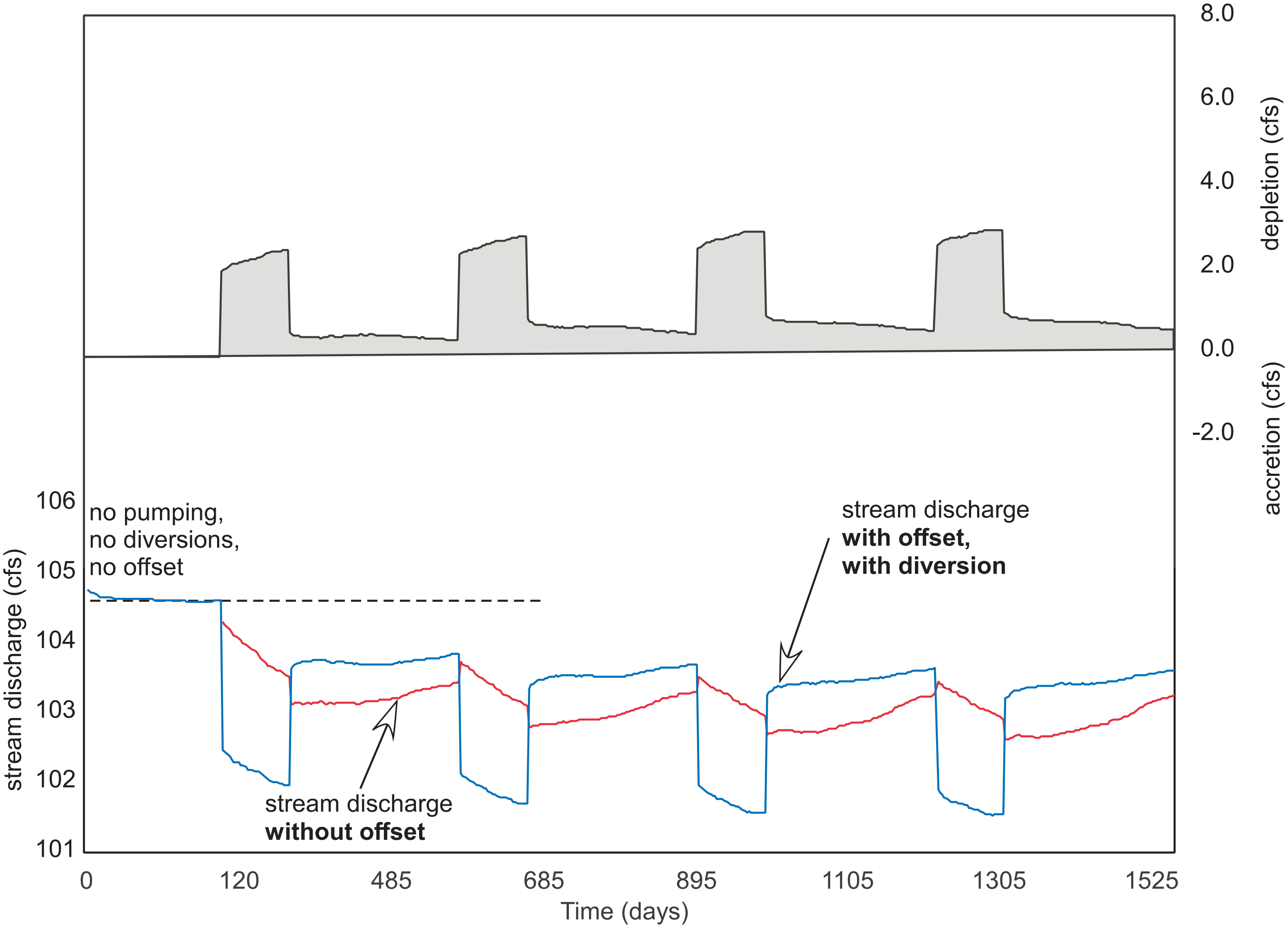


Figure 34. Stream discharge without offset (red line and left axis) shows a declining trend. Stream discharge with offset roughly equal to the 2 cfs predicted stream depletion, but applied to ground water, shows less of a downward trend (blue line and left axis). Stream depletion relative to pumping without offset approaches zero (gray area graph and right axis) and seems to be having a lasting effect, but the stream discharge is still less than the original discharge (dashed horizontal line).

The effects of evapotranspiration (ET) from crops and phreatophytes were not considered in these evaluations, mainly to keep the differences between the various methods simple and clear. As noted, some have suggested substituting ET or even lumping all consumptive use for the well discharge value. In the analytical methods such as the Schroeder methods used here, the result is a much lower depletion rate. The same is true for the numeric modeling, but the limited value of the application becomes clearer. One does not pump less because one will not be consuming all the water pumped. When and where the unconsumed water is returned to the system is every bit as important as the method(s) of offsetting stream depletion; it should not be treated as an artifact of a budget calculation. The simple example is perhaps the most common condition: pumping from a deep aquifer and recharging a shallow aquifer with the excess water. The deeper aquifer is not likely to be recharged by the shallow aquifer before the excess water is discharged to surface water. These analyses, along with exhaustive research presented in the literature, demonstrate that the timing, the rate, and the location of depletion are affected directly by the timing, location, and rate of pumping. The same is true for offsetting that depletion.

In the summer of 2008, additional surface-water and ground-water data will be collected during the irrigation season that will be used to “nail down” the relationships among pumping ground water, irrigation canal leakage (recharge), and stream discharge. The new data will also allow the model area to be expanded to include the area west and southwest.

The methods of analyzing stream depletion and offsetting stream depletion for the lower Beaverhead River, while pertinent for local conditions, are easily applied in other parts of this and other watersheds. The area of investigation is of sufficient extent to develop useful models, but not so large as to take many years of study.

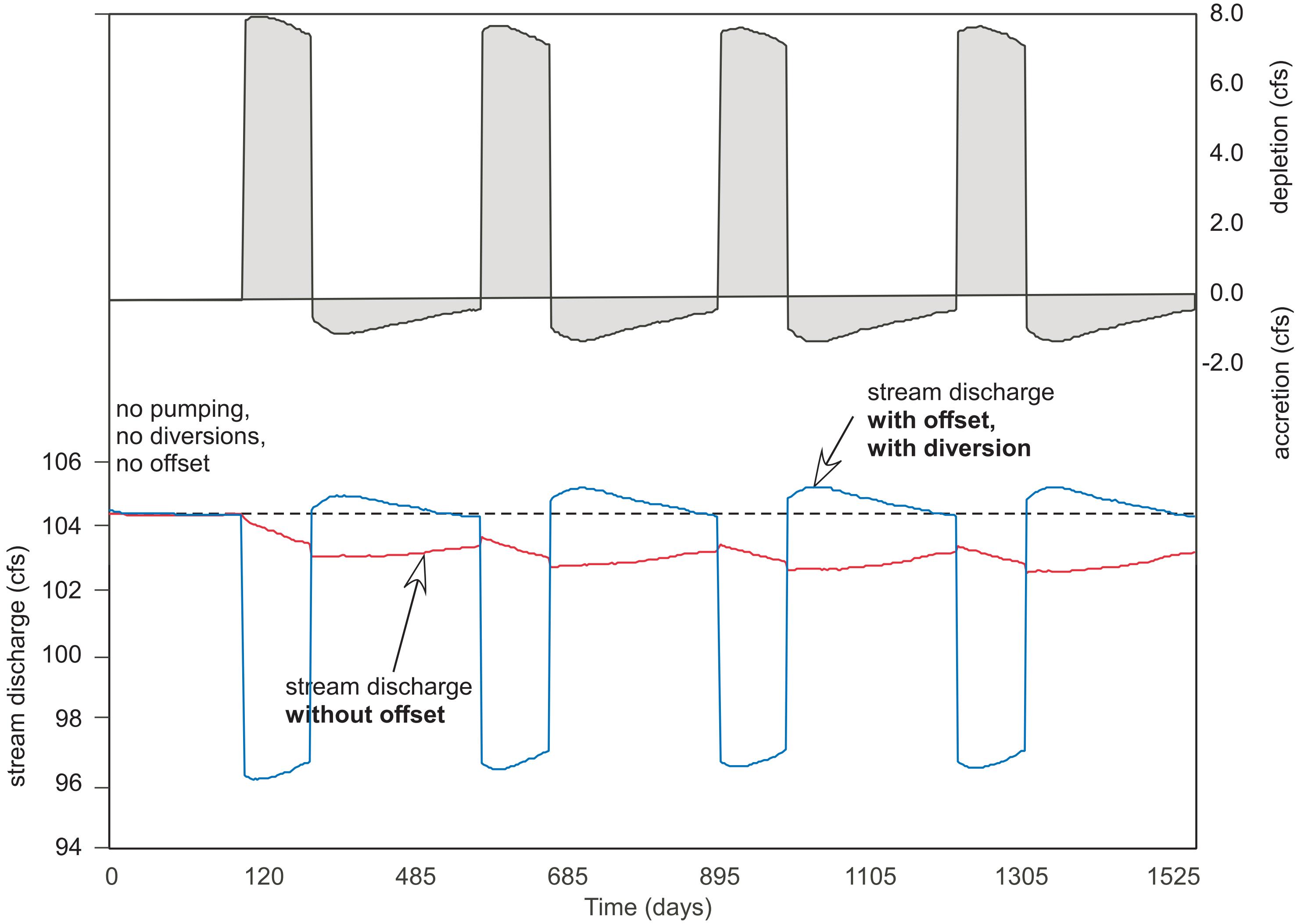


Figure 35. Stream discharge without offset (red line and left axis) shows a declining trend. Stream discharge with offset equal to the 8 cfs volumetric depletion applied to ground water shows the wide range of stream discharge that will result (blue line and left axis). Stream depletion between periods of diversion is much less than zero (gray area graph and right axis) and the offset is actually increasing stream discharge above the original.

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GALLATIN VALLEY CASE STUDY—A BASIN-SCALE GROUND-WATER MODEL TO EVALUATE GROUND-WATER AND SURFACE-WATER INTERACTION

by Kirk Waren, Hydrogeologist

EXECUTIVE SUMMARY

The interaction of ground water and surface water in the Gallatin Valley is investigated using a ground-water model with a stream routing package to calculate surface-water gains and losses in the system. The approach involves portraying the geometry and geology of the basin using digital elevation models, stream locations, estimated aquifer properties, and available water budget data. Because of a detailed water resources study published by the U.S. Geological Survey in 1960 (Hackett and others, 1960) and revisited by the U.S. Geological Survey in cooperation with the Montana Bureau of Mines and Geology (MBMG) in the mid 1990s (Slagle and others, 1995), there are considerable data available to conduct this type of preliminary assessment.

The results of the model demonstrate that the effects of irrigation return flows are extreme in the Gallatin Valley. The model runs generate about 116 cubic feet per second (cfs) of irrigation return flow at the end of a modeled irrigation season of 152 days, and that rate is only about 25 to 30 percent of actual observed irrigation return flows. Furthermore, the model demonstrates that a high-capacity well field, sufficiently removed from the stream by distance and depth, can operate with less than 2 percent of the total withdrawal rate appearing as stream-flow depletions at the end of an irrigation season. Such modest impacts could easily be mitigated by stream flow augmentation from ground water, if necessary.

Long-term surface-water gauging, ground-water level monitoring, and water-quality data collection can be used to constantly monitor the overall health of the hydrologic system, to make sure that ground-water pumping or other activities are not exceeding the ability of the system to function without detrimental impacts to surface-water flow. Pumping water out of the ground-water reservoir creates storage space that can be refilled during times of high surface-water flow. If particular problems in surface-water outflow are identified, there are numerous water management strategies that can be implemented to address such temporal shortages. Water-quality data must continue to be collected and assessed, more so near any activities that may threaten the quality of the ground-water system. By addressing these shortages directly, the use of ground water throughout the basin and its tributaries can continue.

INTRODUCTION

This investigation demonstrates the application of a ground-water model to evaluate certain aspects of the water budget in the Gallatin Valley, including the impacts of periodically pumping high-yield wells to stream flow. Due to time constraints and, to a lesser extent, data constraints, this demonstration is not a definitive work, but a tool to demonstrate the technical methods available and the kinds of results they can provide. These results show how the ground-water model will calculate stream-flow depletions under stresses of pumping wells and how depletions may affect stream flow. However, this model effort does not account for high seasonal flows and other factors that will affect the actual result of stream-flow depletions. There are a multitude of available data that can be used to develop a much more accurate and defensible ground-water model.

The area of investigation is shown in figure 1. The irrigated acreage from water resources surveys shown within the study area boundary was obtained from “An Atlas of Water Resources in Montana by Hydrologic Basin,” undated, circa 1970, Montana Water Resources Board Inventory Series Report No. 11. New irrigation in the area southwest of Manhattan was digitized from recent infrared aerial photographs available on the Montana Natural Resources Inventory System (NRIS) geographical information raster service.

APPROACH

The basic approach entailed assembling applicable geographic and water budget data for the Gallatin Valley and developing an approximate representation of the system using Groundwater Vistas, graphic design software for operating a variety of sophisticated numerical ground-water models. The numerical model used in this work is MODFLOW, a modular, three-dimensional finite-difference ground-water flow model developed by the U.S. Geological Survey (USGS). Due to time constraints, the ground-water model application is limited. The model can be improved by incorporating available borehole and aquifer test data, a more rigorous use of available ground-water and surface-water monitoring data, further grid refinement, the addition of ditches and canals, and extending simulations into more and varied time periods.

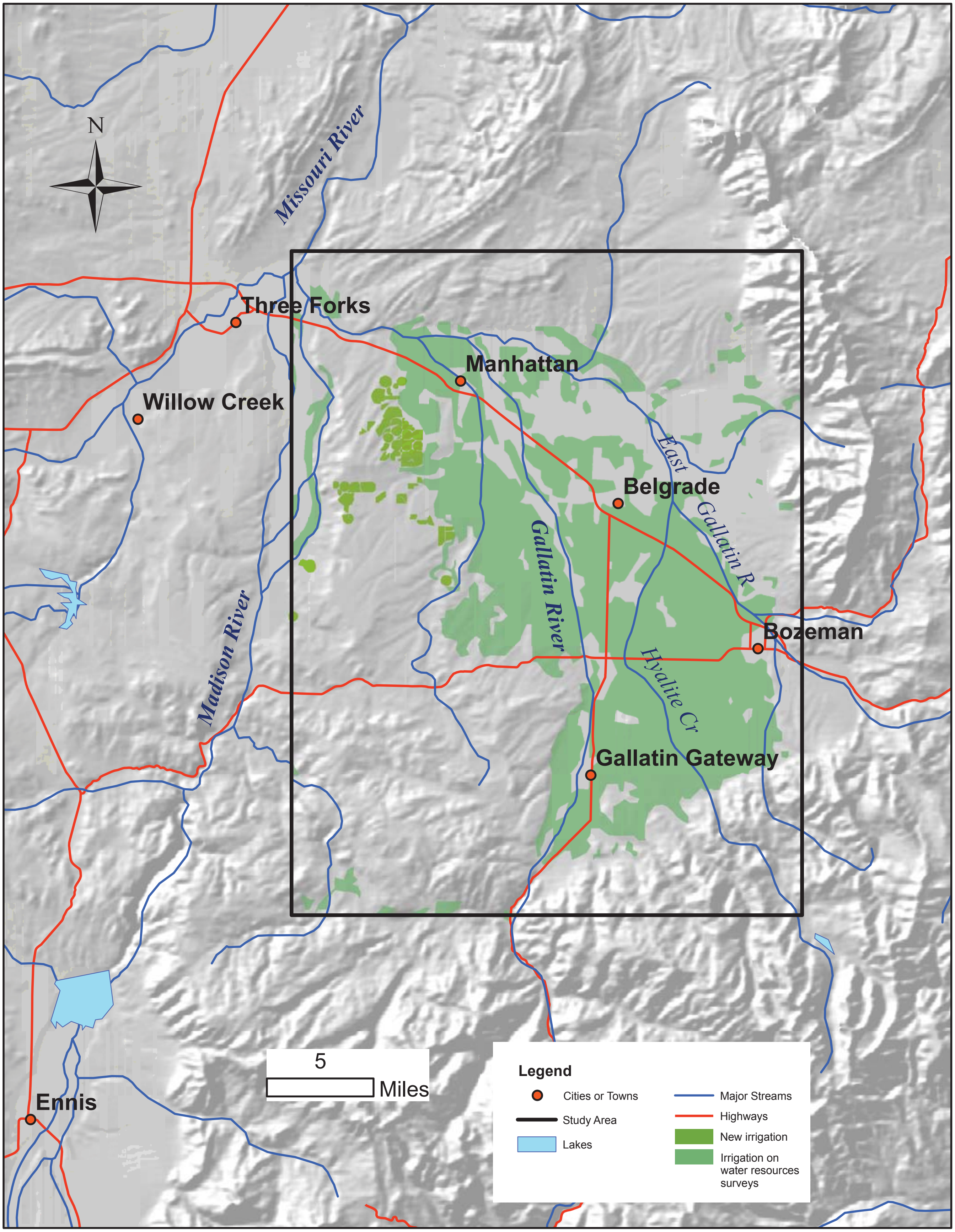


Figure 1. Location of the study area and irrigated lands within the area of investigation.

WATER BUDGET OF THE GALLATIN VALLEY

In 1960, the USGS published "Geology and Ground-Water Resources of the Gallatin Valley, Gallatin County, Montana." This report was authored by O.M. Hackett, F.N. Visher, R.G. McMurtrey, and W.L. Steinhilber, so is sometimes referred to as the "Hackett report" or the "Hackett study." This report provides the results of what is probably the most detailed, comprehensive water budget analysis ever done for such a large area in Montana. While some changes in water use and land use that affect the water budget have occurred within the study area since the study was completed, continued surface-water gauging, ground-water level measurements, and surface- and ground-water quality analyses provide contemporary data to compare with the 1960 study (Slagle, 1995; Montana Ground-Water Information Center (GWIC)).

The Gallatin River is the largest stream in the study area. Its surface-water inflow has been measured near the upstream edge of the study area south of Gallatin Gateway to varying degrees since 1889. The annual mean flow is about 814 cubic feet per second (cfs). The lowest flows tend to occur in the early winter months, with monthly mean flows of just over 300 cfs in January, February, and March. The highest flows are typically in the late spring. The mean monthly discharge for June is about 2,950 cfs.

The outflow from the Gallatin Valley is measured at a nick point at Logan. There, the average mean flow is 1,068 cfs. The mean monthly flow for winter months ranges from about 525 to 560 cfs, so there is a significant contribution of water, over 200 cfs, to the Gallatin River during winter months within the area of investigation between the gauges at the inflow and outflow points. The East Gallatin River accounts for a little less than 40 cfs of these accretions. The rest of the increase in flow consists almost entirely of ground-water discharge.

Figure 2 is a reproduction of the data shown in table 23 and plate 8 of the 1960 USGS report. This figure displays a monthly inventory of the surface-water resources of the Gallatin Valley. It shows monthly inflows and outflows, and depletions and accretions to surface-water resources in the valley. The figure spans two water-years: years beginning October 1st and ending September 30th. The main vertical axis is in thousands of acre-feet. Note that during the months of high surface-water inflow and outflow, the volumes of water are in the range of 100,000 to 400,000 acre-ft of water. The mean annual outflow at Logan for the period of record (1894 to 2002) is 774,000 acre-ft.

The USGS study published in 1960 also included a characterization of ground-water resources in the valley and their interaction with surface-water resources. They assessed and described the principal aquifers in the valley, conducted aquifer tests to determine basic aquifer properties, determined the direction of ground-water movement, and evaluated the quantity and quality of water recharging and discharging from the aquifers. The USGS, in cooperation with the MBMG, revisited the study area in the early 1990s and in 1995 published "Geohydrologic conditions and land use in the Gallatin Valley, Southwestern Montana." The ground-water map in figure 3 depicts the results of the ground-water mapping from the 1995 report, which included portions of the Madison Plateau on the western flank of the Gallatin Valley. The ground-water contours have been converted from feet to meters for consistency with the ground-water model.

According to the 1960 USGS study, the Gallatin Valley ground-water reservoir is recharged principally by stream flow and irrigation water and only in small part by direct infiltration of precipitation and snowmelt. Some reaches of the Gallatin River and other streams in the Gallatin Valley lose water to the aquifer during part or all of the year. An estimated 300,000 to 400,000 acre-ft of water are diverted from streams for irrigation, and at least half of that water recharges aquifers. Using more than one method, the study found that some 240,000 acre-ft of ground water discharges from the aquifers annually. This water discharges prinicipally in the northern, downgradient end of the valley in a ground-water discharge area where the water table is shallow and springs and gaining streams prevail.

The USGS also mapped the rise and fall of the ground-water surface in the Gallatin Valley (USGS, 1960, plates 5 and 6). The water table is less than 10 ft deep in much of the valley. Where the water table is shallow, the storage capacity of the aquifer is limited compared to areas where the water table is deeper, such as in the central part of the valley near Belgrade.

Changes that affect the hydrology of the valley since the study in 1960 include the expansion of urban or subdivided land into areas that were formerly irrigated, and the addition of irrigation wells west of the Gallatin River on the Madison Plateau, mainly between and to the west of the area between the towns of Manhattan and Churchill. As of January 1990, about 36,000 acre-ft of ground water was permitted for use in this vicinity. Actual withdrawals are estimated to be on the order of 27,000 acre-ft per year, based on the reported acreage for water rights in January 1990. This value is available from a compilation of data conducted by DNRC in 1990. It is estimated from recent aerial photographs and anecdotal information that the bulk of ground-water irrigation in the area today was in place by 1990. The 1995 USGS report found that there were no significant water-level changes resulting from increased ground-water withdrawals in the Gallatin Valley.

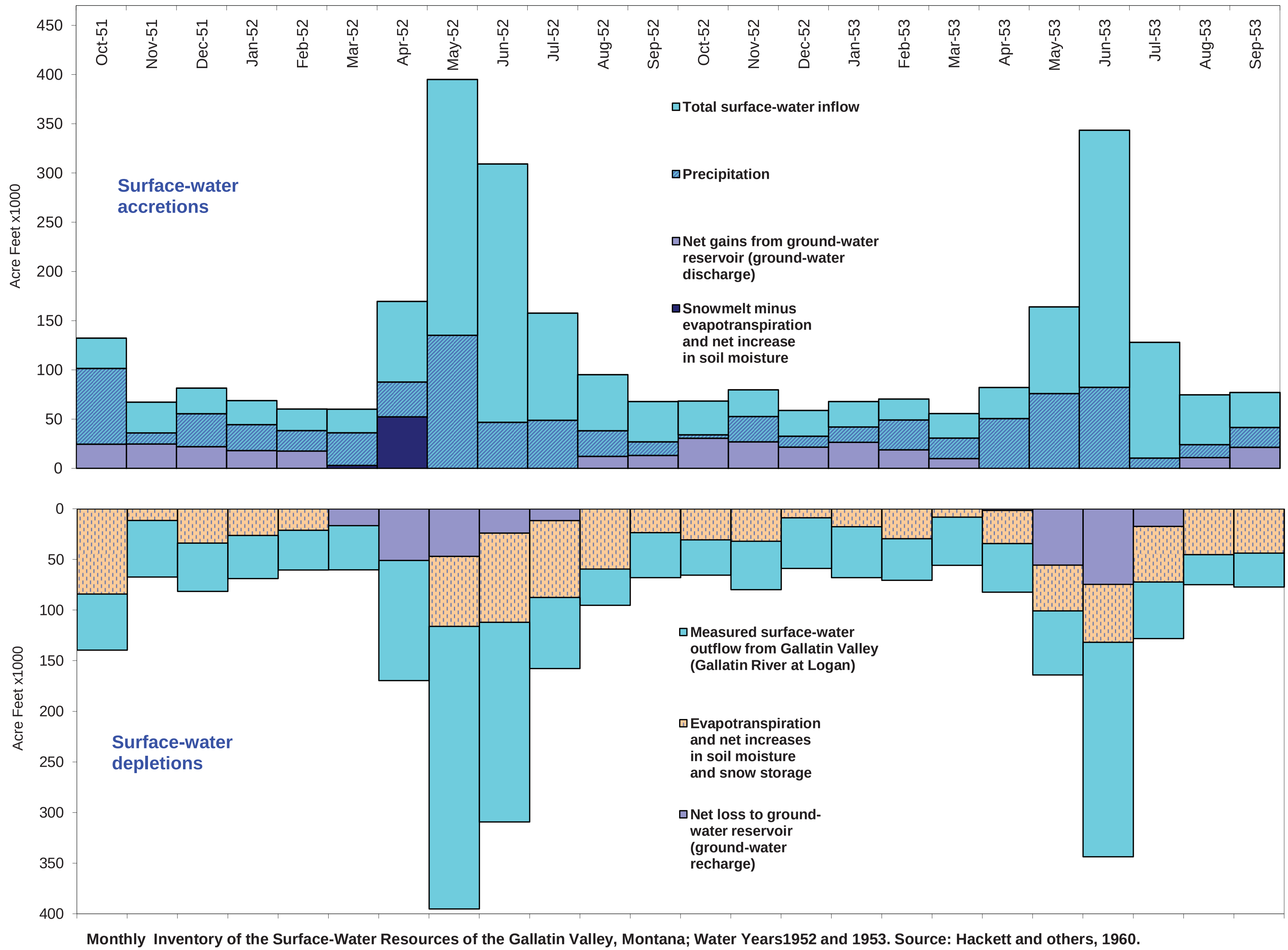


Figure 2. Monthly inventory of the surface-water resources of the Gallatin Valley, Montana: water-years 1952 and 1953. Source: Hackett and others, 1960.

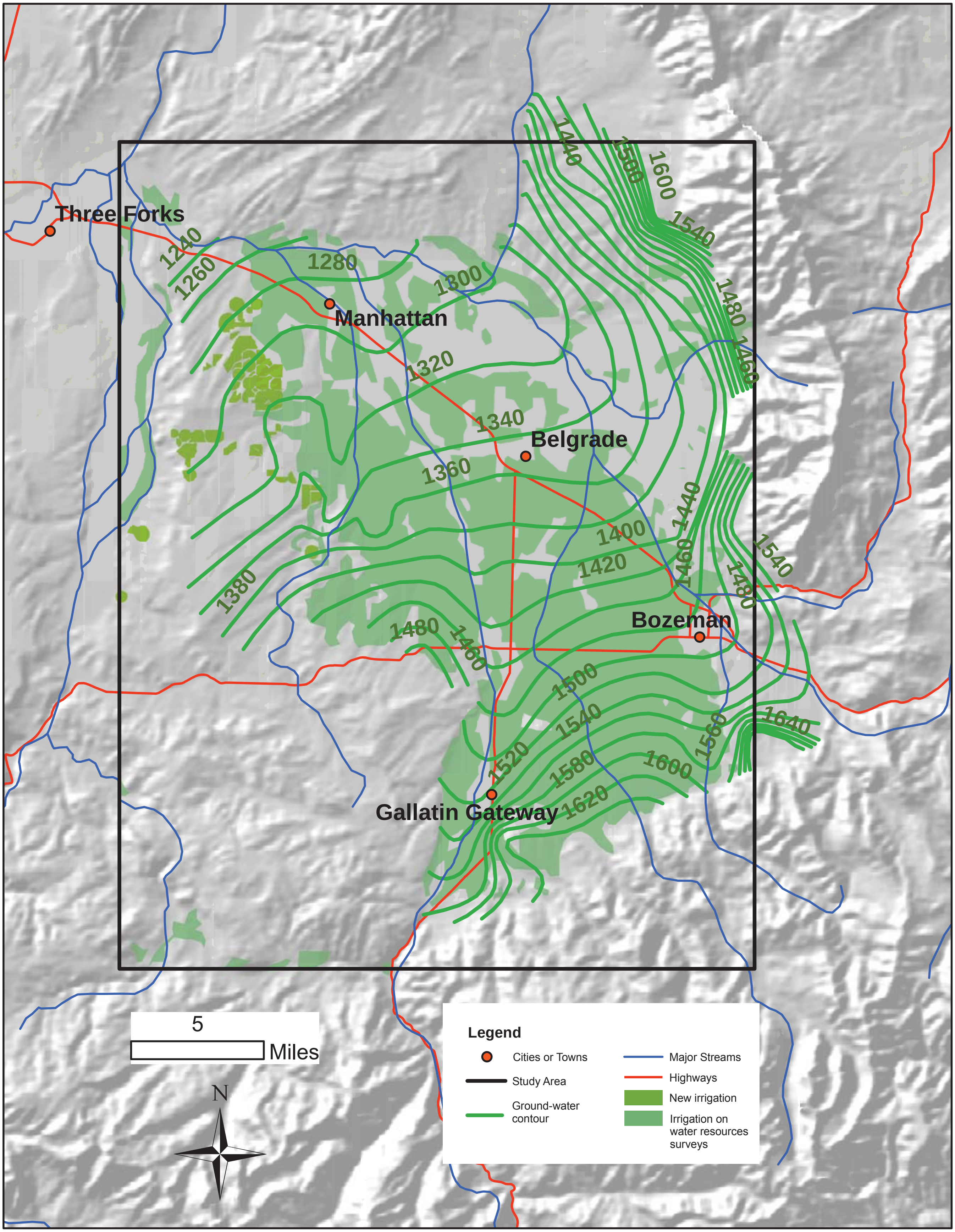


Figure 3. Contoured map of the ground-water surface. Contour interval 20 meters. Source: Slagle, 1995.

GEOLOGY

The geologic materials in the Gallatin Valley constitute the medium in which ground water is stored and transmitted. Detailed geologic maps are available for this area (Hackett and others, 1960; Lonn and English, 2002; Vuke, 2003), and there are considerable data available in the form of well logs and aquifer tests (DNRC, 1992; GWIC). A detailed compilation of available data was not possible in the time frame allotted for this study. Therefore, general aquifer parameters were assigned based on the principal geologic formations present as shown on the Geologic Map of Montana (Vuke and others, 2007). Figure 4 illustrates predominate geologic materials relevant to this effort.

Alluvium and older alluvium in the valley floor areas are stream-deposited sand, gravel, silt, and clay. The gravel cap units are probably either perched aquifers or dry, in either case probably not connected with the principal ground-water flow system. The alluvial fan deposits are less permeable than the valley sediments as evident by the steeper mapped ground-water gradients. The upper and middle Tertiary sediments tend to contain semi-consolidated rocks such as siltstone and mudstone, and are generally less permeable than the alluvium of the Gallatin Valley. An exception are layered, gravel aquifers south and southwest of Manhatten, generally coinciding with the “new irrigation” locations mapped in figures 1 and 3. Indeed, these gravels are probably the source for most if not all of the new irrigation. These aquifers occur in layers underground at depths of about 100 to 1,000 ft deep.

The Central Park Fault (the dashed line south of Manhattan in figure 4) displaces the Tertiary rocks upward north of the fault line, causing the alluvium to thin dramatically north of the fault.

STEADY-STATE GROUND-WATER MODEL

The study area, or ground-water model boundary, was selected to encompass the entire Gallatin Valley. This approach was selected because the water budget information available for the Gallatin Valley is directly comparable and applicable to the modeled area. The basic approach to the model was to avoid constant head boundaries, and use the natural constraints and geometry of the valley aquifers, stream cells to represent major streams and their ability to lose or gain water from the ground-water system, and aerial recharge to simulate recharge from precipitation or from irrigation activities.

A model grid was selected with square cells 500 meters (1,640 ft) wide. Units of meters (m) and days (d) are used for distance and length in the ground-water model. Meters are used because most data available in the NRIS database is provided in the projected coordinate system of State Plane, meters. The model was designed with three layers vertically. Layer 1 is closest to the surface, and represents materials in the upper 30 m (98 ft). Layer 2 is 60 m (197 ft) thick, and layer 3 is 90 m (295 ft) thick, so all three layers represent the uppermost 180 m or 590 ft of geologic materials.

A model topography was generated by creating a surface from available USGS digital elevation model (DEM) data available through NRIS. This is used to calculate surface elevations for each cell representing the top of layer 1. The tops and bottoms of deeper layers were then computed from the top elevation of layer 1 and the desired depths.

Figure 5 shows the ground-water model grid developed for the model, with horizontal hydraulic conductivity zones for layer 1 shown by the various colors. Hydraulic conductivity is a measure of the ability of sediments to transmit water. Throughout the model, in all layers, the vertical hydraulic conductivity is 1/10th of the horizontal value assigned. This is a convention commonly applied for stream-deposited materials. In layer 1, the white cells in the valley floor area represent alluvium and have an assigned hydraulic conductivity of 40 m/d, which is a value typical for a relatively high-permeability material such as sand and gravel, and is equivalent to nearly 1,000 gal/ft2/d in common units. This value is lowered north of the Central Park Fault (gray cells) to 25 m/d, which has the same effect in the model simulation as making layer 1 thinner. The green areas at the valley margins have an assigned hydraulic conductivity of 1 m/d, and the yellow area has a value of 2 m/d.

The locations of modeled stream cells representing major streams are shown in figure 5 as the blue cells along the stream courses of the Gallatin River, East Gallatin River, and Hyalite Creek. In the model simulation, these cells calculate stream-flow gains and losses to the ground-water system.

Figures 6 and 7 show the assigned properties for layers 2 and 3. The color assignments are the same as in layer 1, so much of layer 2 is assigned hydraulic conductivities of 1 or 2 m/d, except for cells in the white area in the valley which are assigned a hydraulic conductivity of 40 m/d, representing deeper alluvial materials in the valley south of the Central Park Fault. In layer 3, the peach-colored areas have an assigned hydraulic conductivity value of 0.6 m/d, the pink areas in the valley 5 m/d, the blue area 10 m/d (representing the buried gravel layers), and the orange area 2 m/d.

Figure 8 shows a cross-section view of the model in an east–west direction about in the middle of the model. This is shown to illustrate how the basic model is constructed. Note the extreme vertical exaggeration.

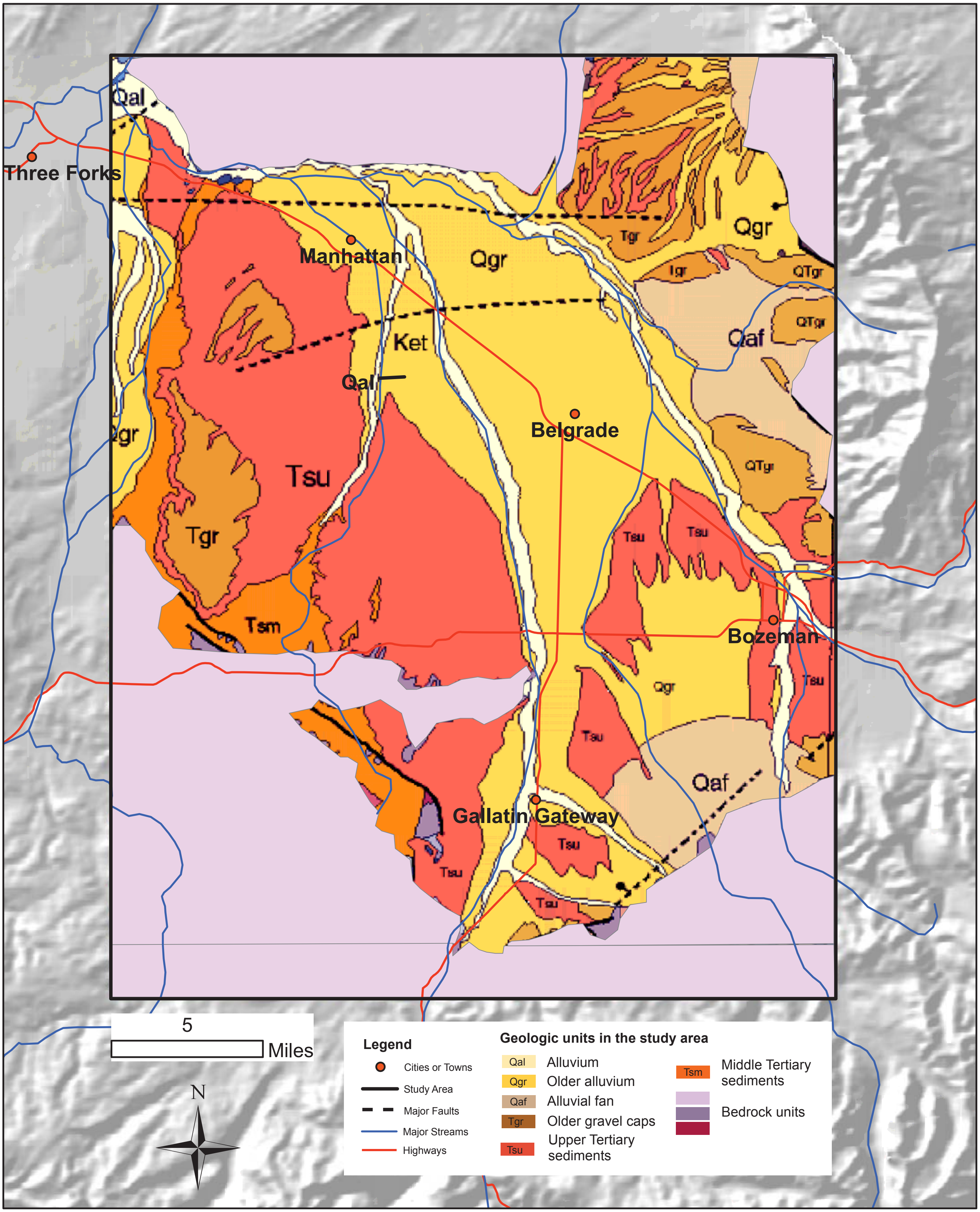


Figure 4. Generalized geology of the Gallatin Valley. Modified from Vuke and others, 2007.

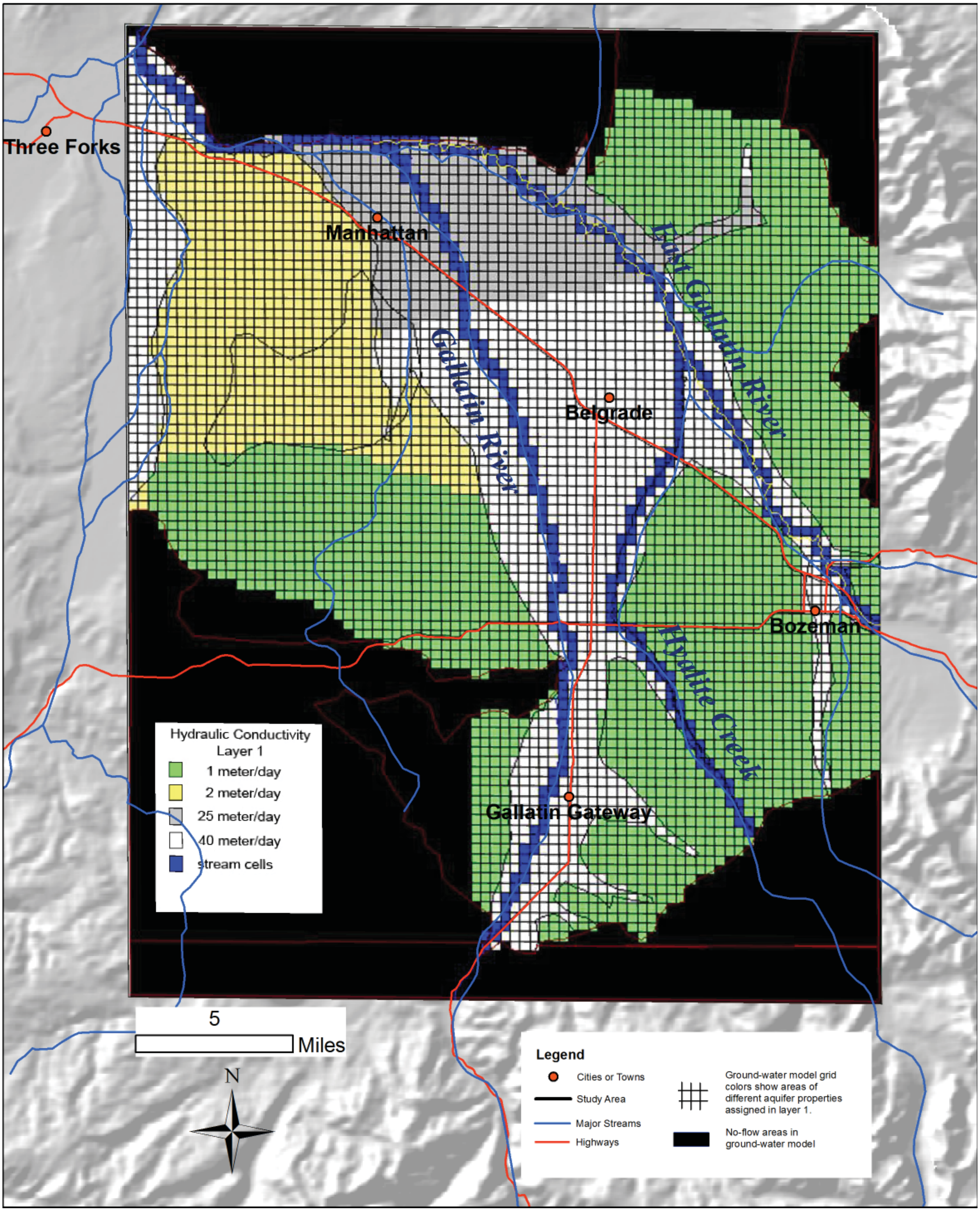


Figure 5. Ground-water model grid, hydraulic conductivities, and stream cells in layer 1.

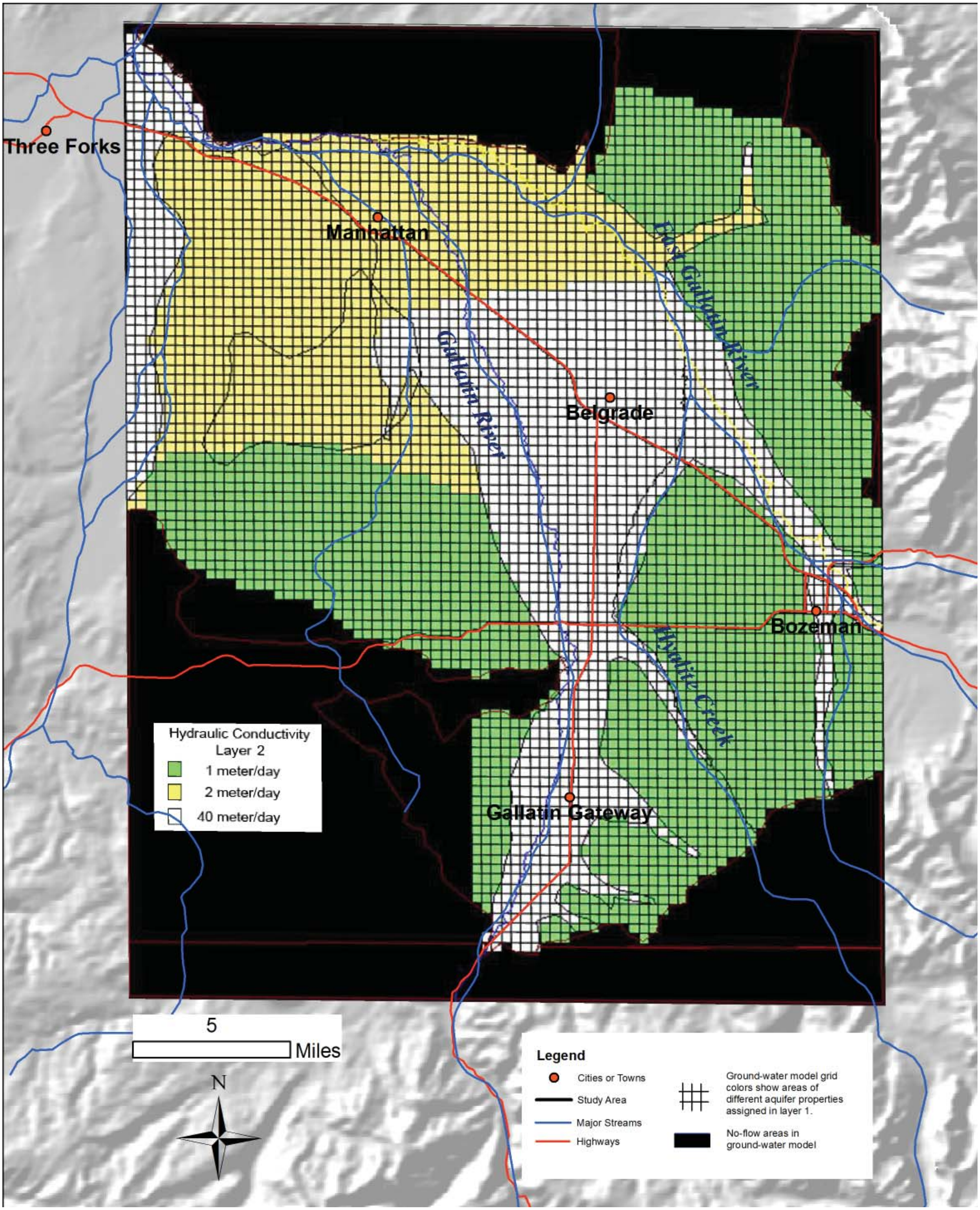


Figure 6. Ground-water model grid and hydraulic conductivities in layer 2.

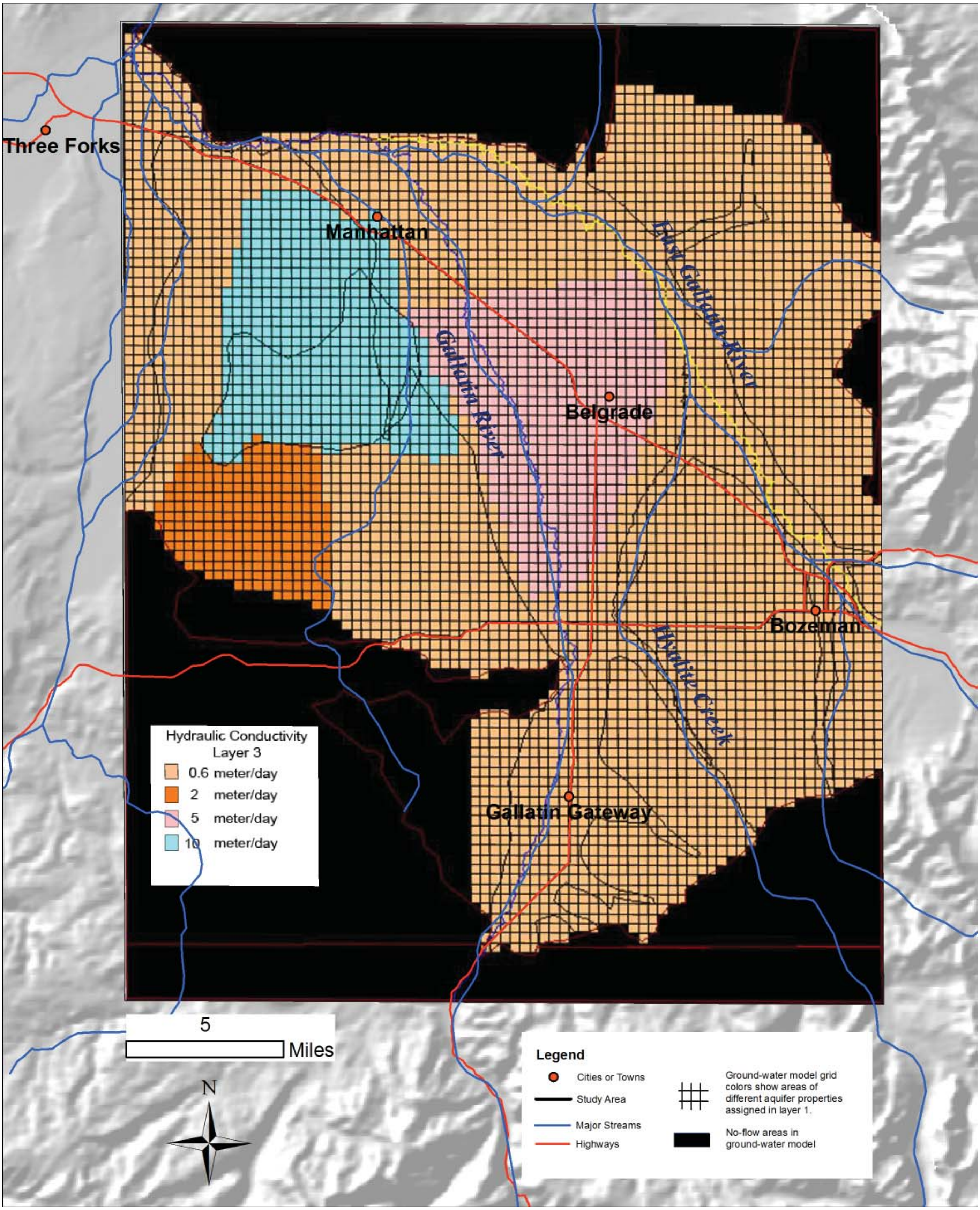
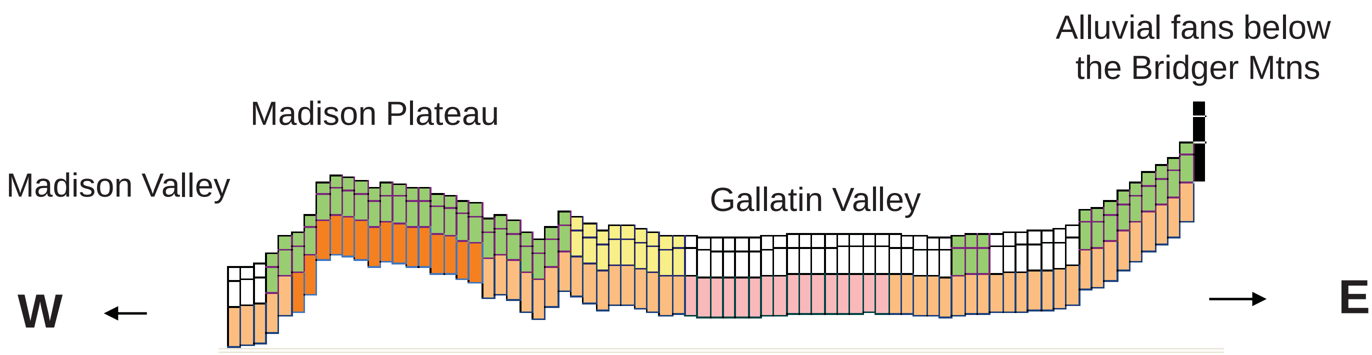


Figure 7. Ground-water model grid and hydraulic conductivities in layer 3.

Profile View of the Ground-Water Model



Vertically exaggerated. Width of model: 38,500 m; height 180 m

Figure 8. East–west profile view in about the middle of the ground-water model. The hydraulic conductivies assigned in each layer are explained in figures 5 through 7.

The steady-state model was developed to approximate conditions observed during low stream-flow conditions in the Gallatin River, typically observed during winter months. Due to the effects of irrigation, natural low water conditions are probably not often achieved in portions of valley aquifers. Numerous monitoring wells in the valley indicate declining ground-water levels throughout winter months as aquifers charged by excess irrigation water discharge water as irrigation return flow. This continues at many sites until the next irrigation season begins and both stream and ground-water levels rise. An example of this phenomenon is shown in figure 9. Note how during winter months the ground-water level in this well, located about 2 miles north of Belgrade, falls until around the start of April each year. This behavior is observed in numerous area monitoring wells. Therefore, the natural low ground-water setting can only be approximated by using seasonal low ground-water level data.

Typically, all available ground-water elevation data are assembled and used as calibration targets for a numerical ground-water model. A considerable amount

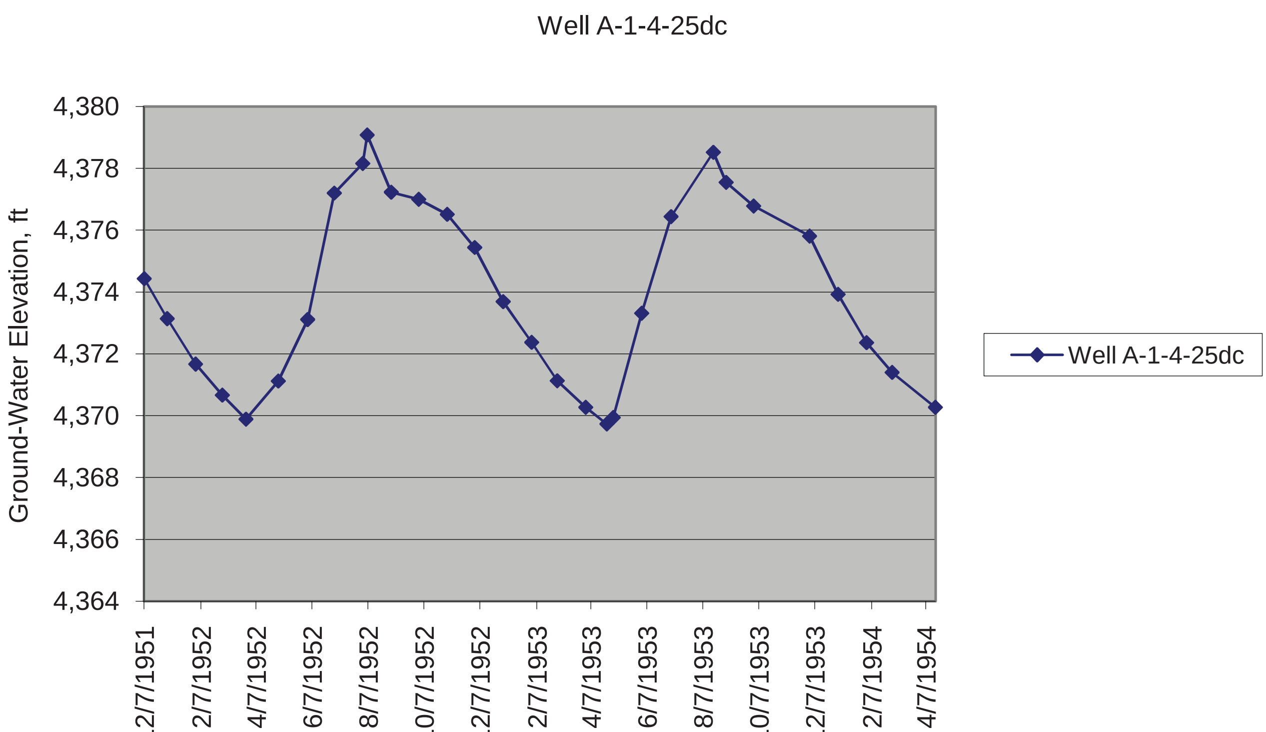


Figure 9. Seasonal ground-water level trends in a typical well about 2 miles north of Belgrade.

of such data is available for the Gallatin Valley. Due to the limited time available for this analysis, this approach was not feasible for the current effort. Instead, an approximate visual calibration of modeled conditions was undertaken using ground-water surface elevation contour maps. In addition to the map rendered in figure 3, a ground-water level map completed by the Montana Department of Natural Resources and Conservation (DNRC, 1992) for the Madison Plateau area was also used to develop a conceptual idea of what the ground-water surface model should look like. One point of data that results in a very different rendition of conditions at the north end of the Madison Plateau between the two maps is due to a reported water level of 550 ft in a well of record located about 5 miles west–southwest of Manhattan, in a remote area high on the plateau (Section 26, T. 01 N., R. 02 E.). This well was recently visited by MBMG personnel (May 2008) and the water was determined to be more than 500 ft deep, the maximum depth measurable with the equipment available. Another visit is planned with instruments capable of deeper measurements.

As noted above, stream cells are used in the model to simulate major streams. The stream routing package for MODFLOW was used for this application. Streambed conductance was adjusted until simulated water movement between the ground-water and surface-water systems were comparable to observed values. A logical early step in developing a more accurate ground-water model would be to add stream cells for smaller water bodies such as irrigation canals and drainage ditches. This would provide a more realistic rendition of how excess irrigation water enters and discharges from the aquifer.

Aerial recharge was added to the entire active model area at a value of about 6 inches per year. This value was derived by adjustments in recharge and aquifer properties until a reasonable rendition of the ground-water surface was achieved. This rate of application generates about 133,000 acre-ft of recharge in the model. This is about 30 percent of the average annual precipitation measured for the Gallatin Valley during water-years 1952 and 1953 (USGS, 1960, table 23). This value may have to be higher than actual values of recharge from precipitation to force the model to generate a water table configuration that is likely influenced by irrigation activities.

Figure 10 shows the results of the steady-state ground-water model. A steady-state model represents the calculated state of the system given conditions that remain steady over time. The contoured image shows the numerically generated ground-water surface. The purple areas are dry cells in the uppermost layer, layer 1. Most cells in the lowest layer, layer 3, are saturated, generating the ground-water contours in these areas where there are dry cells in layer 1.

An evaluation of the steady-state ground-water model budget indicates that stream cells generally north of an east–west line about 2 miles south of Belgrade gain water at a total rate of about 14,000 acre-ft per month. Cells upstream of these reaches lose some 2,500 acre-ft per month, so net gains for the entire model are about 11,500 acre-ft/month. This compares with estimated net winter monthly gains ranging from about 10,000 to 31,000 acre-ft for the Gallatin Valley (USGS, 1960, table 23). Since the model design attempts to replicate low flow conditions, the modeled gains compare reasonably with the water budget for the purposes of this demonstration.

TRANSIENT MODELING

Transient modeling allows the ability to analyze time-dependent problems or activities. For this demonstration, the effects of recharge from excess irrigation water and the introduction of a well field are analyzed. In the first simulation only the irrigation recharge is applied; in the second simulation both irrigation recharge and the well field are applied. The results demonstrate how the ground-water model responds to these stresses and how the model calculates impacts to stream flow.

The USGS (1960) study resulted in a map showing the difference in position of the water table seasonally in the Gallatin Valley. Generally, ground-water mounding occurs in some areas due to summer irrigation activities. Its pattern is controlled by variety of factors such as the locations of canals and irrigation activities, the available unsaturated space in the water table aquifer, the drainage characteristics of the soil and underlying material, the presence of drainage ditches, and so forth.

An advanced modeling effort would involve a determination of ditches or water courses that act as drains or water sources, or both, and incorporate them into the model. Also, a detailed inventory of water distribution and movement in the valley would be useful in assigning specific values of recharge for various irrigated areas. For example, for an area irrigated by a single ditch, the total flow can be determined and estimates of crop use made using the Montana Irrigation Guide (U.S. Soil Conservation Service, 1987). The remaining water likely returns to the ground-water and surface-water system in some manner. The general conditions at various sites can be used to estimate how much water might be assigned as ground-water recharge from excess irrigation water. Other useful information would be measurements or estimates of ditch leakage and the type of irrigation methods used, such as flood or sprinkler irrigation.

For the purposes of this demonstration, recharge is added in areas where ground-water mounding is expected based on the USGS (1960) study and ad-

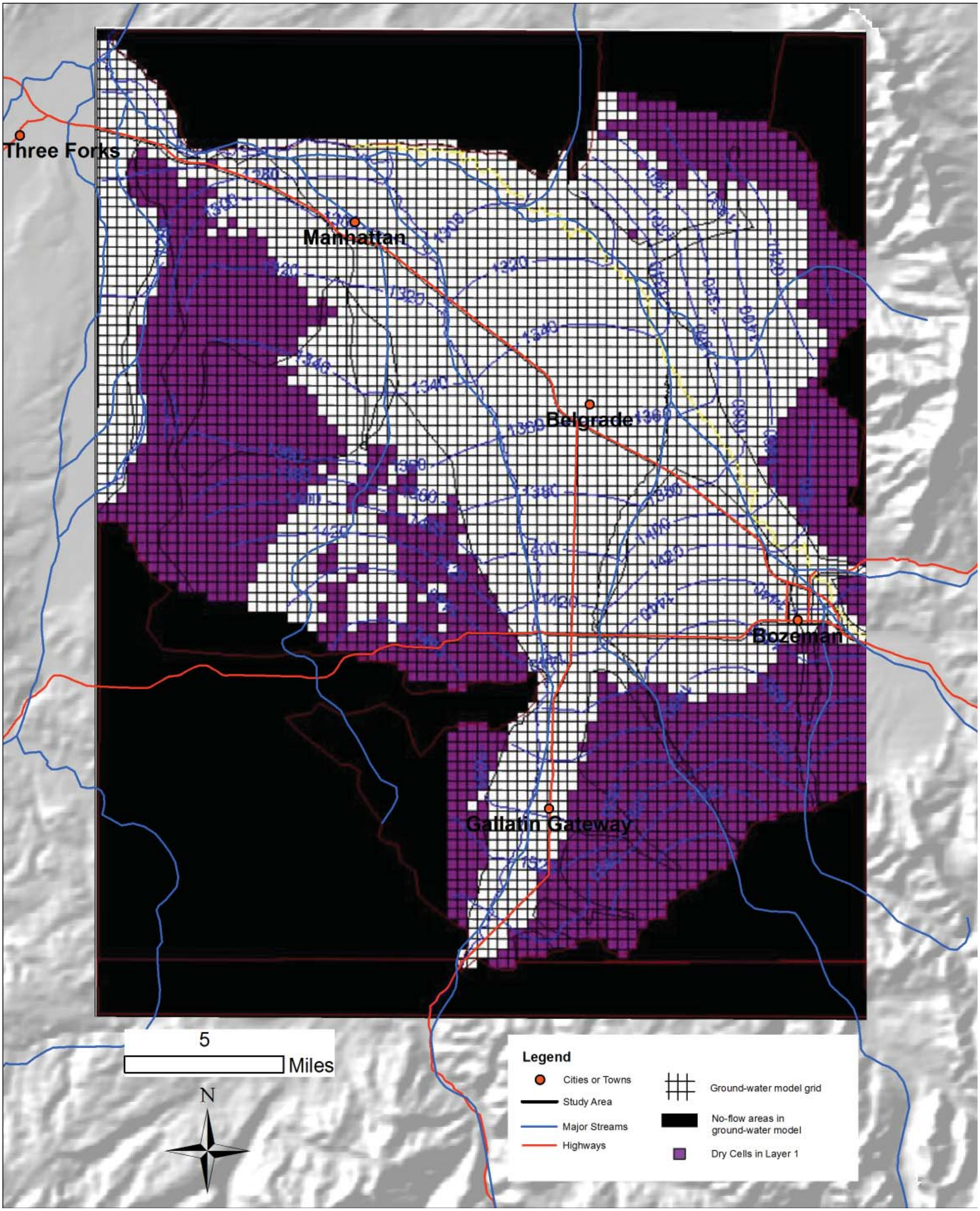


Figure 10. Ground-water model calculated steady-state ground-water surface. Contour interval 20 m (66 ft).

ditional data gathered by Slagle (1995) and DNRC (1992). The distribution of this additional recharge to simulate recharge from excess irrigation water is shown in figure 11. This recharge is applied in the model simulation during stress period 2, which has a modeled length of 152 days, representing a 5-month irrigation season. The actual irrigation season is typically as much as a month shorter in this area, but five months is chosen to reflect a maximum expected season length for irrigation and, in later modeling, pumping of irrigation wells. The recharge of the darker green areas on the valley floor is equivalent to 8 ft of recharge for the irrigation season, simulating recharge from flood irrigation activities on the porous materials of the valley floor. The recharge of the lighter green areas on the areas adjacent to the valley floor are assigned recharge equivalent to 2 ft over the irrigation season, representing a lesser amount of recharge to the ground-water system due to less porous materials in the subsurface and perhaps the use of sprinkler irrigation systems. Notice that the total acreage modeled is probably somewhat less than the areas irrigated as illustrated in figures 1 and 3.

The model has six stress periods. Stress period 1 has a length of 213 days, and is essentially a continuation of the steady-state model. This stress period is used to verify that the appropriate starting conditions are applied to the transient model. Stress period 2 has a length of 152 days, and represents the irrigation season as noted above. Stress periods 3 through 6 vary with different model runs, and serve to calculate incrementally how conditions recover after the applied stress of stress period 2 is stopped and modeled conditions return to those of stress period 1.

The resulting mounding calculated by the model is shown in figure 12. During this effort, it was found that in general, the modeled valley aquifer will not saturate and drain the amount of aquifer material indicated by actual observations in a similar timespan. This is attributed largely to the likelihood that drainage ditches or stream courses likely serve to drain various irrigated areas more directly than can be simulated by only having the main streams acting as stream cells in this model. The model results indicate approximately 5.5 m, or 18 ft, of mounding in the vicinity of Belgrade. Actual seasonal observations show mounding of up to 43 ft near Belgrade. Similar maps for the end of the following stress periods show a gradual reduction of the mounding effect as calculations simulate water draining away from the mounded areas and showing up as stream-flow gains.

Note the slight drawdown occurring in the northeast part of the image in figure 12. Although the color flood scale renders this drawdown to be readily visible, it represents less than 1 m of calculated drawdown. The calculated drawdown in that particular spot is due to boundary effects caused by the configuration there of the units with higher conductivity in layers one and two. The calculated drawdown effects here may or may not occur in nature. Since it occurs in all of the transient models in a distal portion of the valley, it does not have any significant effect on the model results.

Additional stream-flow gains attributed to the modeled irrigation-related recharge of stress period 2 is calculated to be 116 cfs at the end of stress period 2, as shown in figure 13. That return flow rate to streams decreases over the following stress periods to about 54 cfs at the end of stress period 6. This represents about 25 to 30 percent of return flow rates observed in streams. A detailed accounting of water distribution and irrigation methods and the addition of a more refined ditch drainage system for more advanced modeling could refine this result. This simulation provides an example of how the model reacts to the assigned stress of added recharge to simulate excess irrigation water entering the ground-water system, and how the model calculates its eventual return to streams.

Another scenario modeled is the introduction of a well field in the Madison Plateau area, with well locations spanning the approximate locations of known wells. DNRC (1992) estimated some 36,000 acre-ft of water was permitted in the area as of 1990. New permits since that time have been limited, so this is probably a reasonable value to start with for this effort. Since water rights commonly are calculated based on full season operation, and on average a variety of activities usually reduce the amount of water used (for example, periods of unusually high precipitation, drying of hay for cutting, etc.) it is estimated that actual water use may be about 25,800 acre-ft per year. This amount of withdrawal was spread around 31 wells in the locations shown in figure 14. All of the wells draw water from layer 3 at a modeled rate of 6,760 cubic meters per day (1,240 gallons per minute (gpm)). The pumping rate of all the wells combined is 38,440 gpm, or 87 cfs.

This portion of layer 3 has intermediate values of hydraulic conductivity assigned to represent productive, deep gravel layers known to be present in the area (DNRC, 1992). These gravel layers are suspected to connect to deeper alluvial aquifers to the east, or simply extend eastward beneath the alluvial sediments of the valley floor, or both.

Figure 15 shows the drawdown in layer 3 resulting from the addition of the well field in stress period 2, at the end of stress period 2, and including the same mounding due to irrigation water recharge calculated in the earlier model run described. This result shown is for layer 3, the bottom layer of the model, showing the calculated impacts of ground-water mounding on the shallow layer being propagated to the deeper layer, as well as the impacts of well withdrawals. Figure 16 shows the remaining drawdown impacts at the end of stress period 6, after both the irrigation recharge and well pumping have been off for a model time of about 14 months (424 days).

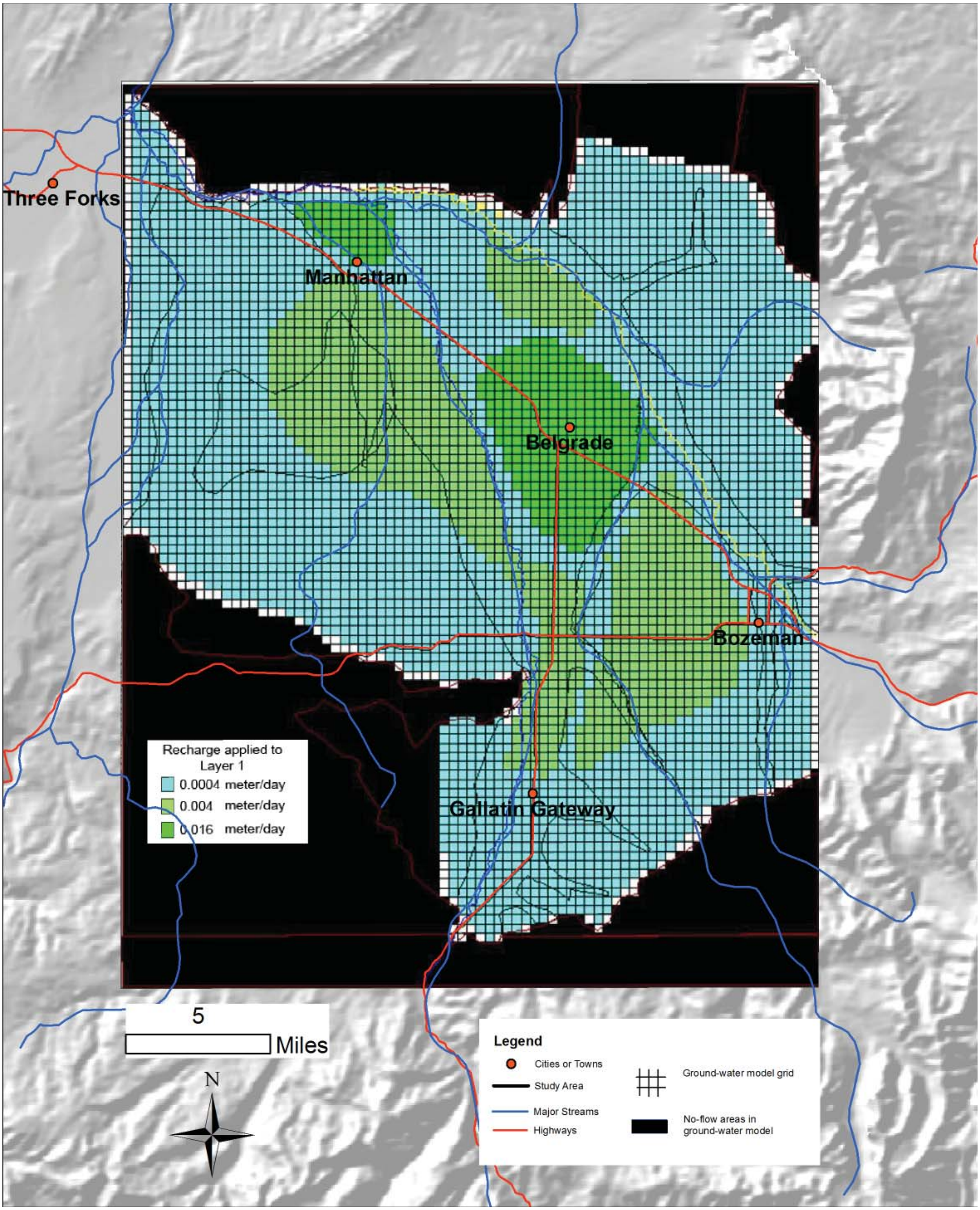


Figure 11. Recharge applied to simulate recharge from irrigation acitivies.

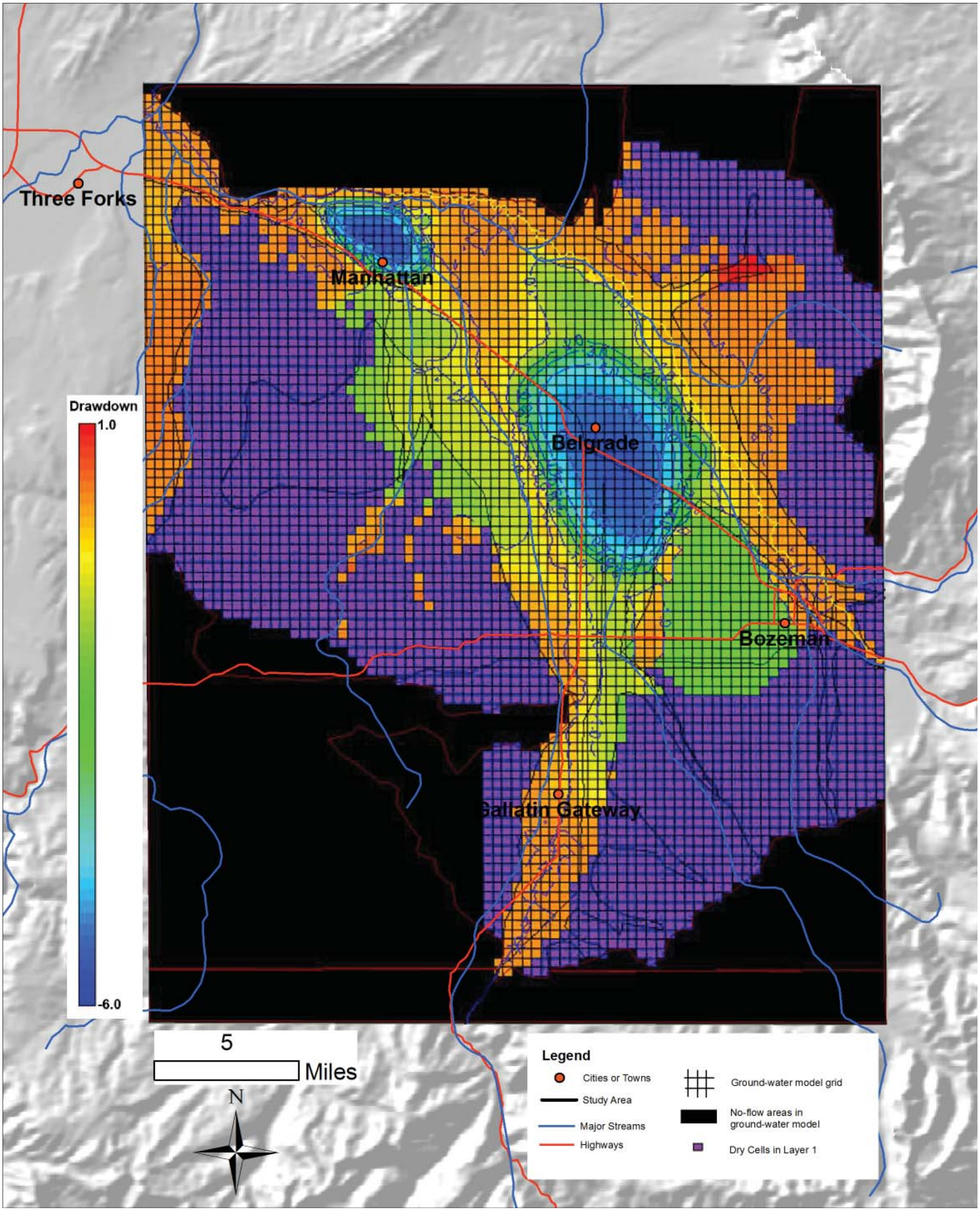


Figure 12. Drawdown map of layer 1 produced by Groundwater Vistas showing mounding of the water table due to recharge as negative drawdown at the end of the irrigation period (stress period 2).

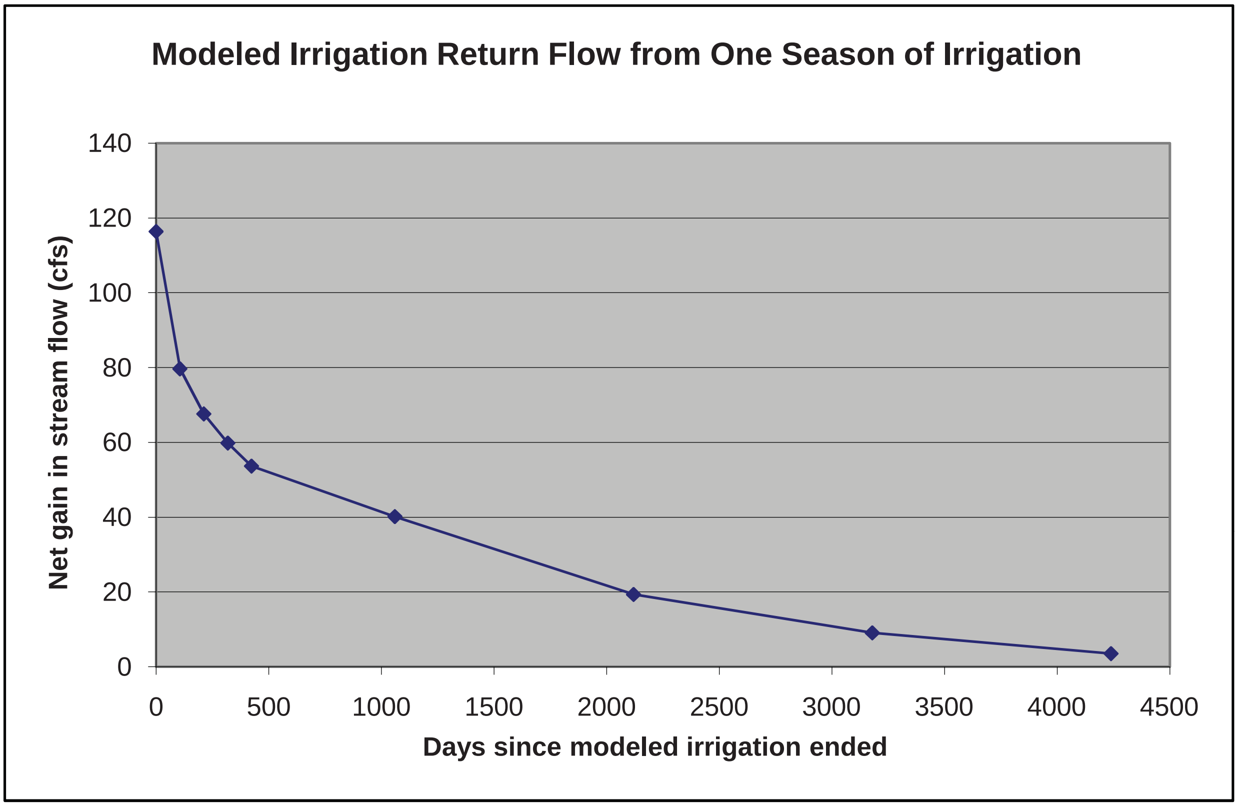


Figure 13. Modeled irrigation return flows from one season of irrigation.

The resulting stream-flow depletion calculated by the ground-water model for pumping the 31 wells in the well field described, pumping 87 cfs, is shown in figure 17. Note that the maximum stream-flow depletion value is approximately 1.4 cfs, or about 1.6% of the pumping rate of the wells. The calculated depletion extends past 10 years (3,650 days), although by that time it diminishes to about 1 cfs. Because stream flow is the only means for water to discharge from the model domain, all 25,800 acre-ft of water must eventually show up as stream-flow depletion, although as this figure suggests, it would take decades. The model simulation calculates all water withdrawn from wells as outflow. In actuality, some amount of that withdrawal would appear as return flow to streams or to the shallow ground-water system.

The ground-water model simulations run to generate the data shown in figure 17 do not include any later pumping periods. Stream-flow depletions from multiple years of pumping would be cumulative. In this model construction, if enough successive years of pumping and recovery were simulated, eventually the stream-flow depletion would equal the amount of water pumped annually spread out over a period of 1 year. It would be 5/7 of 87 cfs, or about 62 cfs, since there is 5 months of pumping and 7 months of recovery each year. However, to achieve this depletion would require many tens of years of pumping. In nature, cyclic wet years, irrigation, and high runoff periods would tend to recharge aquifers at times of high water, making such long-term calculations meaningless.

CONCLUSION

This model shows that a relatively simple and straightforward approach to modeling a large basin readily yields a reasonable rendition of the configuration of the ground-water system with reasonable water budget results. These results can be compared directly with actual data available for basin. The irrigation return flows calculated by applying excess water to selected areas in the model to simulate irrigation are about 25 to 30 percent of actual values observed.

Significantly, the simulation of the well field in the Churchill and Manhattan area suggests that depletions of stream flow from pumping may be both delayed and highly attenuated in such a manner that the impact to the stream from a single season of pumping is less than 2 percent of the pumping rate for all wells combined in the well field. This demonstrates that if wells are sufficiently removed by distance or depth from streams, their impact to streams becomes negligible relative to other influences to stream flow, such as irrigation return flow. This simulation only provides one exit for water from the system, and that is out the stream cells. Furthermore, there is no confining layer in the simulation. The only separation between the well field and the streams in the model is distance and the properties of the aquifer materials modeled.

In this example, a mitigation strategy for the presence of 31 high-yield wells that generate 1.4 cfs of stream depletion at the end of a pumping season would be to simply add one more well capable of pumping 1.4

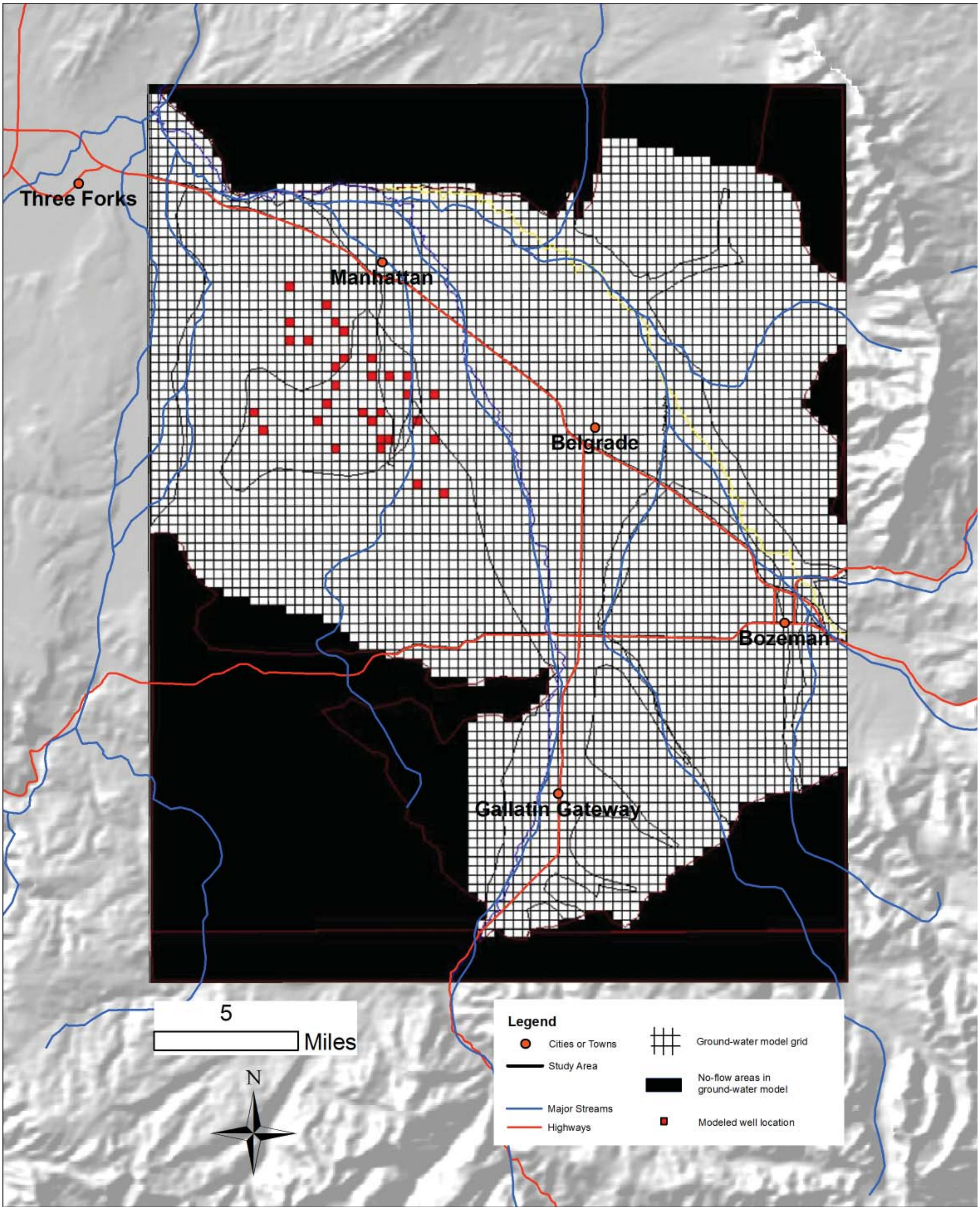


Figure 14. Locations of wells in the well field modeled (layer 3).

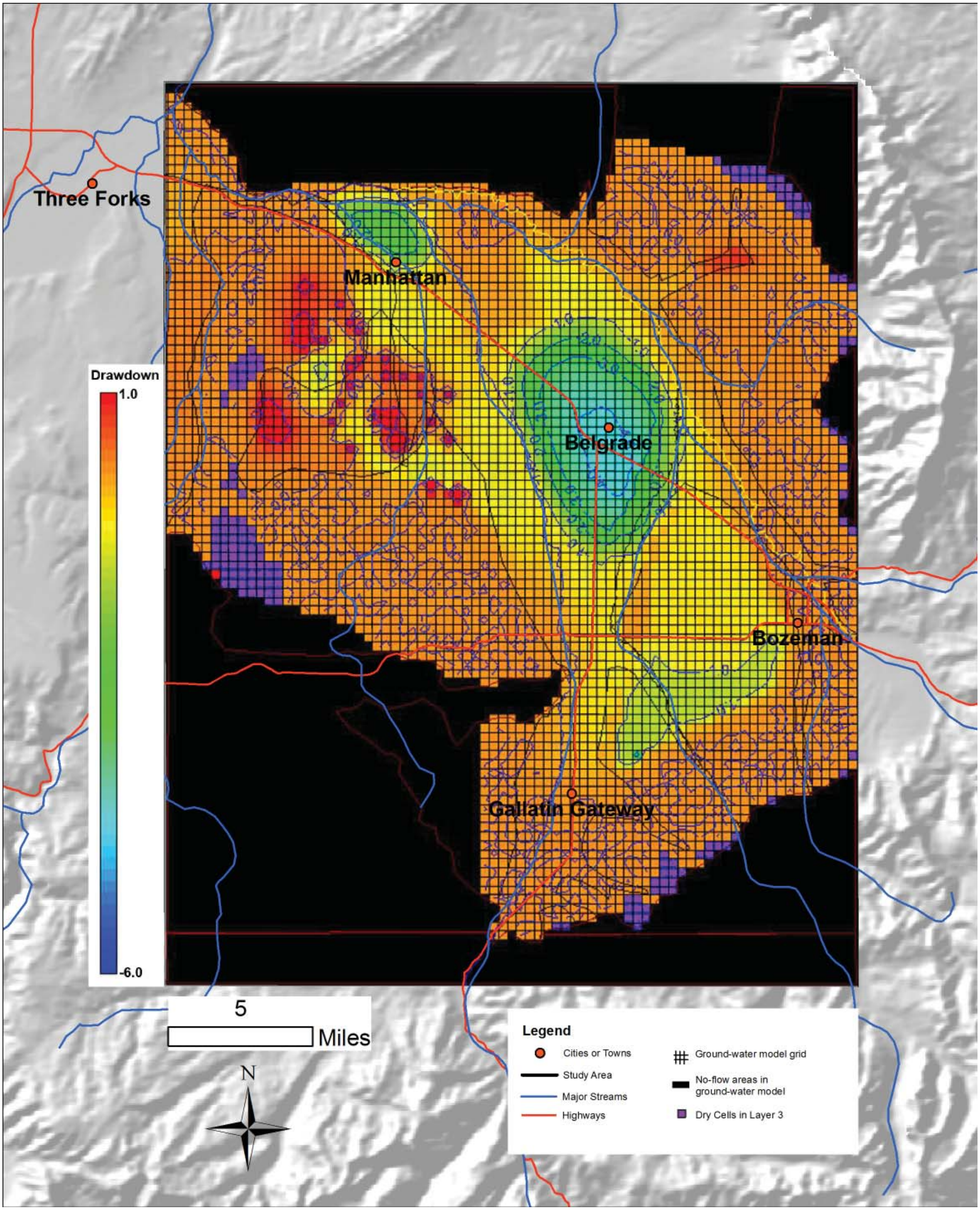


Figure 15. Drawdown map of layer 3 produced by Groundwater Vistas showing both mounding of water in layer 3 due to recharge applied in layer 1 as negative drawdown, and the drawdown due to pumping wells, at the end of the irrigation and pumping period (stress period 2).

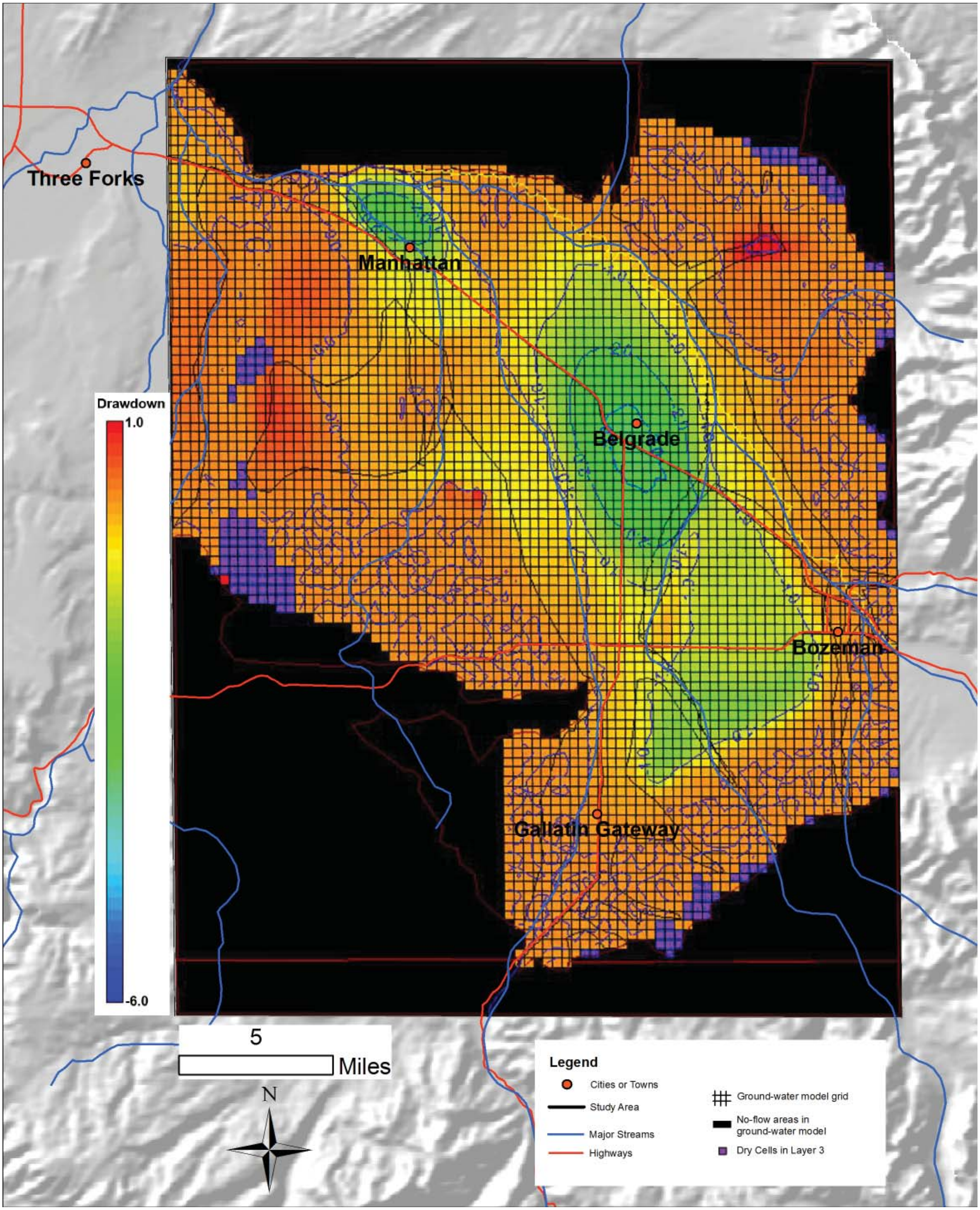


Figure 16. Drawdown map of layer 3 produced by Groundwater Vistas showing both mounding of water in layer 3 due to recharge applied in layer 1 as negative drawdown, and the drawdown due to pumping wells, 318 modeled days after irrigation and pumping ends.

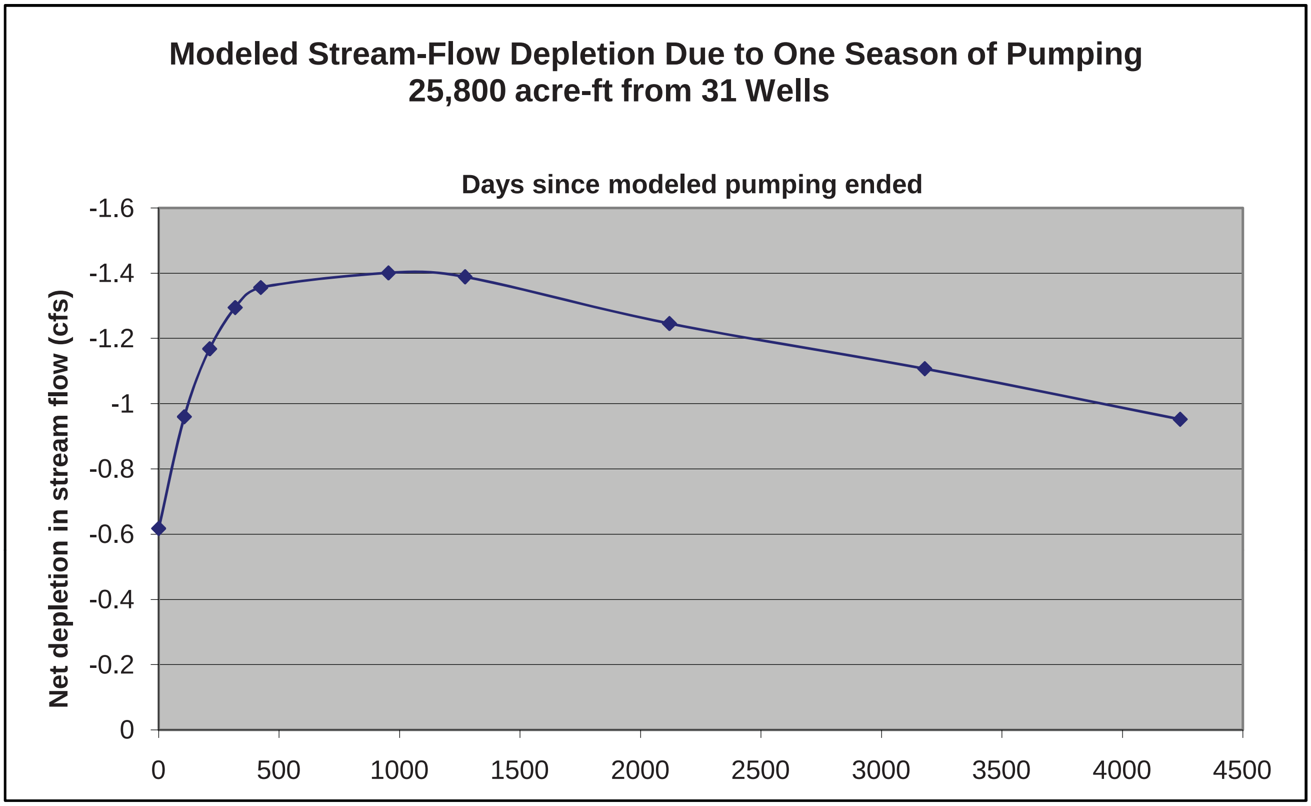


Figure 17. Modeled stream flow depletions due to one season of pumping 25,800 acre-feet of water from 31 wells.

cfs (628 gpm) to run only at times that there is a water shortage in the surface-water outflow, if such a time exists.

Attempting to track or model cumulative depletions over many years becomes problematic due to the dynamics of water movement into and out of the basin. Consider again the water budget shown in figure 2. There are tens and hundreds of thousands of acre-feet of water moving through the system on a monthly basis. It would be problematic to try to discern a depletion on the order calculated by the modeled well field in this setting.

Long term surface-water gauging, ground-water level monitoring, and water-quality data collection can be used to constantly monitor the overall health of the hydrologic system, to make sure that ground-water pumping or other activities are not exceeding the ability of the system to function without detrimental impacts to surface-water flow. Pumping water out of the ground-water reservoir creates storage space that can be refilled during times of high surface water flow. If particular problems in surface-water outflow are identified, there are numerous water management strategies that can be implemented to address such temporal shortages. By addressing these shortages directly, the use of ground water throughout the basin and its tributaries can continue.

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GROUND-WATER USE AND DEVELOPMENT IN THE BITTERROOT WATERSHED

by John LaFave, Associate Research Hydrogeologist

Introduction

The Bitterroot watershed (fig. 1) is one of the fastest growing regions in the closed basin area and has the highest density of wells in the State. It is also a relatively constrained watershed with long-term surface and ground-water monitoring data.

The Bitterroot watershed is a distinct part of the closed basin area with boundaries that correspond closely to those of Ravalli County; it is also an area of concentrated ground-water development. Growth in the Bitterroot Valley has resulted in increased use of ground water, mostly for municipal and domestic purposes. Exploitable ground-water resources occur primarily in the basin-fill deposits. The Bitterroot watershed contains a productive shallow basin-fill aquifer along the floodplains and terraces. Most people in the watershed reside in the valley lowlands between Darby and Missoula, within a few miles of the river. The population growth has resulted in the conversion of much agricultural land to home sites along the valley bottom and upland terraces.

Ground water supplies most of the municipalities and all of the private residences in the watershed. Records from GWIC show a more than fivefold increase in the number of wells (mostly private domestic, or exempt wells) installed in the watershed between 1970 and 2000 (fig. 2). Domestic wells attain a density of about 300 per square mile in two sections in the valley (T. 6 N., R. 20 W., sec. 31 and T. 8 N., R. 20 W., sec. 30), the highest density in the state (fig. 3). About 3 percent of the wells are used for irrigation; most of the wells are for domestic use and exempt from any water rights permitting (93 percent, or nearly 16,000 wells as of 2006).

Development of ground water disrupts the natural flow of water into and out of an aquifer. Ground water removed from the system must be balanced by a decrease in ground-water storage and some combination of increased recharge and/or decreased ground-water discharge. The combination of increased recharge and decreased discharge in response to pumping is termed capture. Usually ground-water discharge (rather than increased recharge) is captured during development.

Significant capture from ground-water development in the Bitterroot Valley should result in a decline in ground-water storage and/or a decrease in Bitterroot River baseflow. Most of the ground-water development has occurred between Darby and Missoula (fig. 3). To assess the impact of ground-water development on stream flow, the difference in annual Bitterroot River baseflows between Darby and Missoula (USGS gauges 12344000 and 12352500, fig. 1) was calculated. Baseflow changes over time were compared to ground-water development, as measured by the cumulative number of wells over time, precipitation data, and changes in ground-water storage.

Baseflows over Time

Long-term stream-flow data are available for the Darby station (1938–present); however, the Missoula records only date back to 1989. Flows on the Clark Fork River upstream and downstream of the Bitterroot's confluence at Missoula have been monitored from 1929 to present and were used to extend the historical record of Bitterroot flows at Missoula (fig. 4). The increase in

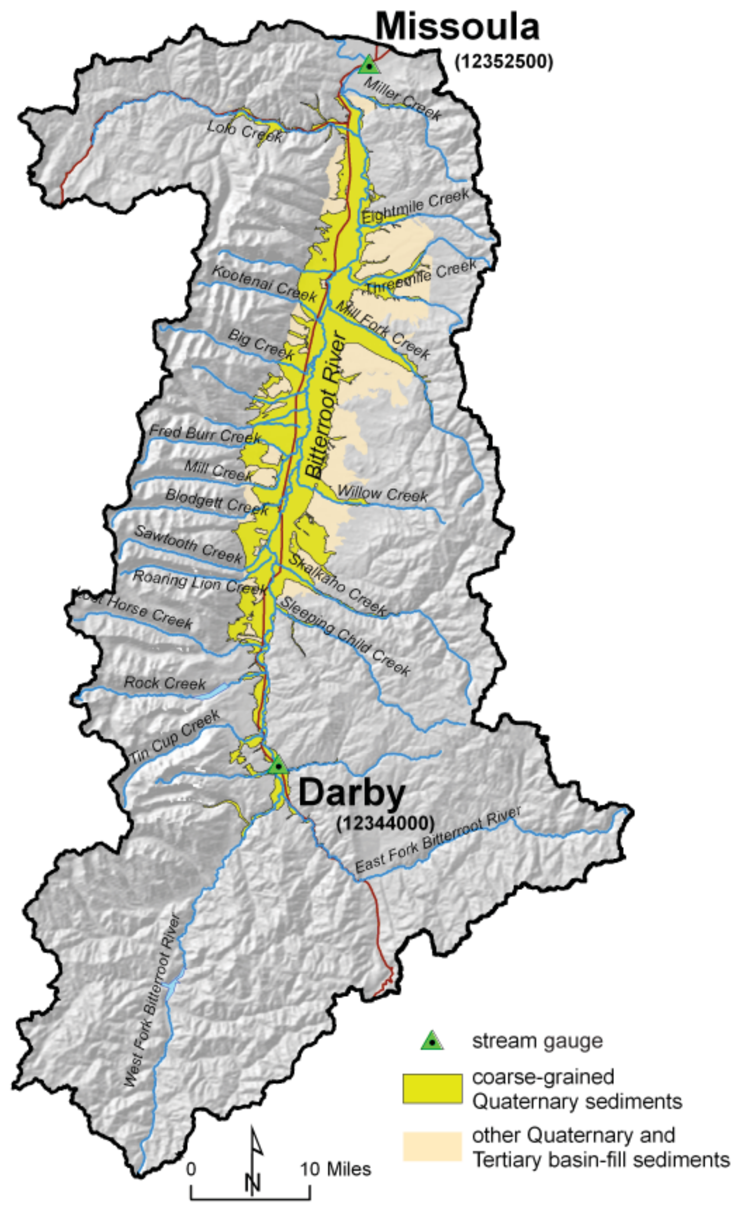


Figure 1. The Bitterroot watershed covers about 2,860 square miles and is gauged at Darby and Missoula.

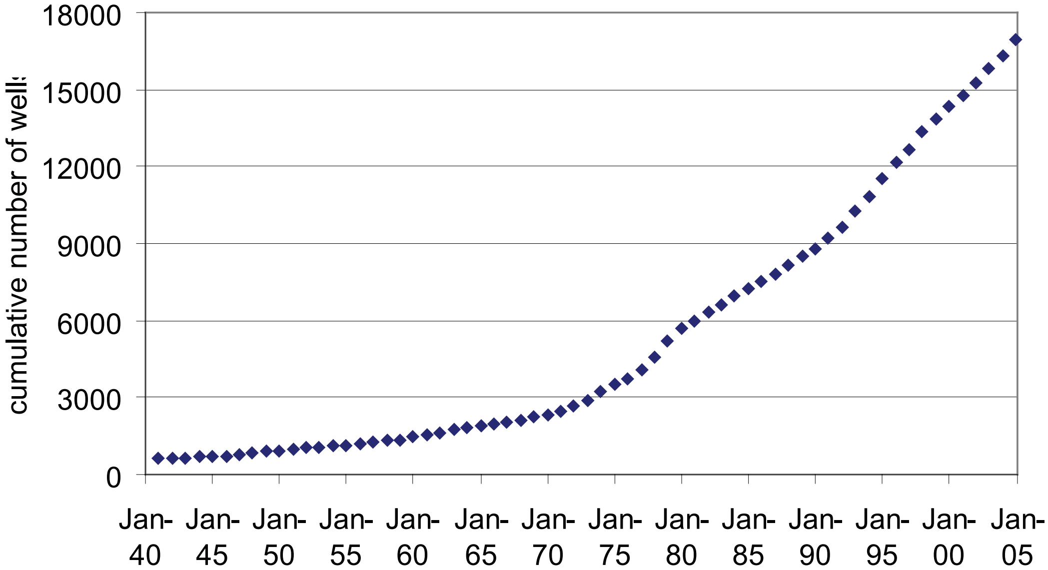
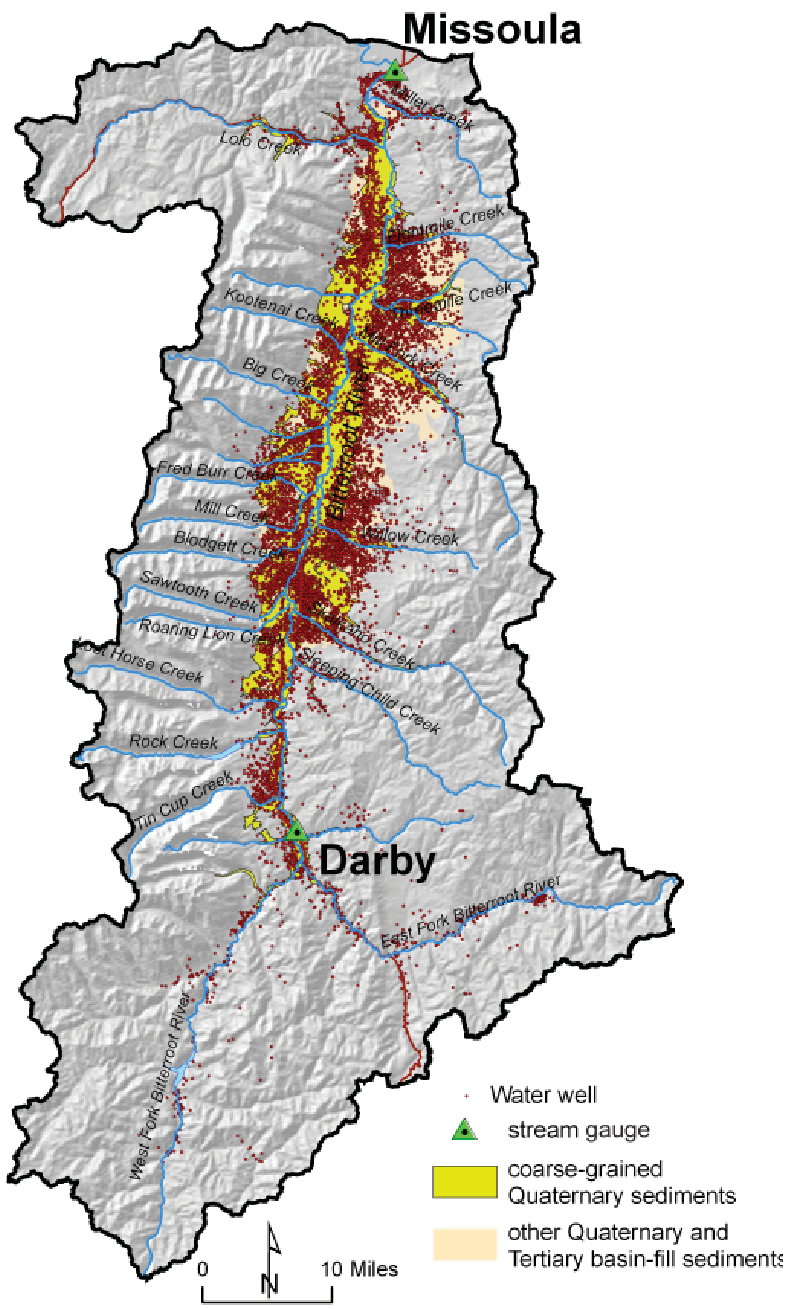


Figure 2. The increasing number of wells in the Bitterroot basin reflects the population growth.

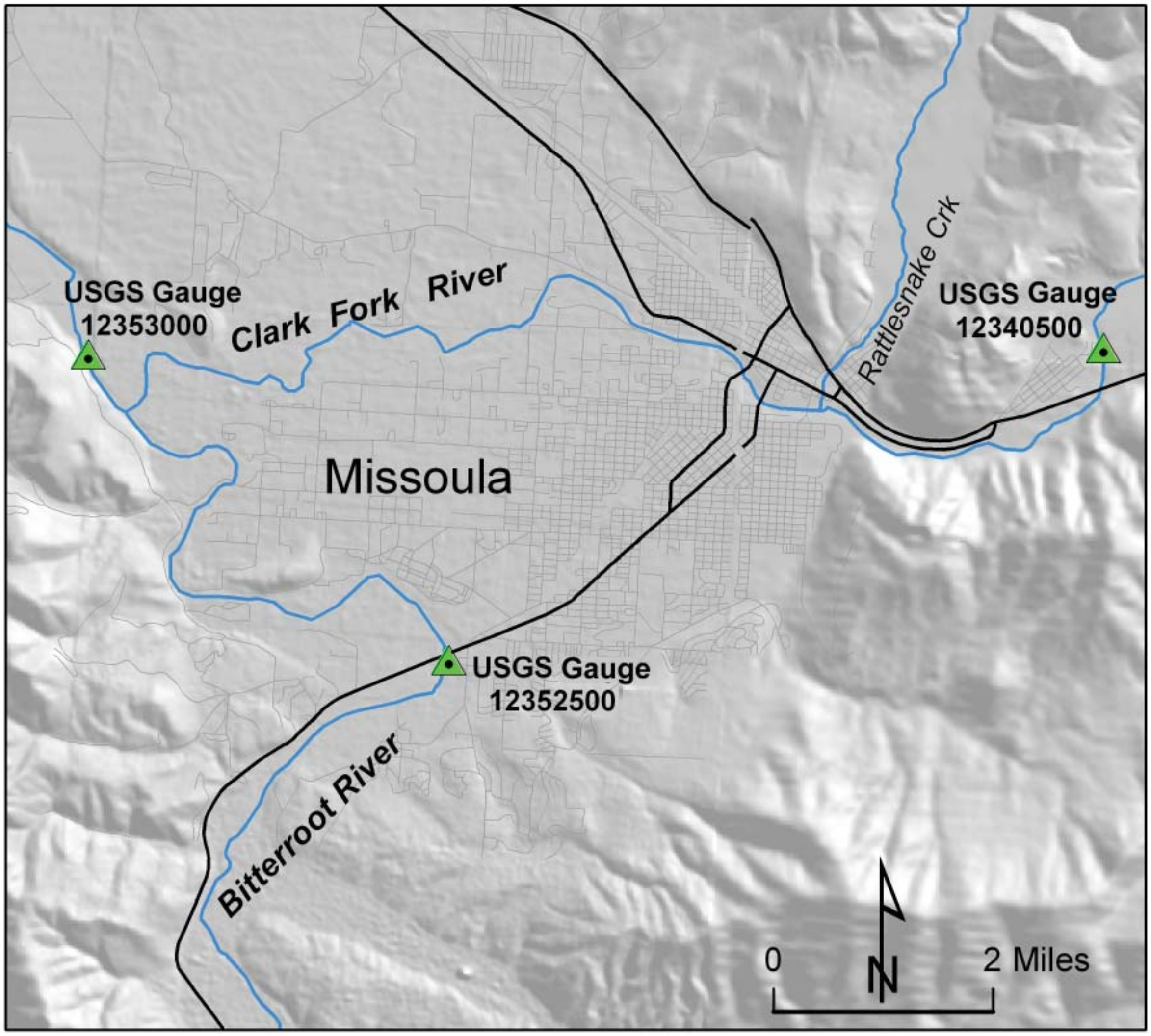
Figure 3. Most of the wells are concentrated between Darby and Missoula reflecting the population distribution.



monthly flows for the Clark Fork was regressed against the Bitterroot River flow data from 1989 to 2005 (fig. 4). Based on 195 measurements, the correlation coefficient (R^2) was 0.996 and the resultant regression equation was $y = 0.9687x$. There is essentially a 1:1 relation between the Bitterroot flows and the increase in flow in the Clark Fork River; about 97 percent of the increase in the Clark Fork River, from above to below Missoula, is due to the inflow from the Bitterroot River (fig. 4). This relationship was used to produce a history of monthly flows for the Bitterroot River at Missoula from 1940 to 1988.

The monthly flow data from 1940 to 2005 for the Bitterroot River at Darby and Missoula were averaged to develop composite annual hydrographs for each station (fig. 5). Hydrographs of the average daily flows in 2000 for each station provide a more detailed picture of the annual stream-flow response (fig. 6). The stream hydrographs show a typical annual response. The rising limb is from March to May in response to increasing flows from snowmelt. Peak flows are in June; the falling recessional limb from July to September marks the onset of drier conditions. The steady low flows during fall and winter months represent periods of baseflow, when ground water contributes most or all of the flow in the stream.

Based on the composite hydrographs, the baseflow period was determined to be between October and January ("baseflow" in fig. 5). For this analysis the monthly flows during that period were averaged for each year to provide a measure of the annual baseflow rate. The baseflow values calculated for Darby, which represent input from the undeveloped headwaters area, were subtracted



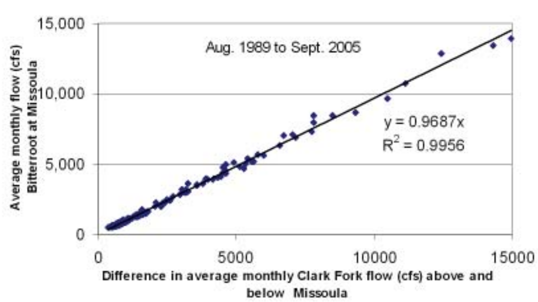


Figure 4. Map showing gauging station locations and a plot of average monthly Bitterroot River flows vs. the difference in flows on the Clark Fork River above and below Missoula.

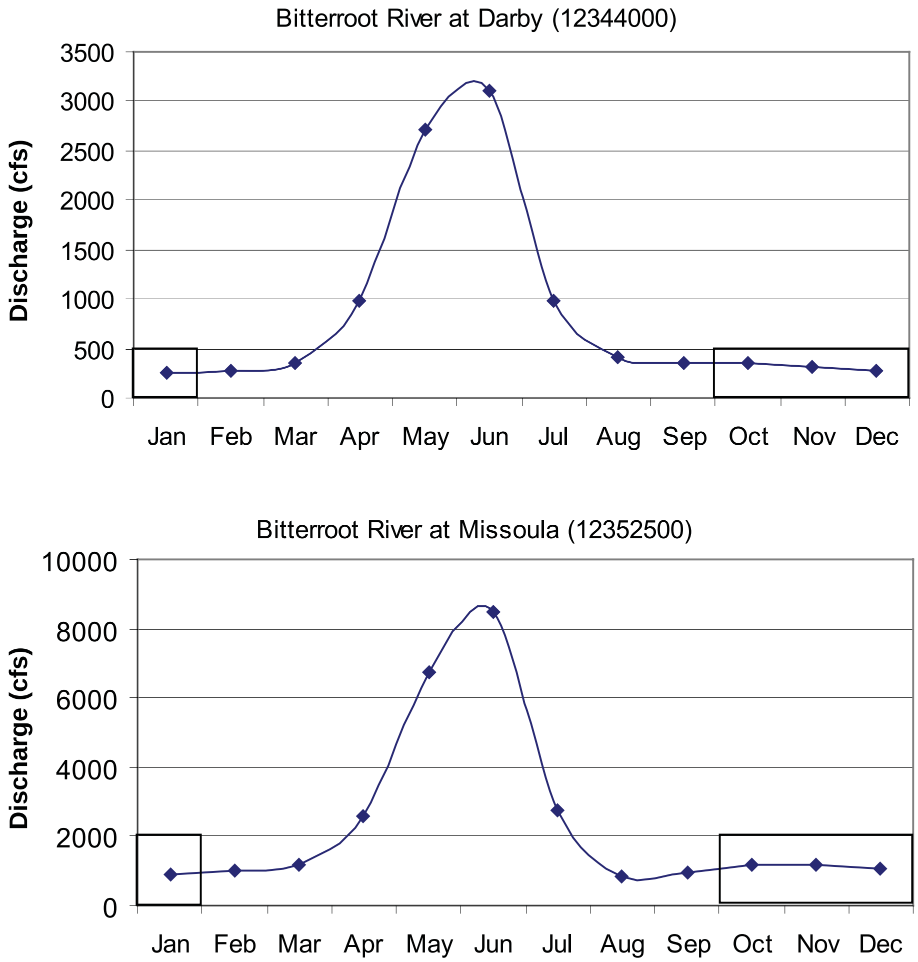


Figure 5. Average Bitterroot River discharge at Darby and Missoula between 1940 and 2005. The shaded area indicates baseflow periods when all or most of the stream flow is sustained by ground-water discharge.

from the values calculated for Missoula. This difference represents the average gain in baseflow in the developed part of the valley, and is also a measure of ground-water discharge between Darby and Missoula.

Figure 7 shows the average Bitterroot River baseflows (average of monthly flow from October to January) from 1940 to 2005, and the difference in baseflows between Darby and Missoula over the same time. The average baseflows at Darby and Missoula and the difference in baseflows show slightly decreasing trends over the period of record; however, there is significant variability from year to year. The variations are generally consistent between the upper and lower parts of the watershed; higher measured baseflows at Darby correspond to higher baseflows at Missoula, below the area of ground-water development.

Baseflows and Well Development

Figure 8 is a plot of the number of wells against the difference in baseflows between Darby and Missoula from 1940 to 2005. If ground-water development is capturing significant volumes of water, then the difference in baseflows should decrease over time. The plot shows a slight negative slope; however, the scatter of the data and the low correlation coefficient (R^2) suggest no significant relationship. Ground-water development, as represented by the number of wells, does not appear to be significantly related to or explain the baseflow variability. Even the period of rapid ground-water development, 1970 to 2005 (fig. 8), does not correspond to an increase in the difference in baseflows over the same time period.

Baseflows and Precipitation

To assess impacts due to climatic variation, changes in baseflow were compared with trends in precipitation data over the same years. Long-term monthly precipitation data (1940–2005) for the western climatic division of Montana (which includes the Bitterroot watershed) were obtained from the Western Regional Climatic Center

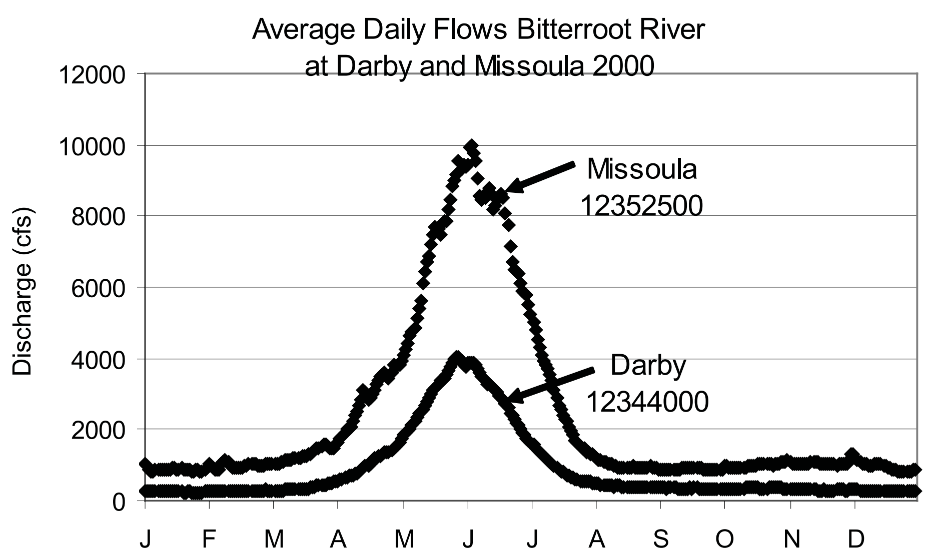


Figure 6. Average daily Bitterroot River flows at Darby and Missoula in 2000.

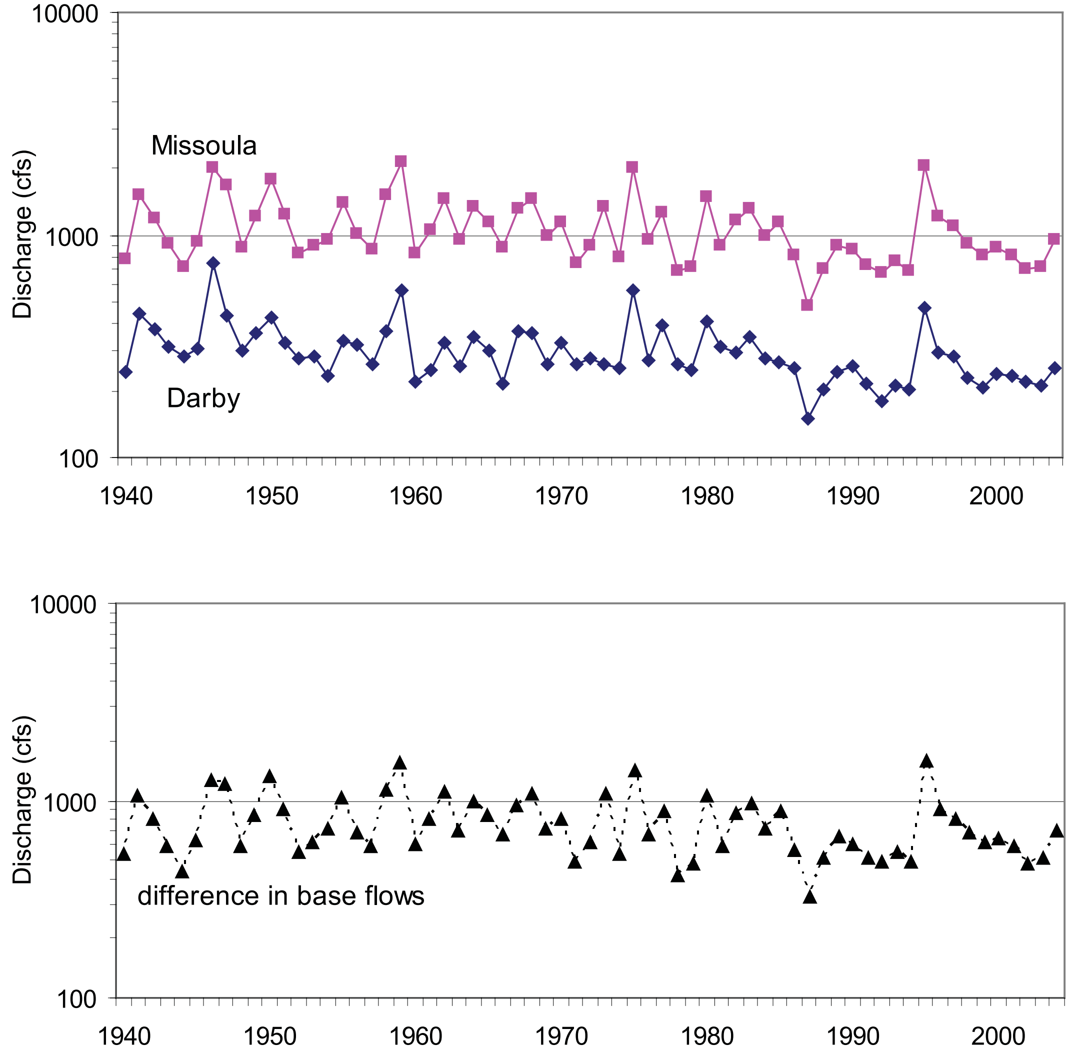


Figure 7. Upper plot shows the average annual baseflow rates at Darby and Missoula; lower plot is the difference in baseflows between Darby and Missoula.

(http://www.wrcc.dri.edu/). The monthly values were summed to obtain annual precipitation totals. Figure 9a shows annual precipitation totals and the difference in baseflows between Darby and Missoula from 1940 to 2005, departures from average of precipitation and baseflows for the same period, and a regression analysis of precipitation against the difference in baseflows. There is a general correspondence between the precipitation highs and lows and the difference in baseflows highs and lows. Both annual precipitation and baseflows show decreasing trends over time. The plot of annual precipitation against the difference in baseflows (fig. 9b) reveals a positive slope and correlation coefficient value of 0.35, indicating that about 35 percent of the variability seen in the annual baseflow data can be explained by annual variation in precipitation.

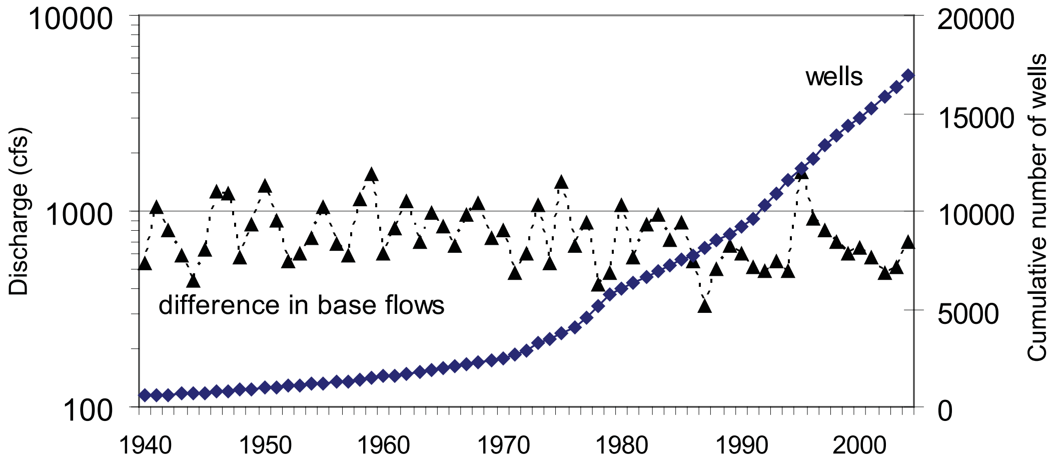
The influence of annual precipitation on the gain in baseflow between Darby and Missoula appears to be more significant than ground-water pumping. Some of the variability seen in the baseflow data is explained by a trend of decreasing precipitation over the last 65 years. The remaining variability cannot be quantified at this scale with the available data and our current understanding of the ground-water–surface-water system.

Ground-Water Storage

Ground-water development can result in declines in ground-water storage. As a first step to assess the significance of the ground-water storage in the Bitterroot watershed, the volume of water in the shallow basin-fill aquifer was estimated by the following method:

Volume = (area of the aquifer) x (saturated thickness) x (Sy, specific yield).

Based on the mapped extent of coarse-grained Quaternary sediments between Darby and Missoula (Smith, 2006), the area of the shallow basin-fill aquifer that is directly connected to the Bitterroot River was calculated to be 375 mi^2 (1.045 x 1010 ft^2) with a thickness of about 50 ft (Smith, 2006b). Ground-water level measurements from 112 wells in the aquifer



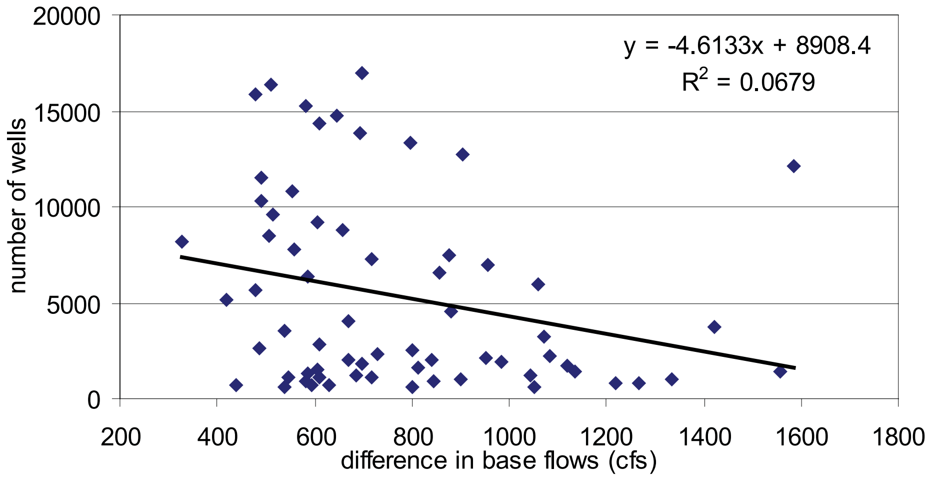


Figure 8. Upper plot shows the average annual difference in baseflows with the cumlative number of wells; the lower plot shows the number of wells vs. difference in baseflows. There appears to be little relation between the variability in baseflow and the number of wells.

ranged from 2 to 43 ft below the land surface with a median value of 10 ft (LaFave, 2006). Therefore, for the purposes of this estimate the saturated thickness was assumed to equal the thickness of the sediments (50 ft) minus the median depth to the water table (10 ft), or 40 ft. The specific yield (Sy) is a unitless factor equal to the volume of water released by a unit volume of an aquifer when drained by gravity. Measured specific yield values for coarse sand range from 0.2 to 0.35 (Johnson, 1967). For the purpose of this estimate, the lower, more conservative value was used.

Given these parameters the estimated volume stored in the shallow basin-fill aquifer is (fig. 10):

$(1.045 \times 1010\ ft^2) \times (40\ ft) \times (0.20) =$
$8.4 \times 1010\ ft^3$ = 625 billion gallons

This estimate does not represent the total amount of recoverable ground water—it is virtually impossible to remove all water from storage with pumping wells. Rather it is meant to highlight the volume of ground water stored in the shallow basin-fill aquifer. During 1940 and 2005 the average gain in baseflow between Darby and Missoula was about 772 cfs, or about 180 billion gallons per year. This volume represents a conservative estimate of annual ground-water discharge from the shallow basin-fill aquifer. Therefore, on an annual basis the volume transmitted through the shallow ground-water system represents about ⅓ of the water stored in the system.

Changes in Ground-Water Storage

Pumping ground water will remove water from aquifer storage. Long-term water-level data were used to evaluate the changes in ground-water storage in the Bitterroot Valley. Ground-water levels in the shallow aquifer have been monitored at three sites (fig. 11), with varying frequency. One of the wells has been monitored

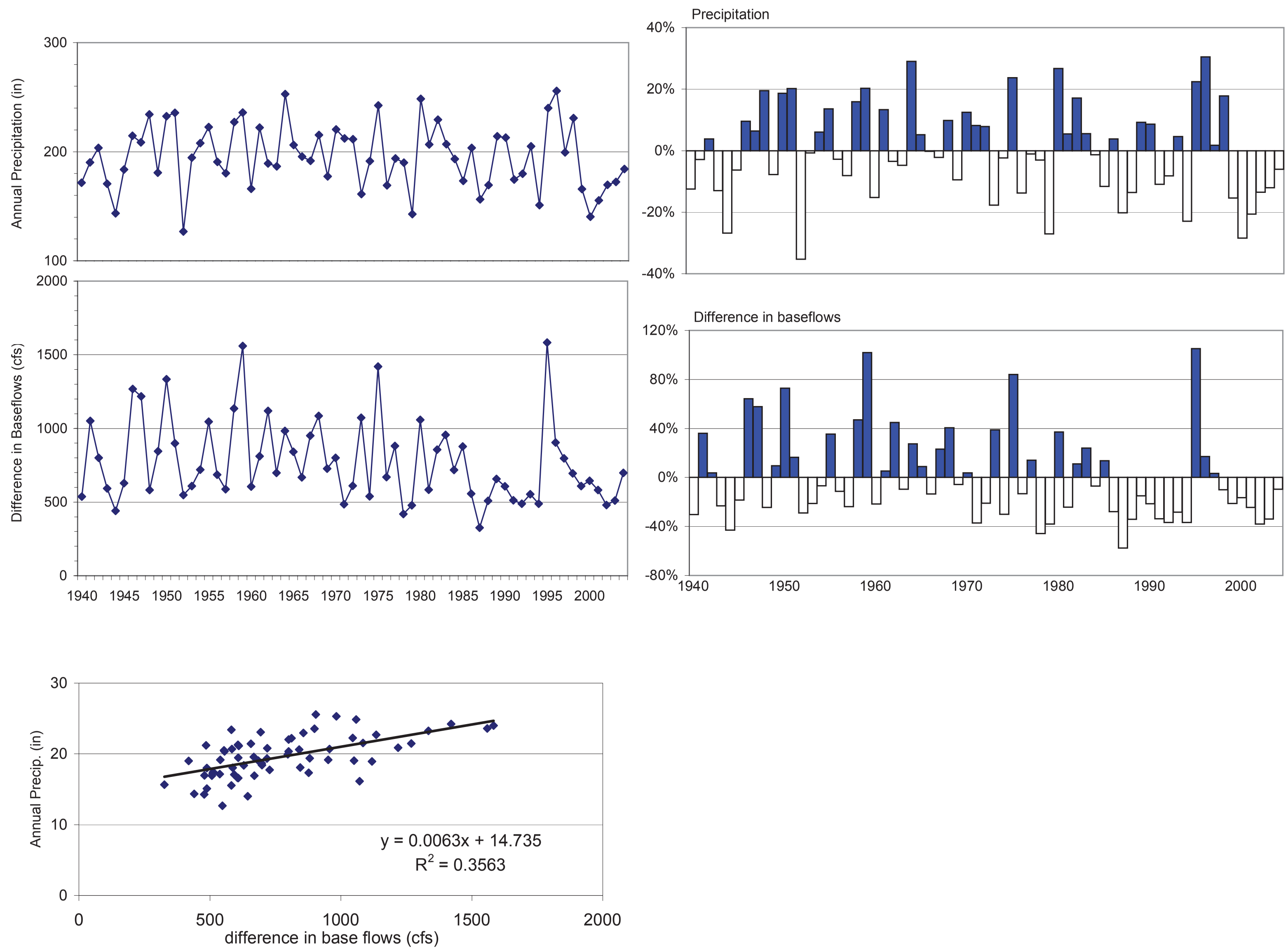


Figure 9. Plots of annual precipitation and difference in baseflows and plots of percent departure from average show correspondence between highs and lows in precipitation and baseflow. The lower plot of precipitation vs. difference in baseflows indicates that precipitation variability is related to baseflow trends.

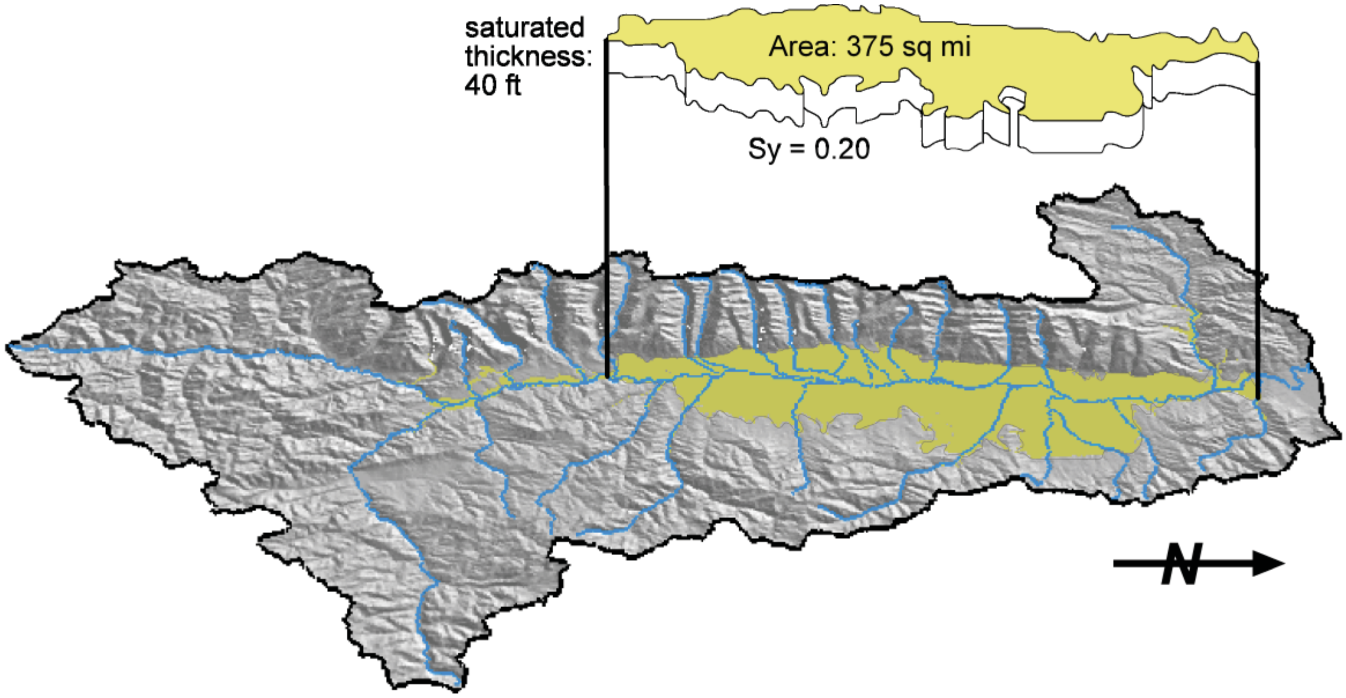


Figure 10. The volume of water stored in the shallow basin-fill aquifer, 625 billion gallons, was estimated by multiplying the area of the aquifer by the saturated thickness, by the specific yield.

since 1970 and the other two since the late 1950s (fig. 11). All three wells are part of the Ground-Water Assessment Program Statewide Monitoring Network; data and hydrographs are available from the GWIC (http://mbmggwic.mtech.edu/). The wells are completed in shallow alluvium, at depths from 40 to 52 ft deep, and the water table is typically within 20 ft of the land surface. For each well, the water-level measurements during a given year were averaged to produce an average annual ground-water level (fig. 12).

Of the three wells only well 136964 shows a slight declining trend (fig. 12); it is the furthest upgradient well in the watershed, and in a part of the aquifer that is heavily influenced by irrigation recharge. There is a slight declining trend in the average annual precipitation for the period from 1960 to 2005. The annual hydrographs for the other two wells show flat or stable trends, indicating no long-term depletions in ground-water storage in the vicinity of these wells.

Seven other wells in the shallow basin-fill aquifer have been monitored since 1993 as part of the statewide monitoring program (fig. 13). Water-level fluctuations in these wells vary with regard to the timing and magnitude of water-level changes, reflecting different recharge sources across the valley. However, the long-term trends are stable, indicating little or no depletion of the shallow basin-fill aquifer.

Ground-Water Withdrawals

The estimated surface-water and ground-water withdrawals in the watershed for the year 2000 totaled about 347 MGD; however, the estimated consumptive use was only 102 MGD. Surface-water withdrawals account for 96 MGD, and ground water 6 MGD of the consumptive use (Cannon and Johnson, 2004) (table 1). Consumptive ground-water use, about 2 billion gallons per year (bgy), represents a small fraction, about 1 percent, of the total annual ground-water discharge (180 bgy) in the Bitterroot Valley (fig. 14).

CONCLUSIONS

Although there are numerous wells in the Bitterroot Valley, ground-water use has not produced measurable impacts, on a basin-wide scale, to Bitterroot River baseflows or ground-water storage. Observed variability in baseflows is more strongly correlated to annual precipitation variability than to ground-water development. Consumptive ground-water use represents a minor fraction (1 percent) of the estimated annual ground-water discharge from the shallow basin-fill aquifer. On a basin-wide scale, there has been little capture due to ground-water development, and long-term water level trends are mostly stable. The declining trend in the average annual water levels noted in well 136964 may be a harbinger of ground-water storage

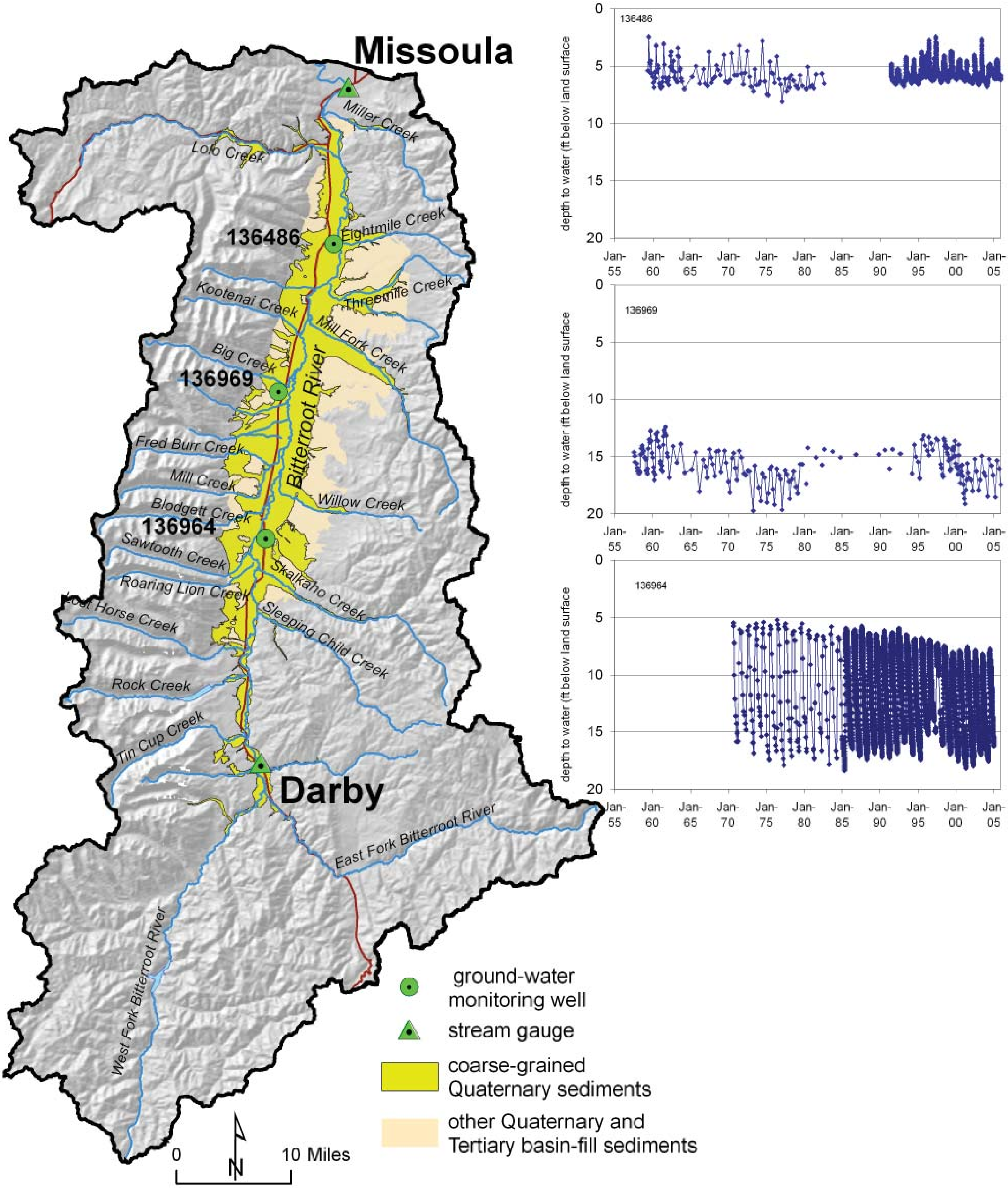


Figure 11. Long-term hydrographs from three wells completed in the shallow basin-fill aquifer.

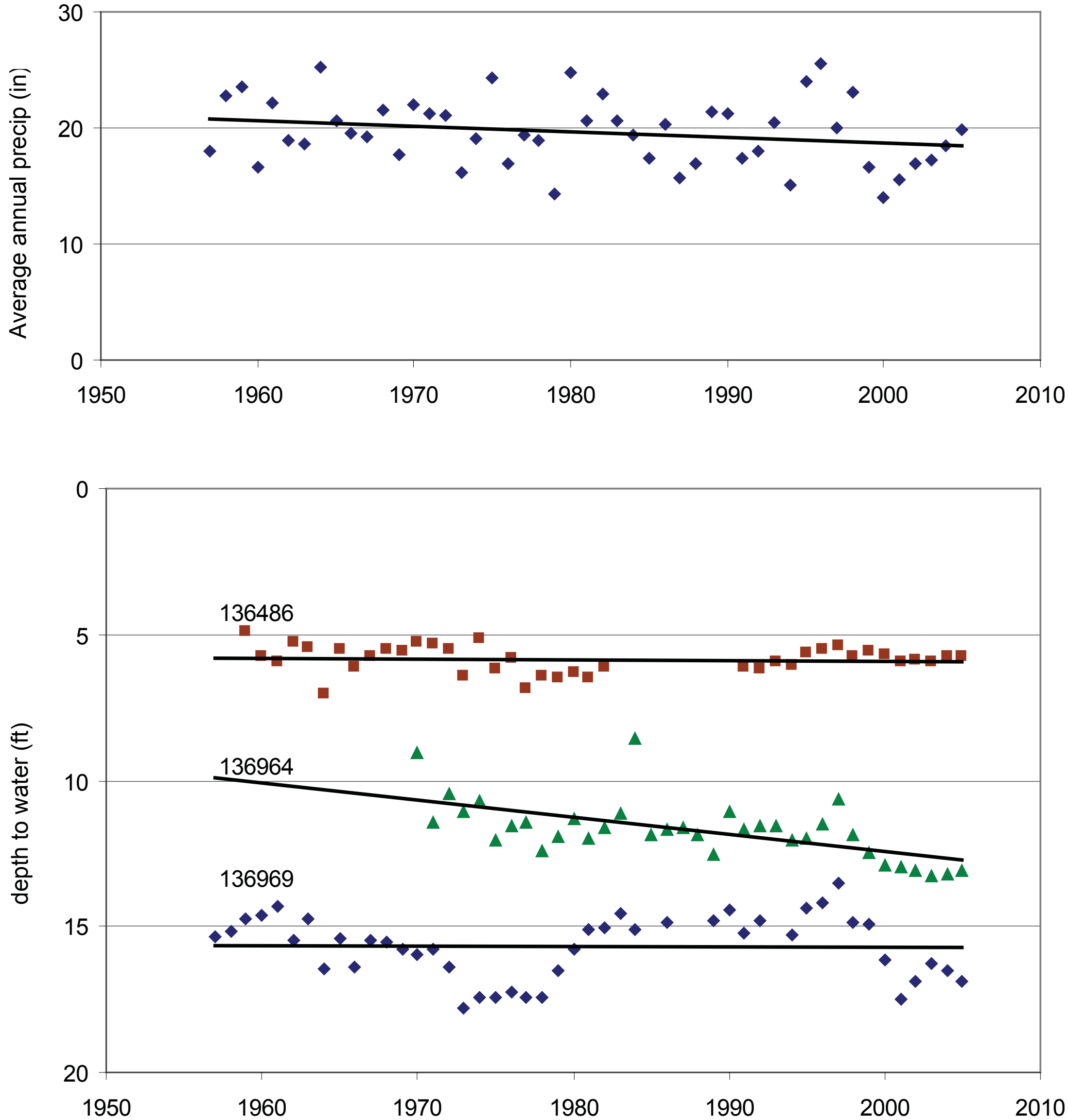


Figure 12. Average annual water-level measurements indicate stable trends for two of the wells. Well 136964 is completed in a part of the aquifer that is strongly influenced by irrigation recharge; the downward trend may be related to land-use change, climate, ground-water pumping or some combination thereof.

Table 1. Estimated ground-water and surface-water withdrawals in the Bitterroot watershed (data from Cannon and Johnson, 2004).

	Ground Water		Surface Water	
	Withdrawals (mgd)	Consumptive Use (mgd)	Withdrawals (mgd)	Consumptive Use (mgd)
Irrigation	6.17	1.73	333.18	95.12
Public water supply	3.01	1.11	0.73	0.27
Self-supplied water	3.03	3.03	0.03	0.03
Industrial	0.12	0.02	0.00	0.00
Stock	0.14	0.14	0.42	0.42

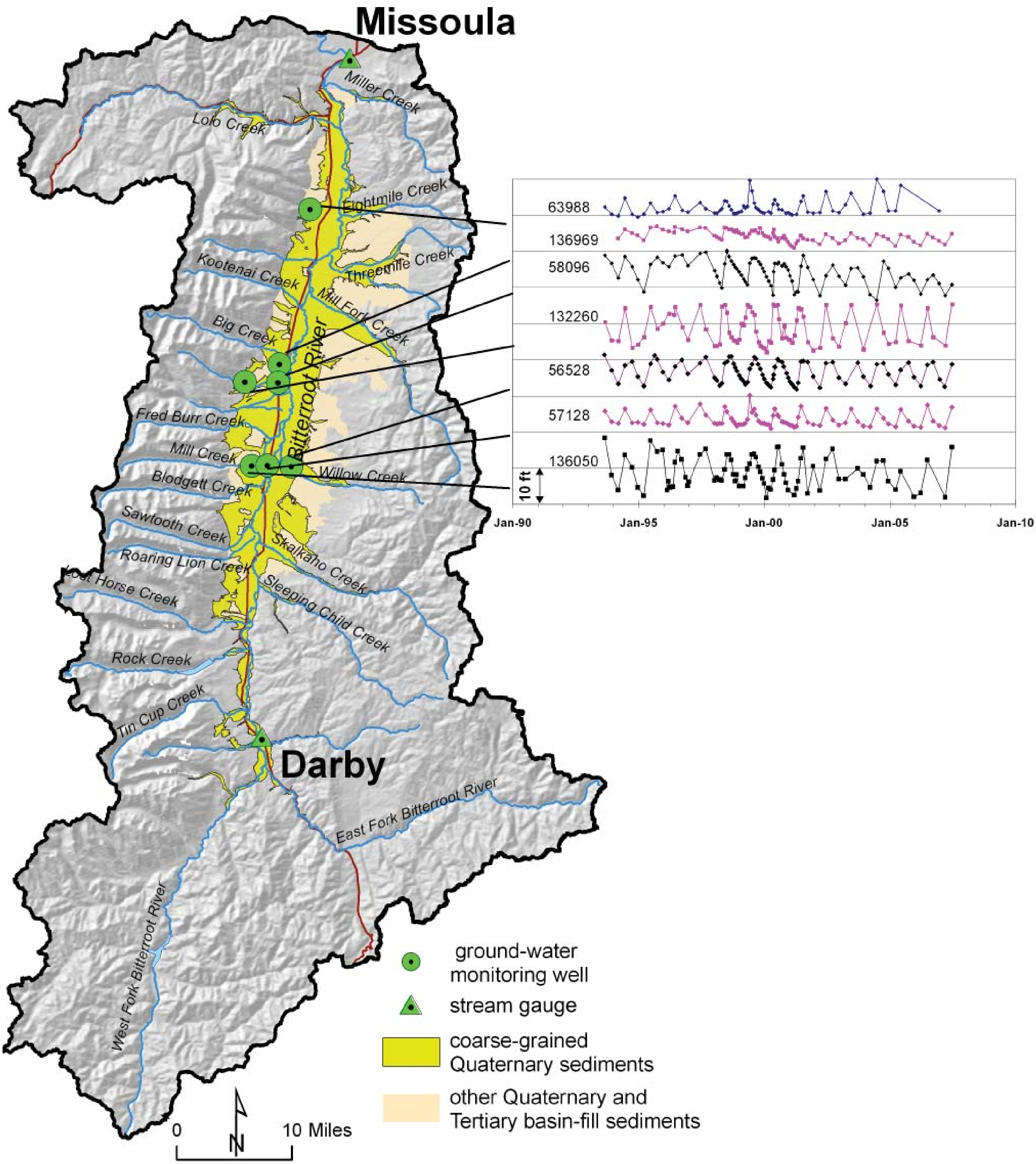


Figure 13. Water-level data obtained since 1993 for additional wells in the shallow basin-fill aquifer show seasonal variability but stable long-term trends.

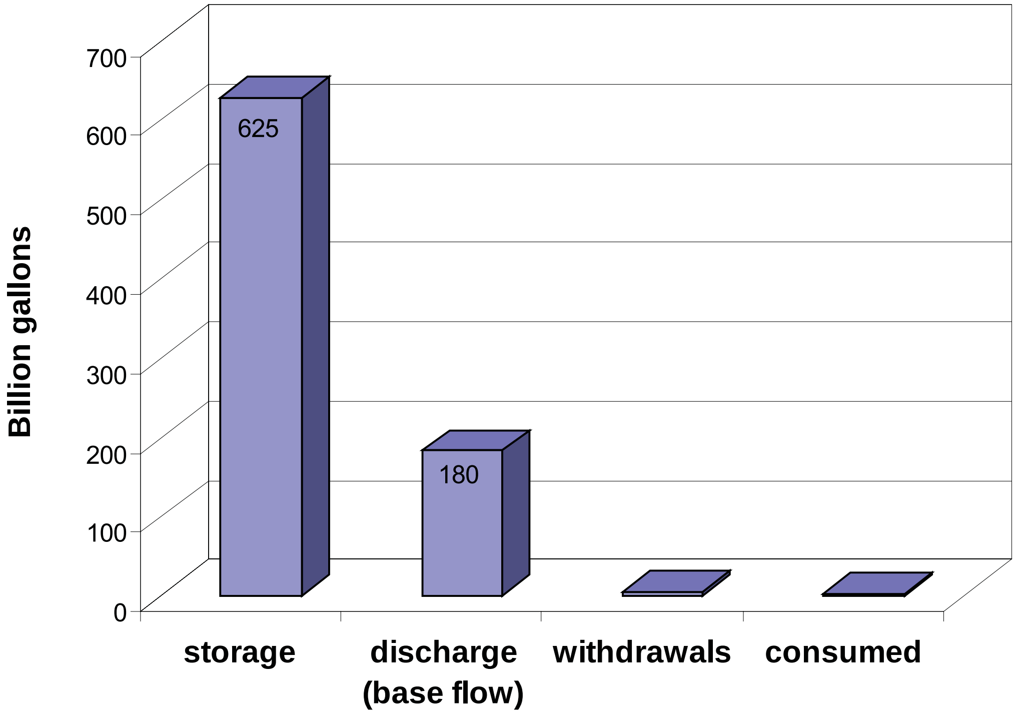


Figure 14. In the Bitterroot Basin ground-water withdrawals represent only a small fraction of the amount of water stored in and flowing through the shallow basin-fill aquifer annually.

depletion, at least locally; however, the trend may simply be reflecting regional climatic (precipitation) variability or land-use changes.

This is a gross water-budget analysis based on a coarse spatial and temporal scale (deep ground-water storage is not accounted for). While it helps provide some perspective with regard to water in the basin relative to water use, it also highlights the importance of spatial scale when assessing the impact of ground-water withdrawals. The results suggest that on a basin-wide scale, that the total use of ground water when compared to recharge, discharge and ground-water storage has had little or no impact. But this analysis does not account for more local effects of ground-water pumping that might impact small, but important stream reaches or other surface-water features such as wetlands or tributary streams.

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SECTION 3: EVALUATION OF HYDROLOGIC ASSESSMENTS

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EXECUTIVE SUMMARY

To satisfy the requirements of HB 831, the Montana Department of Natural Resources and Conservation (DNRC) developed new rules regarding the process for applying for a new ground-water use. This process includes submittal of a hydrogeologic assessment with each new application. A hydrogeologic assessment requires: a description of local conditions, aquifer testing on the source aquifer to develop physical characteristics, use of these characteristics in a net depletion analysis, and an evaluation of physical and legal water availability.

Aquifer testing standards ensure development of credible data at the location of the new appropriation and use. Commonly used methods for net depletion analysis (used to determine adverse affects to senior water uses) contain inherent limitations that may result in inaccurate estimates. Cumulative effects of prior, current, and future ground-water uses are not considered as part of the water-right application process.

DNRC and the MBMG have developed procedures to include hydrogeologic assessment reports and data in the Ground-Water Information Center database, so that this information is made available to the public, as required by HB 831.

REVIEW OF APPLICATION PROCESS AND HYDROGEOLOGIC ASSESSMENT REQUIREMENTS

For a new water-use application to appropriate ground water in closed basins, DNRC requires the following, at a minimum:

1. A completed Application for Beneficial Water Use Permit and the applicable criteria addendum;
2. A hydrogeologic assessment that predicts whether the proposed appropriation will result in net depletion of surface water; and
3. An aquifer recharge plan or other mitigation plan, if required.

Additional data or technical information may be required, at the discretion of DNRC, beyond the minimum hydrogeologic assessment requirements for testing and evaluation. DNRC provides a suggested format for reporting and requires submittal of Form 633 to report aquifer testing details and results.

Components of a Hydrogeologic Assessment

Part 1: Introduction

Background

Description of each proposed use, the purpose of use, period of use, rate and volume of multiple uses, location of diversions and places of use, water-supply calculations, source of water, and topographic location base map with production and monitoring wells identified.

Part 2: Physical Characteristics

Physical Setting

1. Physiography overview: Mountain ranges, rivers, streams, lakes, springs, irrigation canals or drains, and other relevant physiographic information.
2. Geology overview: Descriptions of geomorphology, stratigraphy, and geologic structure, accompanied by a geologic map and geologic cross section(s) of the area of interest.
3. Hydrogeology overview: Description of aquifers (thickness, boundaries, hydrogeologic properties), ground-water levels/hydrographs, ground-water flow direction, and relevant surface-water flow data/hydrographs.
4. Water quality: HB 831 requires a water-right applicant to provide DNRC with a copy of any relevant discharge permit.

Aquifer Testing

Includes information on well drilling, including well logs and well design / construction, aquifer testing methodology, data reduction / analysis, and results / interpretations. This requirement includes either a 72-hour or 24-hour constant-discharge aquifer test for the purpose of determining aquifer characteristics.

Hydrogeologic Assessment

Either a model (numeric or analytical) or hydrogeologic data per MCA 85-2-361 must be included in a hydrogeologic assessment. A model must be developed by a hydrogeologist, a qualified scientist, or a qualified licensed professional engineer. The data or model for the new application must include the following information:

1. The area or estimated area of ground water that will be affected;
2. The geology in the area identified in subsection (1), including stratigraphy and structure;
3. The parameters of the aquifer system within the area identified in subsection (1), to include, at a minimum, estimates for:

(a) the lateral and vertical extent of the aquifer;

(b) whether the aquifer is confined or unconfined;

(c) the effective hydraulic conductivity of the aquifer;

(d) transmissivity and storage coefficient related to the aquifer; and

(e) the estimated flow direction or directions of ground water and the rate of movement;

4. The locations of surface waters within the area described in subsection (1) that are subject to an appropriation right, including but not limited to springs, creeks, streams, or rivers that may or may not show a net depletion;

5. Evidence of water availability; and

6. The locations of all wells or other sources of ground water of record within the area identified in subsection (1). See MCA 85-2-361.

Part 3: HB 831—Net Depletion Evaluation

A net depletion analysis must be submitted with the water-right application and must include, but is not limited to, analysis of the following factors within the affected area:

1. Evidence of the nature or the hydraulic connection between the source aquifer and all surface water;

2. Evidence of propagation of the drawdown from pumping a proposed well and volume, rate, timing, and location of any resulting surface-water effects;

3. Comparison of the proposed flow rate and period of diversion to similar types of existing water uses;

4. Estimates of the volume of water consumed monthly by a proposed project through evaporation, evapotranspiration, and all other forms of consumption associated with the proposed project;

5. An assessment of potential return flows to a source aquifer or surface-water source and the volume, rate, timing, and location of return flows;

6. Return flow—includes but is not limited to any treated wastewater if the treated wastewater will be used as part of an aquifer recharge plan;

7. The volume, rate, timing, and locations of accretions to surface water that are not consumed and subsequently return to surface water;

8. A water balance table that describes the monthly and total annual water balance for the proposal; and

9. A list and map of the points of diversion of surface-water appropriation rights and ground-water rights on record with the department that are located within the potentially affected area.

Information required by the hydrogeologic assessment may not be sufficient to meet applicable criteria under MCA 85-2-311, including but not limited to adverse effect to a prior appropriator. The applicant for a beneficial water use permit pursuant to MCA 85-2-311 is responsible for providing sufficient evidence to meet all applicable criteria (see Permit Application Criteria Addendum).

Part 4: HB 831—Net Depletion and Adverse Affect

An applicant must analyze whether any net depletion to surface water will result in an adverse effect on a prior appropriator. If the applicant determines that no adverse affect will occur as a result of the net depletion analysis, DNRC will proceed to process the application. If the applicant determines that an adverse effect will occur, the applicant must provide a mitigation or aquifer recharge plan that will offset the net depletion. DNRC will be proposing rules to describe net depletion requirements; however, adverse effect created as a result of net depletion will be determined on a case-by-case basis.

Part 5: HB 831—Water Quality

A hydrogeologic assessment must include any predicted water-quality changes to surface water within the potentially affected area that may result and the following information, as required by MCA 85-2-361(2)(b):

1. The location of any existing documented hazards (generally—geological, hydrological, or human-caused) that could be affected or exacerbated by the proposed appropriation, such as subsidence, along with a plan to mitigate any conditions or impacts;

2. Water-quality information to comply with MCA 85-2-361(2)(b)(ii) and HB 831, Section 19. The applicant must provide the department with a copy of a relevant discharge permit from the Department of Environmental Quality (DEQ) if the "aquifer recharge" (i.e., mitigation) plan involves use of sewage. Submit a copy of the discharge permit with the water right application form; and

3. A description of any water treatment method that will be used at the time of any type of

injection or introduction of water to the aquifer to ensure compliance with 75-5-410 and 85-2-364, and the water-quality laws under Title 75, chapter 5. See MCA 85-2-361(2)(b)(iii). Enclose a copy of the water treatment method used with the water-right application form.

Part 6 : Permit Application Criteria Addendum

Information required by the hydrogeologic assessment may not be sufficient to meet applicable criteria under MCA 85-2-311, including but not limited to adverse effect to a prior appropriator.

The following water-right HB 831 Ground Water Permit Application Information criteria must be addressed as required by existing rules for a correct and complete application. Address information pertaining to the criteria in a separate document. If information in the hydrogeologic assessment pertains to the water-right criteria, repeat that information in the criteria document. DNRC will not search through an application for information pertaining to the water-right criteria.

ARM 36.12.1703, Permit Criteria:
Physical Ground-Water Availability

1. Applicants for ground water must provide substantial credible information demonstrating that water is available for their use from the source aquifer in the amount the applicant seeks to appropriate during the proposed period of diversion.
2. Information demonstrating physical ground-water availability must include an evaluation of drawdown in the applicant's production well for the maximum pumping rate and total volume requested in the permit application.
3. The drawdown projected for the proposed period of diversion must be compared to the height of the water column above the pump in the proposed production well to determine if the requested appropriation can be sustained.
4. The requirements of ARM 36.12.121 must be followed.

ARM 36.12.1704, Permit Criteria:
Existing Legal Demands

1. A proposed water-right application may affect prior appropriations and water reservations. These existing legal demands will be senior to a new application and the senior rights must not be adversely affected.
2. The applicant must identify the existing legal demands on the source of supply and those waters to which it is tributary and which the applicant determines may be affected by the proposed appropriation.
3. The applicant must provide an abstract of those water rights identified.

ARM 36.12.1705, Permit Criteria:
Comparison of Physical Water Availability and Existing Legal Demands

1. To determine if water is legally available, the applicant must compare the physical water supply at the proposed point of diversion and the legal demands within the area of potential impact. An applicant must become familiar with senior water rights operations to accurately evaluate the effect to the senior water right.
2. Applicants must analyze the senior water rights on a source of supply and those waters to which it is tributary within the area of potential impact. Provide a written narrative comparing the physical water supply at the point of diversion during the period of diversion requested and the legal demands that exist for the water supply during that same period.
3. If known patterns of use differ from the legal water rights filings, an explanation may be submitted addressing the current water use operation. For example, if a water reservation has not been perfected, that information may help to explain if water is legally available.

ARM 36.12.1706, and HB 831 Permit Criteria:
Adverse Effect

1. Adverse effect for permit applications is based on the applicant's plan showing the diversion and use of water and operation of the proposed project can be implemented and properly regulated during times of water shortage so that the water rights of prior appropriators will be satisfied.
2. A written narrative must be provided addressing the potential adverse effect to the water rights identified in ARM 36.12.1704.
3. For ground-water applications, in addition to (1) and (2), the applicant shall describe how water levels in wells of prior water rights will be lowered and the rate and timing of depletions from hydraulically connected surface waters.

Part 7 - Mitigation and Aquifer Recharge Plans

If the amount of net depletion predicted will result in adverse effect, a mitigation plan or an aquifer recharge plan is required. The plan must show how the mitigation or aquifer recharge plan will offset the adverse effect to a senior surface-water right.

REVIEW OF CURRENT PERMIT APPLICATIONS AND ASSOCIATED HYDROGEOLOGIC ASSESSMENTS

To date the MBMG has received nine hydrogeologic assessments and associated aquifer test data from DNRC for inclusion in the Ground-Water Information Center (GWIC) database. These assessments accompany ground-water applications from locations within closed basins that were submitted to DNRC. Discussions with DNRC resulted in the following changes to the current process:

1. Several changes/clarifications for Form 633, required by applicants to submit aquifer test data to DNRC in support of a water-use application.
2. Agreement between DNRC and MBMG that hydrogeologic assessments will be transmitted to MBMG for inclusion in the GWIC database when DNRC has determined that the assessment is adequate for purposes of the application.
3. MBMG will store and make available to the public (through GWIC) the assessment information only. Other information related to the water-use application that may be provided to DNRC will not be included.
4. MBMG will assign a unique identifier for each test that will be related to the GWIC ID number assigned to each well used in the test.

Analytical Methods Used for Net Depletion Analysis

Based on a review of ground-water assessments received by the MBMG, most applicants use an analytical method of depletion analysis (Colorado Method, Well Pumping Depletion Model). These offer the advantage of simple, somewhat standardized methodologies that are likely available to all applicants. Analytical methods are simplifications of the physical systems being evaluated. They involve inherent assumptions that can limit their applicability, such as:

- Homogeneous/isotropic aquifer; and
- Horizontal ground water flow only.

Analytical methods typically generate a zone-of-influence or area-of-impact circle of fixed radius that is shown on a map. The circles do not account for natural and widespread variability in hydrologic systems. This inherent inaccuracy may or may not be important in a given area, and the long-term effect (over- or underestimating actual depletion) on the accumulation of net depletion analyses is unknown.

Numerical Modeling Used for Net Depletion Analysis

Numerical modeling (such as MODFLOW) can be used for net depletion analysis to account for more complexity in the hydrologic system. Numeric modeling also involves the use of simplifying assumptions, but assumptions can be reduced by including some of the complexity of a natural system into the model.

Numeric modeling generally has more demanding data requirements, including information on conditions some distance from the project site. It is generally more time-intensive and requires a higher level of technical expertise. However (again, generally), numeric modeling can provide more effective use of available data and a greater accuracy of solution to specific site conditions.

Conclusions of the Review of Hydrogeologic Assessments

The aquifer testing requirements provide useful, site-specific information for the project area subject to a new ground-water use. The net depletion analysis process is a standardized approach that uses this information to provide estimates of the effects of the new use.

New uses in closed basins are evaluated by including new, permitted uses in each subsequent application's hydrologic assessment. Since each new use results in an increase in depletion, it follows that at some point, the maximum allowable depletion will be reached and no further appropriations should be allowed. However, the cumulative hydrologic effects of prior, current, and potential future uses within a closed basin are not considered in an individual assessment.

The current process is focused, as the law requires, on local impacts to senior water uses. It is possible that overall management of the ground-water resource in closed basins is insufficient to prevent future over-use.