PRINCIPAL AQUIFERS OF MONTANA

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INTRODUCTION

Often called the “hidden resource,” groundwater is one of Montana’s greatest natural assets. In most rural areas, groundwater supplies all the domestic, stock, and ranch needs. Groundwater is the municipal source for some of Montana’s larger cities, including Missoula, Kalispell, and Sidney. High-capacity wells support irrigated agriculture in north-central and northeastern Montana (eastern Sheridan and northern Blaine Counties), and western Montana (the Flathead, Beaverhead, and Gallatin River Valleys). Groundwater also plays a crucial role in sustaining streamflow; about half of the total annual flow in typical Montana streams is derived from groundwater (Wolock, 2003).

Groundwater occurs in aquifers; these are permeable geologic units that store and transmit usable quantities of groundwater, be it to a well, a spring, or as baseflow to a river. Aquifers provide two important functions in the context of the hydrologic cycle: (1) to transmit water through the subsurface from areas of recharge to areas of discharge, and (2) to provide groundwater storage. The characteristics of an aquifer—for example, its productivity, or its baseline water quality—are largely controlled by geology. Understanding Montana’s geology is key to understanding the State’s groundwater resources. The principal aquifers of Montana, here defined as an aquifer or aquifer system able to yield potable water across broad areas, were delineated based on digital data derived from the 1:500,000-scale geologic map by Vuke and others (2007). The map “Principal Aquifers of Montana” (Crowley and others, 2017) shows the surface extent and boundaries of aquifers based on geologic groupings (fig. 1).

This paper will review basic concepts of groundwater, provide an overview of Montana’s principal aquifers, and summarize information regarding water quality and wells completed in the principal aquifers. Water-quality and well data were obtained from the Montana Ground Water Information Center (GWIC), the State’s official repository for water-well logs, groundwater chemical analyses, and other critical groundwater data (http://mbmggwic.mtech.edu/).

Occurrence of Groundwater

Groundwater occurs under either unconfined (water table) or confined (artesian) conditions (fig. 2). In unconfined aquifers the water table represents the upper boundary of the aquifer; pore spaces are fully saturated below the water table. In the vadose zone, the area above the water table, pore spaces are filled with air and water. The water table moves upward or downward in response to water recharging (entering) or discharging 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 within 100 ft of the land surface and occur adjacent to the major streams in Montana; many unconfined alluvial aquifers have direct hydraulic connection to surface water. Water table conditions may also exist in bedrock aquifers, where bedrock crops out or is close to the land surface.

Confined, or artesian, aquifers are permeable geologic units that are: (1) completely saturated and (2) overlain or “capped” by aquitards—relatively low-permeability layers such as clay, silt, or shale—that restrict groundwater flow (confining layers). Groundwater in confined aquifers occurs under pressure, and the water level in a well completed in a confined aquifer will rise above the top of the aquifer. The level to which water will rise in a confined aquifer is referred to as the potentiometric surface. In flowing artesian wells, the pressure head in the aquifer is sufficient to raise the water level above the land surface. As opposed to unconfined aquifers (which release water

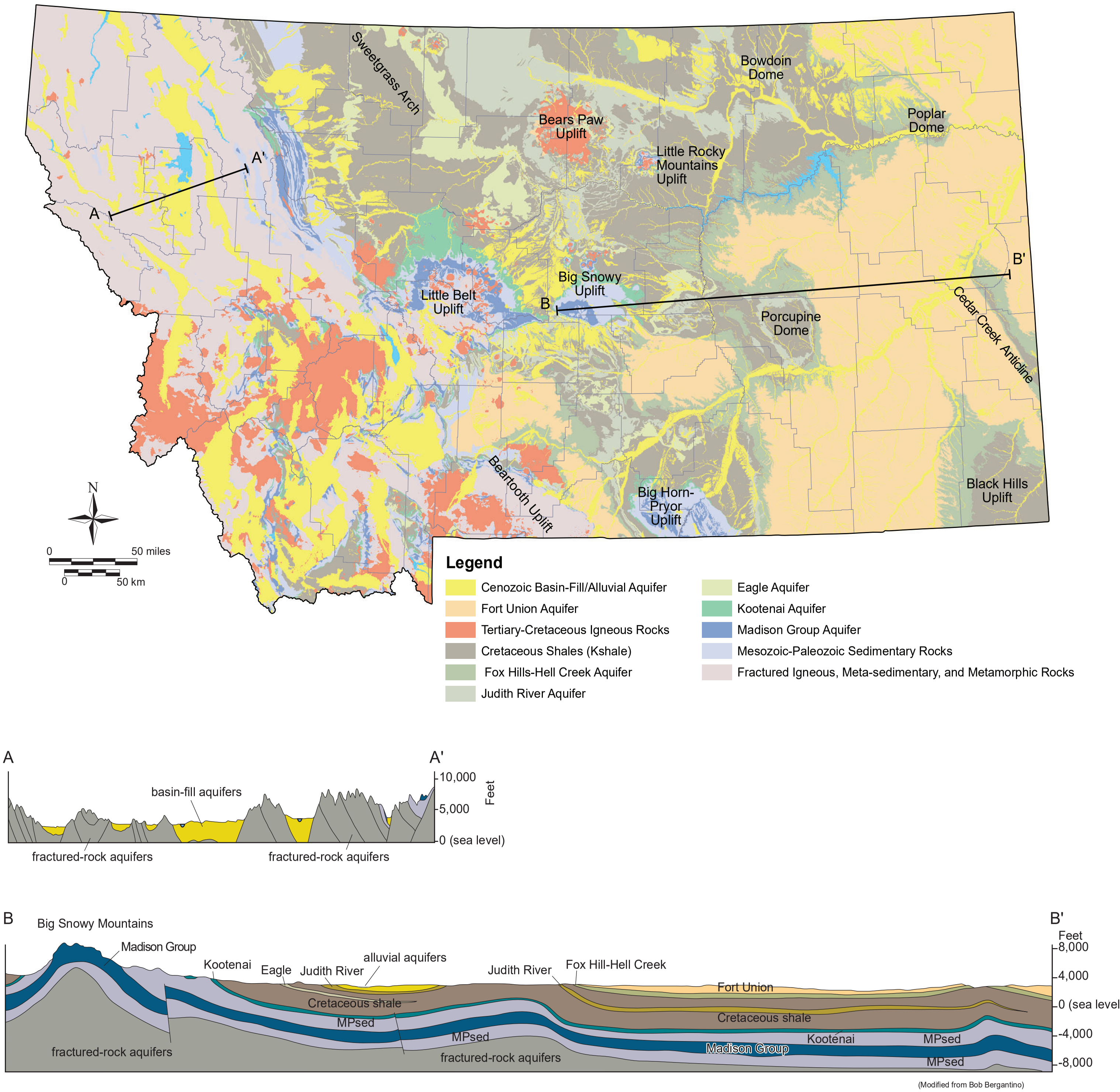


Figure 1. Surface distribution of principal aquifers in Montana. Cross section A: Generalized cross section illustrating principal aquifers through western Montana Intermontane Basins (A–A'). Cross section B: Generalized cross section illustrating principal aquifers through eastern Montana Great Plains (B–B').

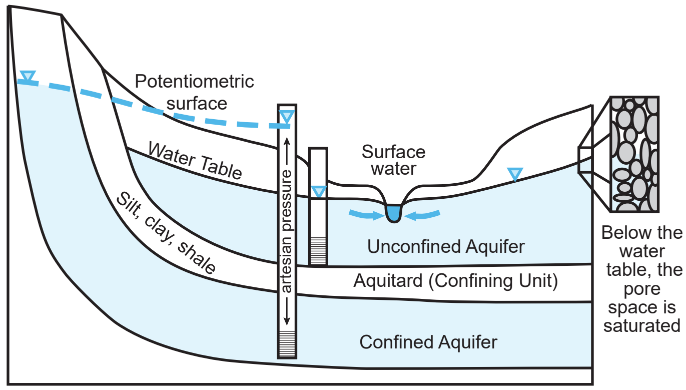


Figure 2. Groundwater occurs in unconfined and confined aquifers. The water table is the upper surface of an unconfined aquifer. Confined aquifers are buried below less permeable layers and the water is under pressure.

from storage by dewatering of pore space), confined aquifers yield water to wells by compression of the aquifer material and expansion of the water. This is an important distinction, because for a given amount of head loss, unconfined aquifers yield significantly more water to wells than confined aquifers. Thus, pumping from confined aquifers generally leads to larger declines in water levels over broader areas than pumping similar volumes of water from unconfined aquifers.

Groundwater flows through aquifers from recharge areas towards discharge points such as rivers, wetlands, springs, and lakes. A groundwater 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 continually added to the system by recharge, and leaves the system by discharge to surface water and evapotranspiration. Groundwater moves much more slowly than surface water. Water in a stream may flow on the order of 1 ft/sec, whereas water in a highly permeable unconfined aquifer might flow on the order of 1 ft/d. In unfractured, deeply buried confined aquifers, the flow rates can be on the order of 1 ft/yr.

Groundwater levels vary in response to changes in recharge, discharge, and aquifer storage. If recharge exceeds discharge, water levels rise as more water is stored in the aquifer. When discharge exceeds recharge, water levels decline as groundwater is lost from storage. Typically, water levels are higher in the spring, reflecting recharge from spring runoff. Water levels tend to decrease during winter months when the ground is frozen, preventing recharge. When recharge and discharge are balanced over the long term, steady-state conditions exist; water levels may fluctuate seasonally, but long-term averages remain stable. Climatic, land-use, or water-use changes (such as groundwater development) can disrupt steady-state conditions so that water levels either rise or fall over the long term.

Stream–Aquifer Interaction

Groundwater and surface water are considered a single resource because nearly all surface-water features (streams, lakes, springs, wetlands, and reservoirs) interact with the underlying groundwater system (Winter and others, 1998). The interactions of surface water and groundwater are governed by: (1) the position of the surface-water bodies relative to the groundwater 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 groundwater flows from the underlying aquifer into the stream (fig. 3). Groundwater discharge into a stream (baseflow) can account for more than 50 percent of annual streamflow, and typically accounts for all streamflow during winter months when the ground is frozen (fig. 4).

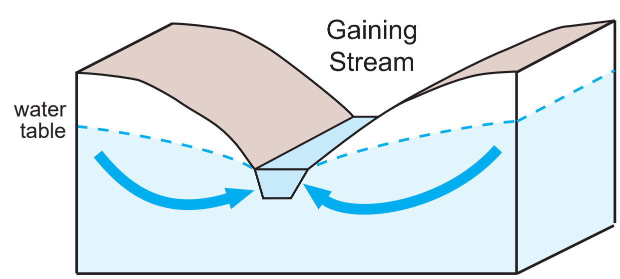


Figure 3. Shallow aquifers discharge water to gaining streams.

Not all streams are gaining; some lose water through the streambed to the underlying aquifer. In losing streams (or stream reaches), the stream stage (height of water in the stream) must be higher than the underlying/adjacent water table (fig. 5). Losing

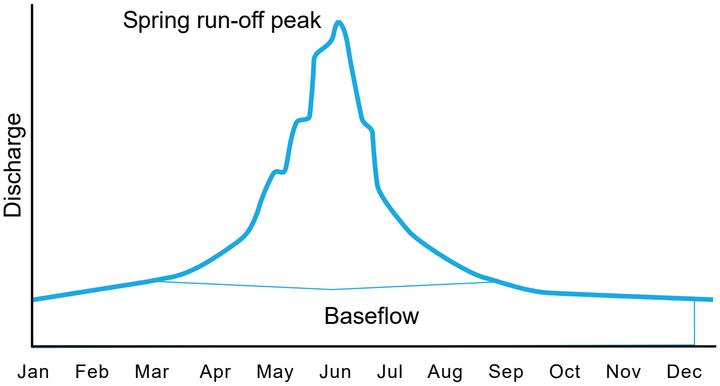


Figure 4. Schematic seasonal hydrograph showing stream discharge against time. Baseflow, the discharge of groundwater into a stream, is a major component of total annual streamflow; during dry and winter months streamflow is sustained by baseflow.

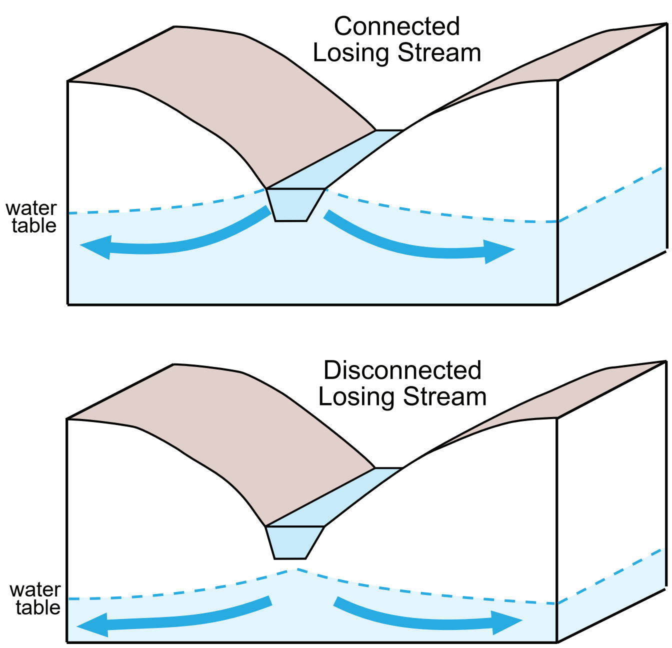


Figure 5. Losing streams lose water to the underlying aquifer; disconnected losing streams are separated from the underlying aquifer by an unsaturated zone.

streams and leaky irrigation canals are an important source of groundwater recharge in Montana.

Seasonal runoff and precipitation events can alter groundwater/surface-water interactions. During spring runoff or times of high flows, stream stage can rise above the adjacent water table, resulting in groundwater recharge. After the high flows decrease, groundwater stored in the streambank may gradually discharge back into the stream.

Climate variability can also affect stream–aquifer interactions. Prolonged drought can reduce the amount of groundwater recharge, resulting in significant storage declines. These conditions can reduce baseflow to streams, resulting in streamflow reductions and, in extreme cases, drying up of stream reaches.

Groundwater development (pumping) can decrease the volume of baseflow by intercepting groundwater that would otherwise discharge to a stream. Pumping may also initiate or increase induced stream leakage to the aquifer. In the idealized stream–aquifer system shown in figure 6, groundwater development in a surficial aquifer directly connected to a gaining stream will progressively reduce baseflow. If development (pumping) continues to the point that the groundwater level adjacent to the stream falls below the stream level, the stream becomes losing, and water is induced from the stream into the aquifer (induced recharge). Figure 6 illustrates the transition from a gaining to a losing stream in response to prolonged groundwater withdrawals. The time lag between the start of pumping and any reduction in streamflow depends upon the distance and depth of the pumping well relative to the stream, the hydraulic characteristics of the aquifer, and the pumping rate. Furthermore, the effect of groundwater pumping on streamflow may persist long after pumping has stopped.

Groundwater in Montana

In Montana, more than 220,000 known wells withdraw water for domestic, stock, industrial/commercial, irrigation, and public water supply uses. Most wells (93%) provide domestic or stock water; irrigation, public water supply, and industrial wells account for the other 7 percent (GWIC database). Montana's water wells withdraw about 285 million gallons of groundwater per day (Maupin and others, 2014). Domestic and stockwater use is volumetrically the smallest, accounting for about 12 percent of withdrawals. Irrigation, public water supply, and industrial wells account for 88 percent of annual withdrawals (fig. 7).

Montana's groundwater resources are closely tied to the geology of the State's two major physiographic provinces: the Intermontane Basins of the northern Rocky Mountains, and the northern Great Plains, glaciated and unglaciated (fig. 8). The Intermontane Basin region covers the western third of the State. It is characterized by several distinct north-trending mountain ranges separated by valleys (intermontane basins). The valleys (basins) contain through-flowing drainages within their valley deposits, and are typically flanked by floodplains and alluvial terraces. The headwaters of Montana's major drainage systems are in the Intermontane Basin province where the Continental Divide separates rivers of the Missouri and Columbia River basins.

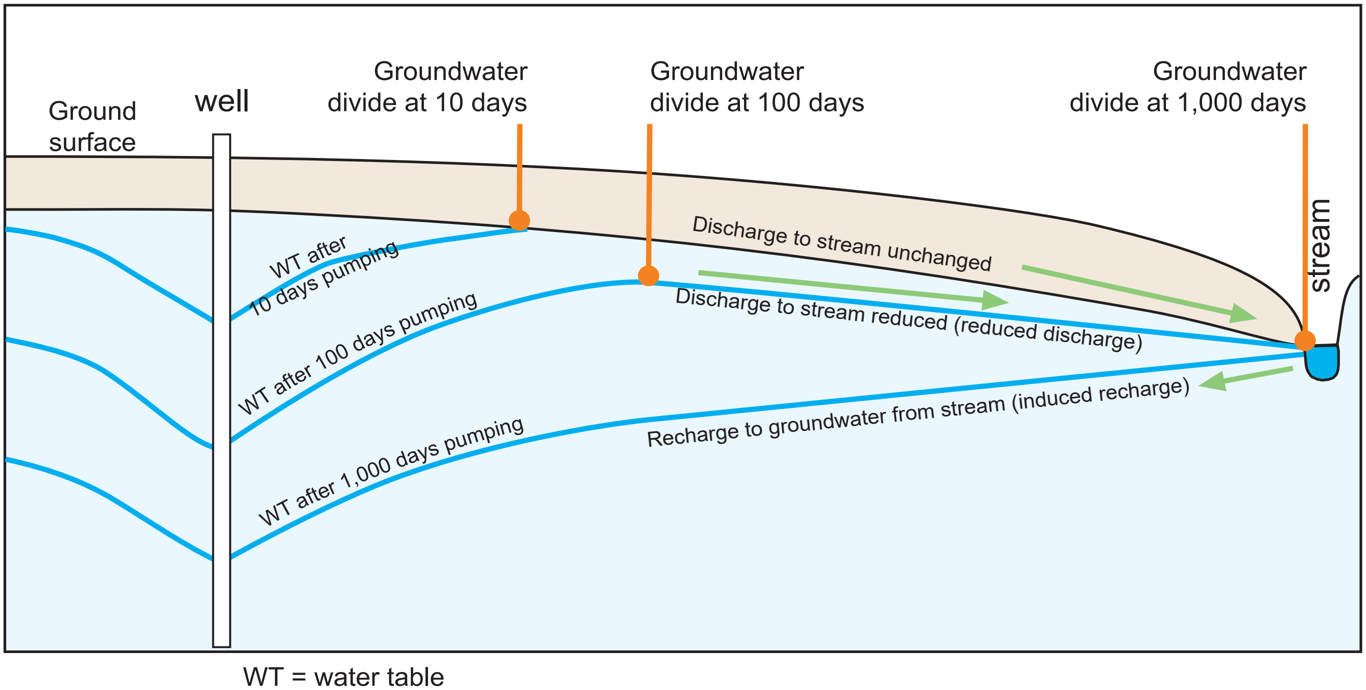


Figure 6. Schematic diagram illustrating the effect of pumping groundwater from a shallow aquifer near a gaining stream. If pumping is at a high enough rate or continues long enough, the stream reach near the well may transition from a gaining to a losing stream.

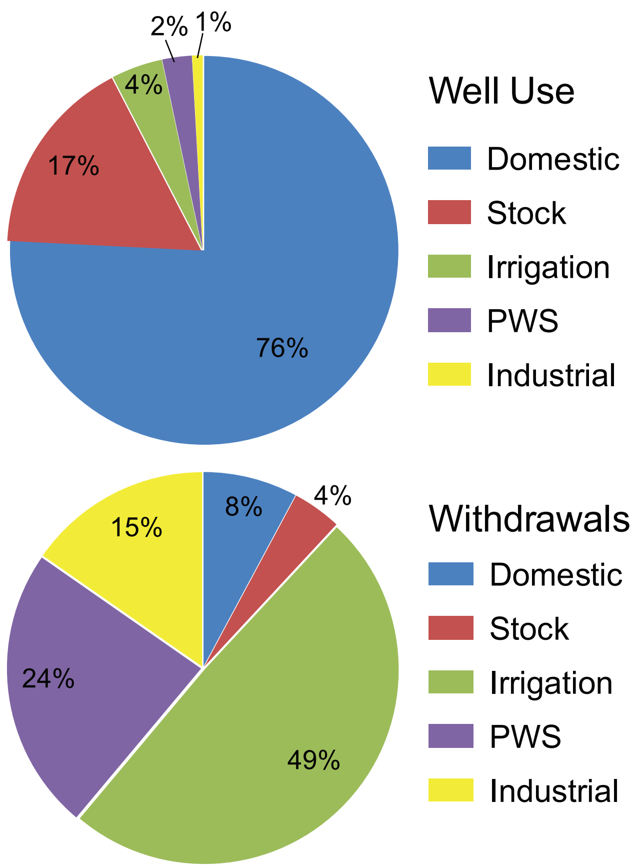


Figure 7. (A) Most Montana wells provide water for domestic and stock use; (B) The large annual withdrawals are for irrigation and public water supply.

The eastern two-thirds of Montana is in the northern Great Plains. This region is characterized by gently rolling to highly dissected topography, but also includes several prominent “outliers” of the Rocky Mountains. The Missouri and Yellowstone Rivers drain Montana’s Great Plains.

Each physiographic province manifests broad differences in geology, geologic history, and climate, which in turn create different hydrogeologic settings and groundwater conditions. Within the intermontane basins groundwater generally occurs in shallow alluvial (sand and gravel) water table aquifers, and in deep, confined to semi-confined basin-fill aquifers. Both aquifer types store large amounts of groundwater, are highly productive, and are utilized. Fractured-rock aquifers occur in the mountains that surround the valleys.

Layers of sedimentary sandstone and limestone form the principal bedrock aquifers in the Great Plains region. These aquifers are not as productive but are highly utilized. Locally, alluvial aquifers within major river valleys are also important sources of water for this region.

Intermontane Basins

The Northern Rocky Mountain Intermontane Basins encompass the western third of the State (41,500 mi^2). The “basins” are topographic and geologic

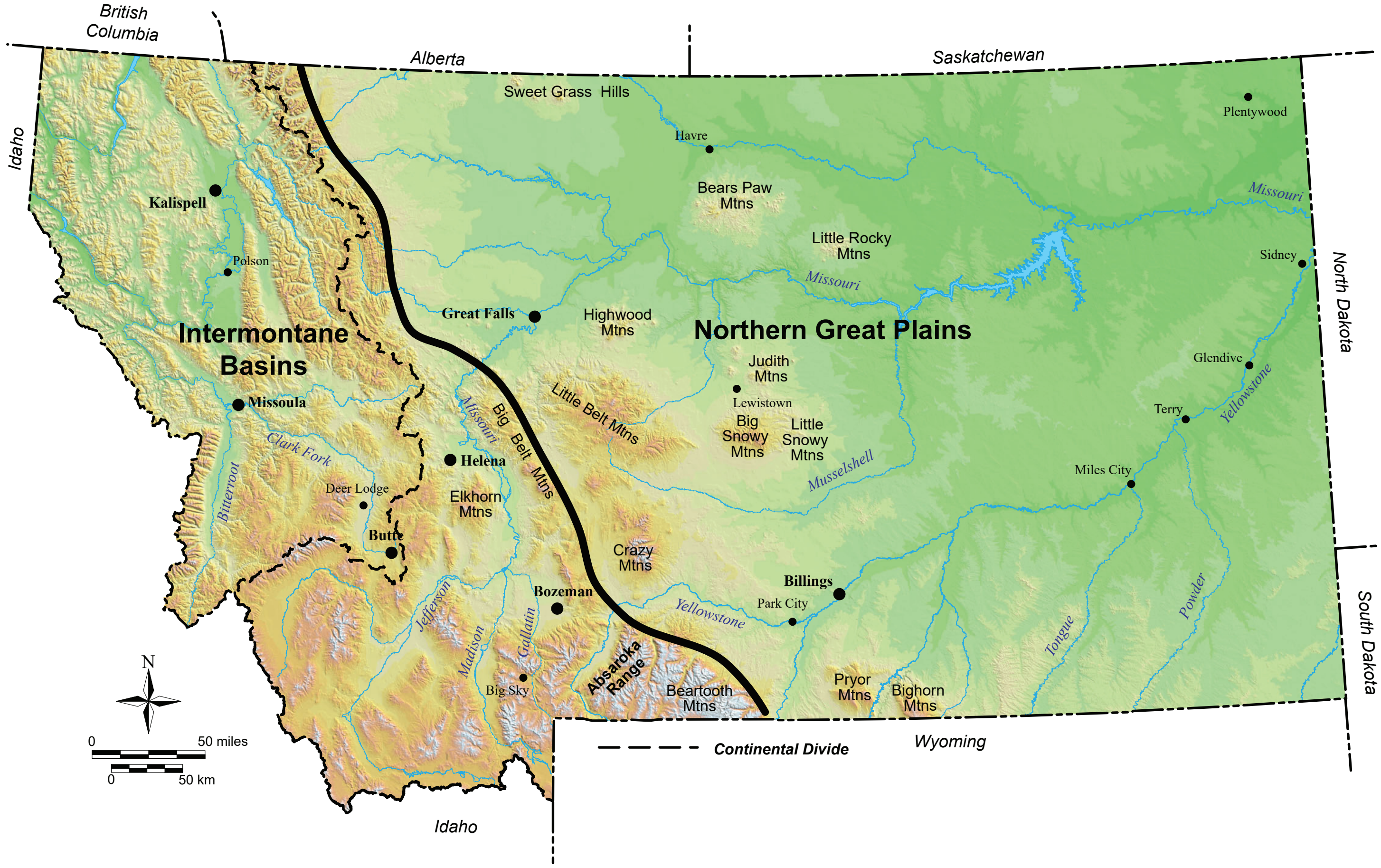


Figure 8. Montana straddles the northern Intermontane Basin and the northern Great Plains physiographic provinces.

features that are structurally down-dropped relative to the surrounding mountains and are filled with unconsolidated to poorly consolidated deposits (Kendy and Tresch, 1996). The topography in the region is varied. Mountain ranges in northwest Montana typically are separated by narrow, steep-sided valleys, whereas intermontane basins in southwest Montana are wide and deep, reflecting a greater degree of structural down-dropping.

The climate in the intermontane basin region is characterized by cold winters and mild summers; the average winter temperature is 24°F, while the average summer temperature is 58°F. The region is wetter than the Great Plains of Montana, receiving an average of 32 in per year from moisture-rich Pacific maritime air in the winter, spring, and fall, and from strong convective systems in the summer. Higher elevations receive heavy winter snowpack that is a significant source of water to valley bottoms. Steep precipitation gradients exist in the intermontane basin region. Many valley bottoms receive less than a foot of moisture, whereas adjacent mountain ranges may receive more than 60 in of annual precipitation. Historical trends show the region becoming slightly warmer and drier; between 1950 and 2015, the average annual temperature increased by 0.4°F/decade while average annual precipitation decreased by 0.6 in/decade (Whitlock and others, 2017).

Physiographic Setting

The intermontane basins contain perennial streams and associated floodplains. In most basins, modern floodplains are adjacent to older, higher river terraces. These features grade to pediments, alluvial fans, or glacial deposits that meet mountain fronts with an abrupt change in slope. Mountain fronts commonly coincide with faults or fault systems along the down-dropped blocks that form the basins. Intermontane basins of Montana range in area from about 30 to 710 mi^2. The basins are filled with unconsolidated to poorly consolidated Tertiary and Quaternary deposits that range from several hundred to several thousand feet thick (Tuck and others, 1996).

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. Many smaller valleys follow the northwest–southeast structural grain inherited from structures in the oldest rocks in western Montana.

During Pleistocene time several Cordilleran ice sheets flowed south from Canada and filled most of the northern basins. Mountain (alpine) glaciers existed in many of the southwestern Montana mountain ranges. The largest lobe of the Cordilleran ice sheets flowed down the Kalispell Valley, terminating at Polson. A lobe that extended down the Kootenai River valley dammed the Clark Fork River near the present-day Montana–Idaho border, forming Glacial Lake Missoula. Glacial deposits include till (a poorly sorted mixture of silt, clay, sand, and gravel), glacial lacustrine silt and clay, and outwash plains of coarse-grained alluvial deposits (Smith and others, 2013).

After the last glaciers retreated, streams deposited unconsolidated sand, silt, clay, and gravel along their channels, floodplains, and low terraces.

Basin-Fill Aquifers

The Cenozoic basin-fill aquifers consist of unconsolidated to semi-consolidated Tertiary age sediments overlain by unconsolidated Quaternary sediments (QTsed). These aquifers are the most productive and developed in Montana. Although the thickness and permeability of the deposits vary among the basins, these aquifer systems are generally similar. They contain: (1) shallow, unconfined alluvial aquifers associated with modern drainage systems, (2) interbedded fine-grained aquitards with variable lateral continuity, and (3) confined to semi-confined aquifers at depth.

Most of the wells completed in the basin-fill aquifers are for domestic purposes (88 percent), with the remainder used for stock water, public water supply, and irrigation uses (fig. 9). Reported well yields range up to 4,000 gpm, with a median of 25 gpm. Most wells with reported yields greater than 500 gpm supply irrigation and public water supply systems. Although well depths range up to 3,200 ft, with eight wells exceeding 1,000 ft deep, the median well depth in basin-fill aquifers is 85 ft.

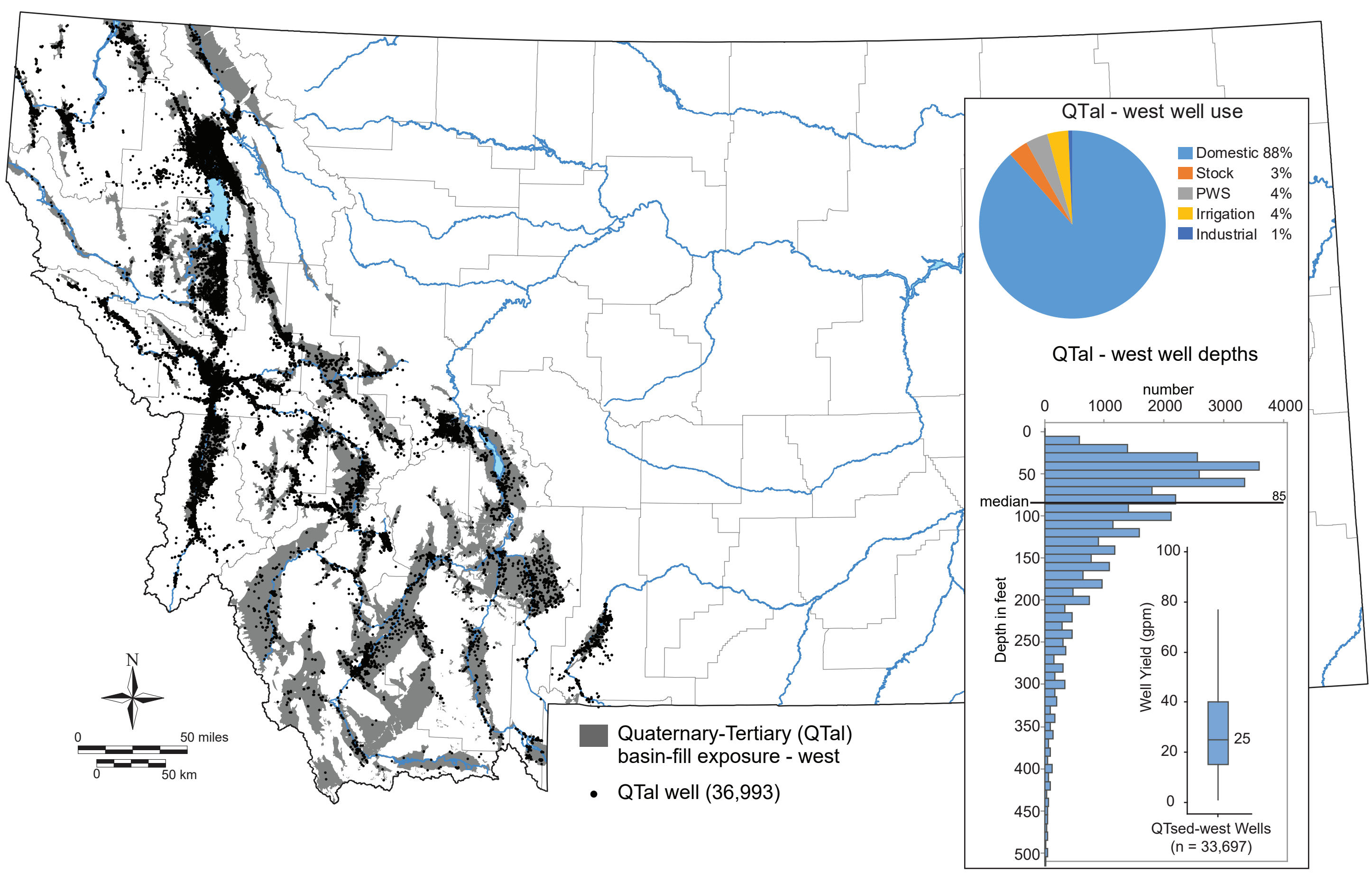


Figure 9. Distribution of wells in the western Quaternary–Tertiary basin-fill aquifers.

Water quality in basin-fill aquifers is generally excellent. The total dissolved solids (of 1,902 samples) range from less than 50 to 4,215 mg/L with a median of 240 mg/L; the water is predominantly a Ca-HCO_3 type water. Locally, high-density, unsewered residential development has resulted in nitrate contamination to shallow basin-fill aquifers in Butte's Summit Valley (LaFave, 2008) and the Gallatin Valley (Kendy, 2001).

Shallow basin-fill aquifers

The shallow basin-fill aquifers contain coarse-grained Quaternary alluvial or Tertiary deposits that consist of glacial outwash and stream alluvium. The aquifers are generally within 75–80 ft of land surface and typically are unconfined. Shallow aquifers are important water sources but are limited in aerial extent. Recharge to shallow basin-fill aquifers occurs by infiltration of precipitation and stream losses. In valleys with unlined irrigation canals, leakage from canals is a significant source of recharge. Groundwater discharge is through springs and seeps along valley bottoms, gaining reaches of perennial streams, transpiration by plants, and wells.

One of the most permeable and utilized shallow aquifers in western Montana supplies water to the city of Missoula. The Missoula Valley is a wedge-shaped intermontane basin that is drained by the west-flowing Clark Fork and the north-flowing Bitterroot Rivers. The basin is underlain by unconsolidated Pleistocene deposits that make up the Missoula Valley aquifer, the municipal water supply for the city of Missoula, and a designated sole-source aquifer by the U.S. Environmental Protection Agency. The aquifer matrix, deposited by glacial meltwater, ranges in size from silt and fine sand to gravel and cobbles. The aquifer is 100 to 150 ft thick and is underlain by relatively impermeable, fine-grained Tertiary sediments. The basal part of the aquifer is composed of 50 to 100 ft of highly permeable, coarse-grained sand and gravel (Woessner, 1988). Wells completed in this zone reportedly yield as much as 4,100 gallons per minute (gpm).

Groundwater in the Missoula Valley aquifer is unconfined, and the water table ranges from 10 to 60 ft below ground surface. Flow direction is generally from the Clark Fork River southwest toward the Bitterroot River (fig. 10). Estimates of leakage from the Clark Fork River suggest that it provides 80 to more than 90 percent of the recharge to the aquifer (Woessner, 1988; Miller, 1991). Groundwater flow rates are on the order of 7 to 18 ft per day. The water quality is characterized by total dissolved solids less than 300 milligrams per liter (mg/L; LaFave, 2002).

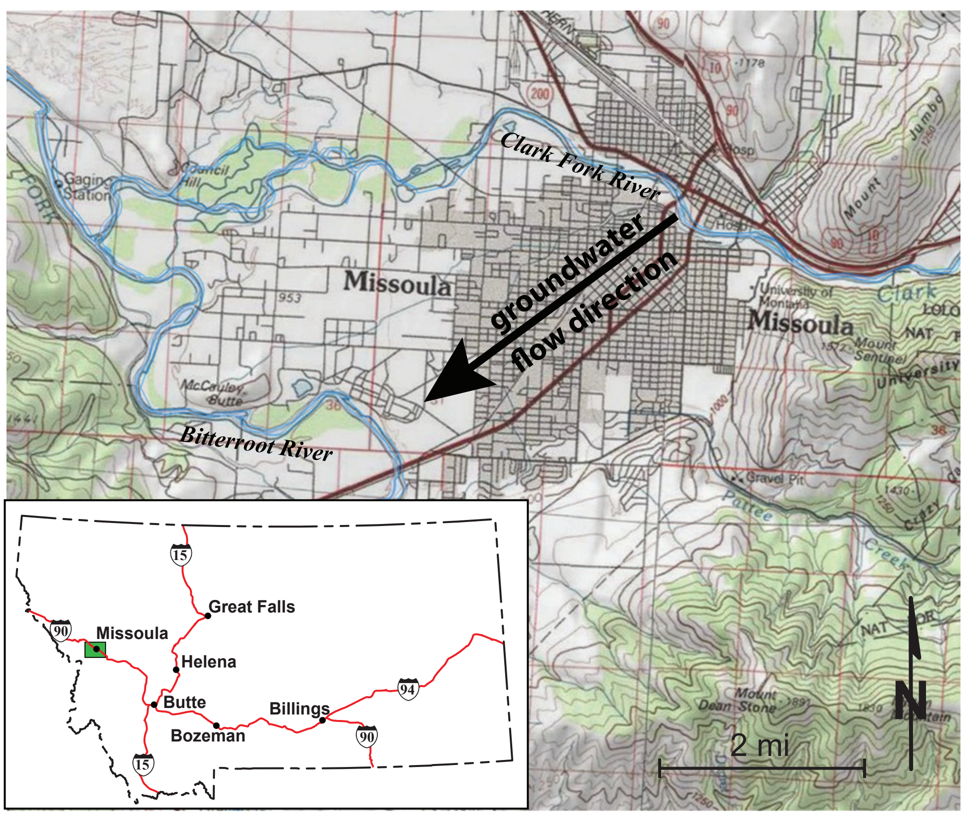


Figure 10. Groundwater flows beneath the city of Missoula from the Clark Fork to the Bitterroot River.

The Missoula Valley aquifer is vulnerable to contamination because of the shallow depth to water, the urban setting, its high transmissivity, and direct connection to surface water. These conditions require vigilant monitoring to support protection of the resource.

Confining units

Clay and silt aquitards, interbedded in the basin-fill, separate shallow and deeper aquifers. The confining layers consist of fine-grained floodplain deposits, silty till, and clay-rich glacial lake deposits. Fine-grained Tertiary deposits make up some of the deeper confining units. Confining units rarely completely seal aquifers; it is more common that the overlying units are partially confining, creating “leaky confined” aquifers. In the Kalispell Valley, a confining layer 100 to 700 ft thick overlies the deep alluvial aquifer there (Rose, 2018). The aquitard consists of thick beds of till deposited by glacial ice and laminated beds of silt, clay, and minor amounts of silty sand and gravel interpreted to be glaciolacustrine deposits from ancestral Flathead Lake (Smith, 2000). The thickness and broad lateral extent of these deposits suggest that the aquitard effectively seals the deep aquifer, allowing artesian conditions to build, and offering good protection from surface contamination.

Deep basin-fill aquifers

Deep basin-fill aquifers contain fine to coarse-grained alluvial deposits and Tertiary sedimentary rocks, generally at depths greater than 75 ft below land surface. The deep basin-fill includes interbedded sand, gravel, silty sand, and local claystone that are widespread but difficult to map with available subsurface data. In most basins it is difficult to distinguish deep Quaternary alluvium from Tertiary deposits based on drillers’ well-log descriptions. The thickness of the basin-fill is variable and poorly constrained. 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 mi 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 sediments before encountering volcanic rocks. Well records from the Bitterroot Valley report that basin-fill deposits are, in places, 2,400 ft thick (Norbeck, 1980). Most Tertiary-age basin-fill deposits are shale and/or sandy mudstone and often are confining units or marginal aquifers. Discontinuous permeable sandstones and conglomerates, however, locally serve as important aquifers, notably in the Bitterroot Valley (Lonn and Sears, 2001) and St. Regis area (Smith, 2006).

In the Kalispell Valley, a laterally continuous Pleistocene sand and gravel deposit that is the valley’s primary water supply is an example of a very productive deep aquifer that occurs below a thick confining unit (LaFave and others, 2004; Rose, 2018). It supports domestic, irrigation, and public water supply wells. The impact of development on this aquifer as a source of irrigation water in the 1970s is well documented as a result of long-term monitoring.

Prior to widespread development, water levels in the aquifer showed a uniform annual cycle, with water levels peaking in early July and reaching a seasonal low in February; the magnitude of annual fluctuations was less than 2 ft. Between 1963 and 1973, the water-level trend was stable (at an altitude of about 2,930 ft above mean sea level at well 131524, fig. 11A), suggesting that the aquifer was in a state of equilibrium. Thus, even though the water level fluctuated seasonally, it fluctuated around a stable mean, indicating that each year the recharge and discharge were in balance.

Between the early 1970s and the mid 1980s, the number of the wells—including irrigation wells—completed in the deep aquifer tripled. During this period of increased groundwater development the water level dropped, indicating that the aquifer was no longer in equilibrium; that is, the increased pumpage was removing water from aquifer storage (fig. 11B). Groundwater development of the deep aquifer, as measured by the number of wells, continued. By the early 2000s, however, the long-term water-level decline slowed, and by the mid- to late 2000s the groundwater development stabilized and a new equilibrium was achieved at a level about 10 ft lower than the pre-development level. The seasonal fluctuations around the new equilibrium are very different, controlled by pumping. Whereas pre-development water levels peaked in early July and reached lows in February, the new response shows water levels reaching seasonal lows in the summer months, when irrigation pumping is greatest, and reaching seasonal highs in the winter months when water levels recover. The magnitude of seasonal change is also much greater—10 to 20 ft per year (fig. 11C).

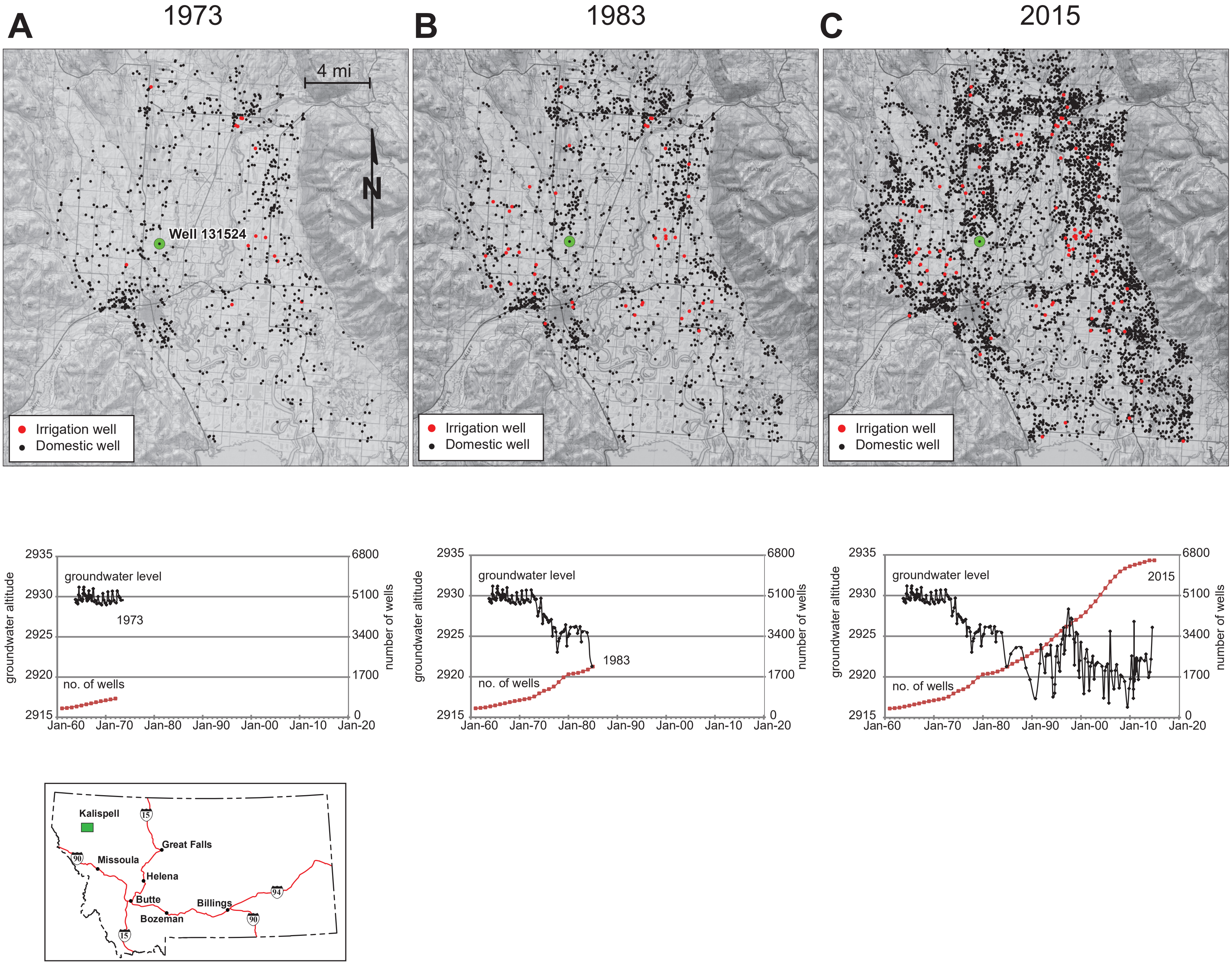


Figure 11. (A) Wells and groundwater levels in the Kalispell valley deep aquifer: 1973. (B) Wells and groundwater levels in the Kalispell valley deep aquifer: 1983. (C) Wells and groundwater levels in the Kalispell valley deep aquifer: 2015.

Fractured-Rock Aquifers

The mountain ranges that separate the basins in western Montana are composed of relatively impermeable bedrock that consists mostly of the Precambrian Belt Supergroup, and of Cretaceous and Tertiary igneous and volcanic rocks. Although relatively impermeable, in many places there is sufficient fracture permeability to provide water to low-yield wells. The Belt rocks consist of a thick sequence of metasedimentary formations that form the mountains and underlie the valleys of northwestern and a substantial part of southwestern Montana. The Belt rocks are generally fine-grained clastic rocks (sandstone, siltstone, and mudstone) and carbonate rocks (limestone and dolomite) that have been subjected to low-grade metamorphism; they are well consolidated and fractured (LaFave and others, 2004; Smith, 2006). The fractured-bedrock aquifers (FB) in western Montana consist mostly of Belt rocks and lesser amounts of Precambrian gneiss and schist in the Absaroka and Beartooth mountains.

In southwest Montana, Cretaceous plutonic igneous rocks (Idaho and Boulder Batholiths) intruded older Belt rock units. Volcanic rocks of Cretaceous (Elkhorn volcanics) and Tertiary (Lowland Creek volcanics, Absaroka volcanics) are present in the eastern part of the Intermontane Basin region. In the Great Plains region, the Tertiary Adel Mountain volcanics occur off the west flank of the Little Belt Mountains, and Cretaceous Sliderock Mountain volcanics occur off the northwest flank of the Beartooth Plateau (fig. 12; Vuke and others, 2007). Collectively these igneous and volcanic rocks are referred to as the Tertiary and Cretaceous igneous fractured-rock aquifers (TKig).

For the remainder of this discussion the fractured-bedrock aquifers (FB) and the Tertiary and Cretaceous igneous fractured-rock aquifers (TKig) are grouped together as "fractured-rock aquifers" because of their similar geographic and topographic setting, and similar hydraulic properties.

Where they are near the surface, the fractured-rock aquifers contain sufficient secondary permeability (fractures) to yield small supplies of water to wells. The occurrence, size, and orientation of fracture openings are spatially variable, resulting in large variations in well yields. Due to low primary porosity, the storage capacity of fractured-rock aquifers is typically

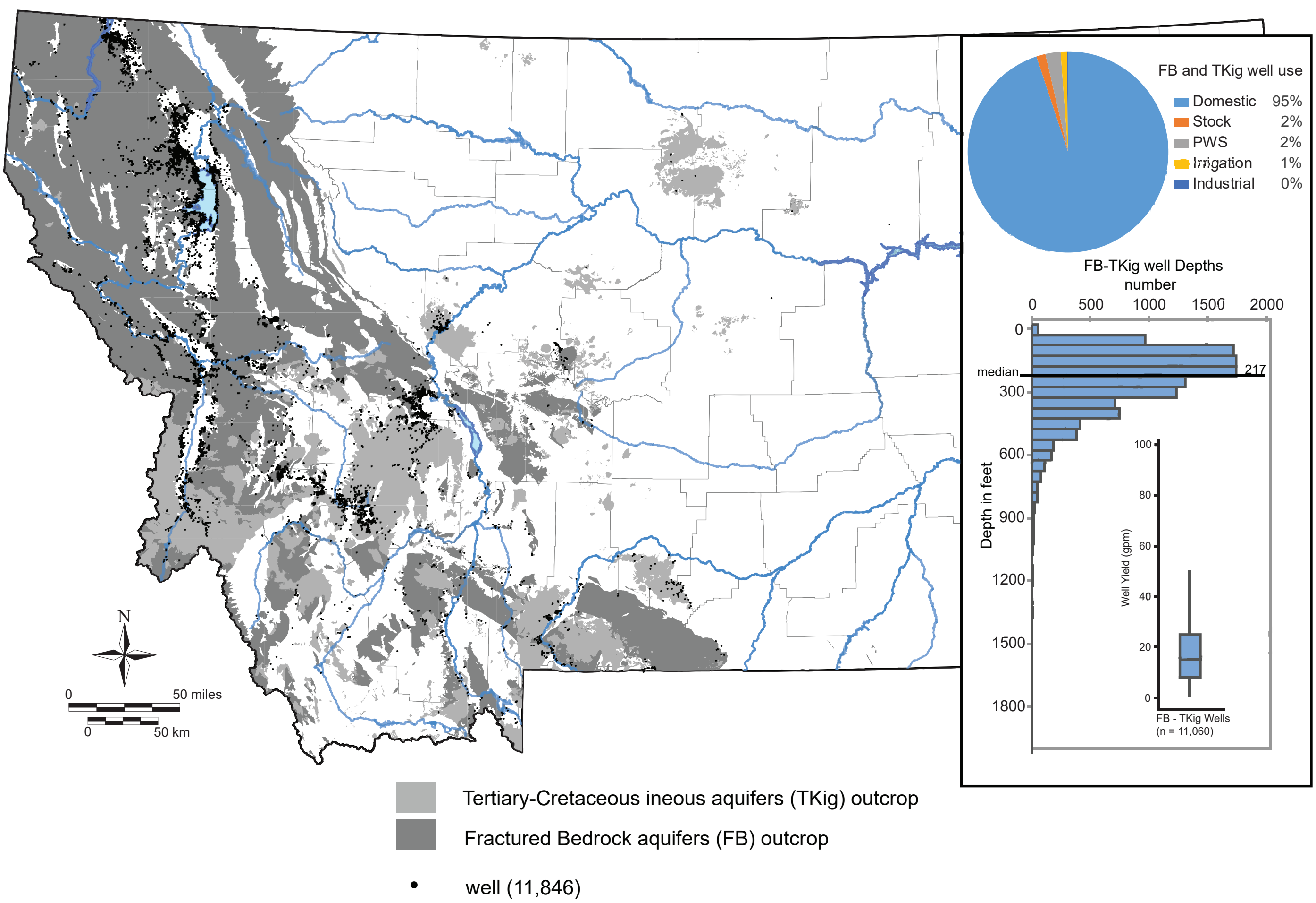


Figure 12. Distribution of wells in fractured bedrock and Tertiary–Cretaceous igneous aquifers.

low; that is, the amount of water stored and released is small relative to basin-fill aquifers. The low storage capacity manifests in large seasonal water-level fluctuations, as shown by the hydrograph from the Madison Valley (GWICID 128327), and interannually as demonstrated in the hydrograph from the Mission Valley (GWICID 77922; fig. 13).

On a regional scale, fractured-bedrock aquifers within the mountainous areas are in hydraulic communication with adjacent deep basin-fill aquifers (LaFave and others, 2004; Smith and others, 2013). Mountains receive more precipitation than the valleys and occupy most of the region's land surface. Infiltration of precipitation and surface water into bedrock aquifers and diffuse flow through fractures towards valley bottoms contribute mountain front recharge to basin-fill aquifers. Mountain front recharge is an important source of water to the basin-fill systems.

Figure 12 shows the distribution of wells completed in fractured-rock aquifers. Most are domestic wells (95 percent), reflecting the residential development in the mountainous fringe along the intermontane basins. Well depths and yields reflect the variable nature of fracture (permeability) distributions. The median well depth is 217 ft; yields are typically between 5 and 25 gpm, with a median of 15 gpm.

In general, fractured-rock aquifers contain low-TDS water; concentrations typically range from about 160 to 400 mg/L. However locally elevated concentrations of arsenic (LaFave, 2006), radon, and uranium (Caldwell and others, 2013) are associated with the Tertiary–Cretaceous igneous intrusive rocks.

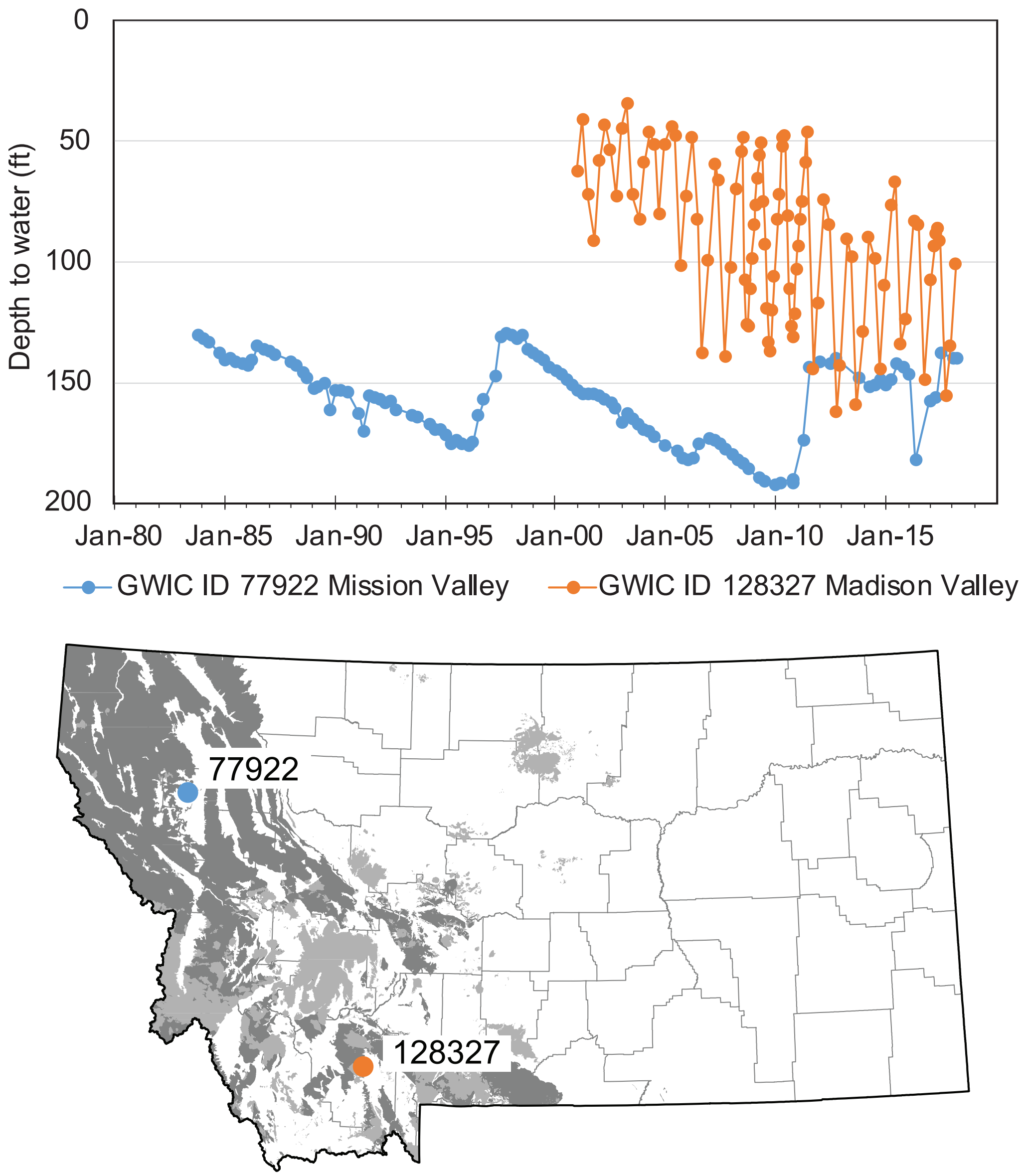


Figure 13. The low storage capacity of fractured-rock aquifers can result in large seasonal (128327) and interannual (77922) water-level fluctuations.

Northern Great Plains Region

The area lying east of the Rocky Mountains is in the Northern Great Plains region (fig. 8). The land surface consists predominantly of gently rolling hills slightly eroded by intermittent streams; several isolated mountain ranges also occupy the area. Grass-covered rangeland is interspersed by non-irrigated farmland in the uplands and, where soil and water permit, by irrigated farmland mostly in the valleys. Locally, badlands have developed in easily eroded shales and siltstones. Badlands are common throughout the Missouri River "breaks" and in the lower Yellowstone River valley between Terry and Sydney where the upper part of the Hell Creek and Fort Union Formations are exposed.

The region is sparsely populated, with most people living in cities and towns. The larger urban areas of Great Falls and Billings are supplied by surface water, whereas rural residents and a few towns rely on groundwater.

The Plains region is characterized by semi-arid-continental climate with cold winters, cool, moist springs, and hot, dry summers. The average minimum winter temperatures in the five eastern Montana climatic divisions range from 8 to 15°F; the average summer highs range from 77 to 83°F. Precipitation in the Plains region is about half that of the Intermontane Basin region. On average the Plains region receives 15.5 in of precipitation per year, primarily in the spring and summer. Trends in annual temperature since 1950 show an average increase in annual temperature of 0.5°F/decade with no apparent historical change in average annual precipitation (Whitlock and others, 2017).

The region is drained by the Missouri and Yellowstone River systems. Most of the region's population and irrigated agriculture are near these rivers. The hydrograph for the Yellowstone River (which is one of the longest undammed rivers in the U.S.) at Miles City shows a large spring peak flow related to snowmelt; the flows decrease through the summer to winter baseflow conditions (fig. 14).

The Northern Great Plains region of Montana is underlain by a thick sequence of sedimentary rocks. Principal aquifers occur in the sandstones and limestones that lie within 1,500 ft of the surface and contain water of sufficient quantity and quality to be suitable for use (fig. 1, cross section B). Aquifers also occur in the alluvial deposits along major drainages, in localized glacial outwash deposits, and in buried valley aquifers in northern Montana.

Sedimentary Bedrock Aquifers

In the Great Plains region, sedimentary bedrock aquifers are the primary source of groundwater outside of the river valleys.

Madison Aquifer

The basal (deepest, stratigraphically) principal aquifer of the Montana plains is the Madison Group. Although this formation underlies most of eastern Montana, it is only a freshwater aquifer near outcrop areas where it is relatively close to the surface. The Madison Group consists of a sequence of Mississippian marine carbonates and evaporites deposited in a shallow-water environment (Downey, 1984). The Madison Group, from oldest to youngest, contains the Lodgepole Limestone, the Mission Canyon Limestone, and the Charles Formation. The Lodgepole is a micritic limestone and dolomite. It grades into the Mission Canyon, which consists of coarse to fine crystalline limestone with evaporite minerals near the top. The Charles Formation, the uppermost unit, consists of the evaporates anhydrite and halite interbedded with dolomite, limestone, and argillaceous units. Pre-Jurassic erosion has removed most of the Charles Formation from central and southern Montana (Downey, 1984), and the Lodgepole generally has very low permeability; therefore, the Madison Group aquifer consists primarily of the Mission Canyon Limestone. The thickness of the Mission Canyon ranges from 300 to 600 ft (Miller, 1981).

In central Montana, the Madison Group crops out mainly along the northern flanks of the Little Belt and Big Snowy Mountains. It is also prominently exposed along the northeast flank of the Pryor Mountains and in narrow exposures in mountain ranges in southwest Montana (fig. 15). The Madison typically dips downward steeply away from the mountain fronts and decreases in slope with distance from them. It is deeply buried (5,000–8,000 ft below the surface) under most of eastern Montana. The limestone is relatively impermeable; however, where fracturing and karst features are well developed, there is the potential for large well yields and fast advective flow paths. The permeability and well yields are variable; in places closely spaced wells have very different reported yields.

Precipitation and stream loss in outcrop areas are the primary sources of recharge. In general, water

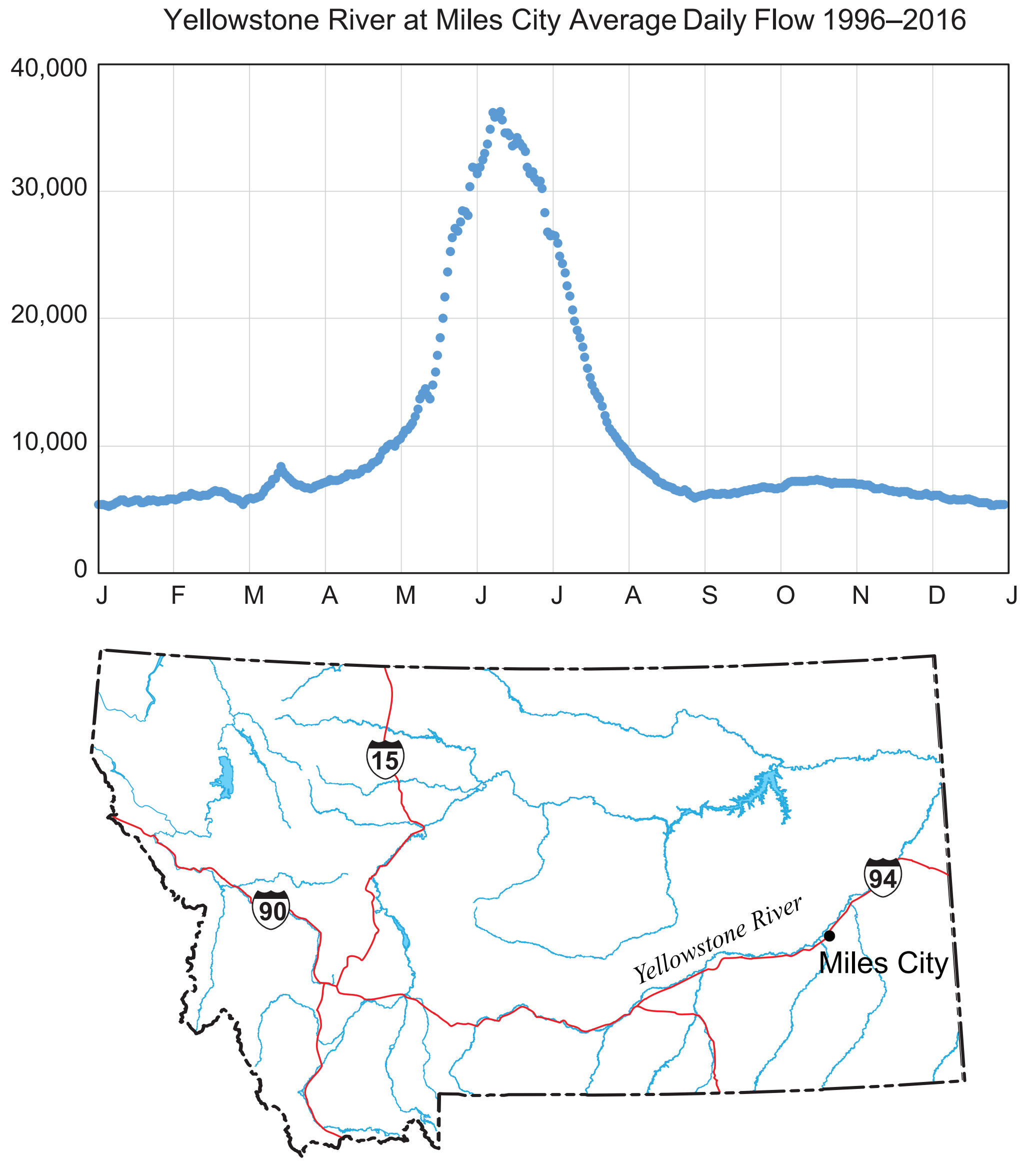


Figure 14. Snowmelt-dominated hydrograph of the Yellowstone River near Miles City, Montana.

moves outward from the mountain recharge areas; water in the Madison is confined except near outcrop areas (Feltis and Shields, 1982; Feltis, 1980; Madison, 2016).

Deep drilling depths and poor water quality at depth have hindered widespread use of the Madison aquifer. Records from the GWIC database indicate there are about 1,100 wells completed in the Madison aquifer, with a reported use of domestic, stockwater, public water supply, or industrial/commercial (fig. 15). Roughly 75 percent of Madison wells are in Cascade County between the Little Belt Mountains and the Missouri River near Great Falls; other wells completed in the Madison are near outcrop areas (fig. 15).

In Cascade county, near Great Falls, the Madison is about 350 ft below the land surface; however, it is exposed about 25 mi to the south–southeast along the north flank of the Little Belt Mountains (Smith, 2008). More than 900 wells (roughly 75 percent of all Madison wells statewide) use the Madison aquifer, mostly between the Little Belt Mountains and the Missouri River near Great Falls (fig. 16). Several streams lose flow to the Madison aquifer in outcrop areas, and groundwater moves from the mountain recharge areas, in the Little Belt Mountains, towards the Missouri River (Feltis and Shields, 1982; Madison, 2016). CFC-age dating of Madison aquifer water near the Missouri River at Great Falls returned an apparent age of 26 yr, suggesting that flow rates between the Little Belt Mountains (recharge area) and The Missouri River (discharge area) could be as much as 15 ft per day (LaFave, 2012). The fracture and solution permeability and the relatively rapid flow rates make water storage in the Madison especially sensitive to drought cycles.

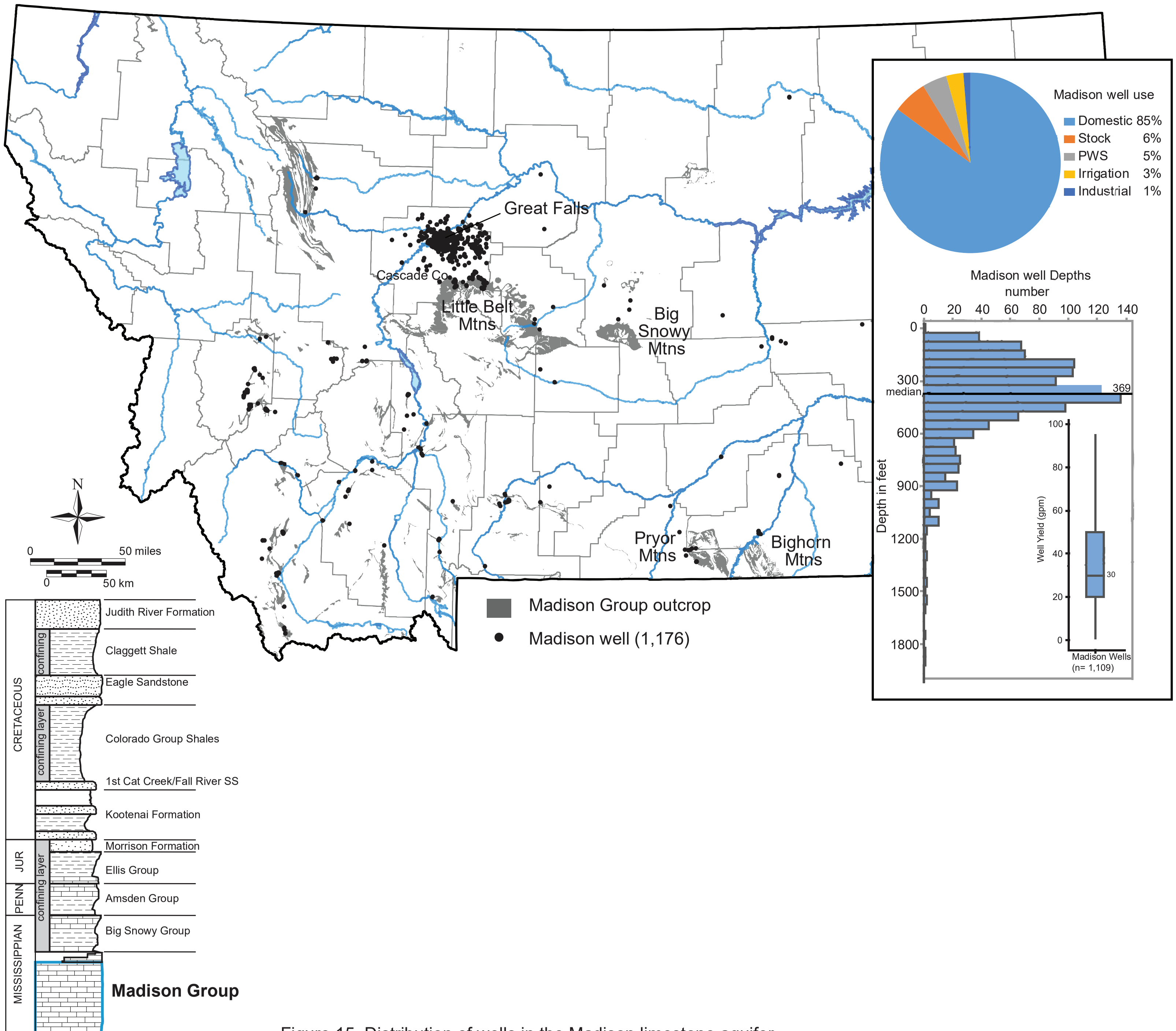


Figure 15. Distribution of wells in the Madison limestone aquifer.

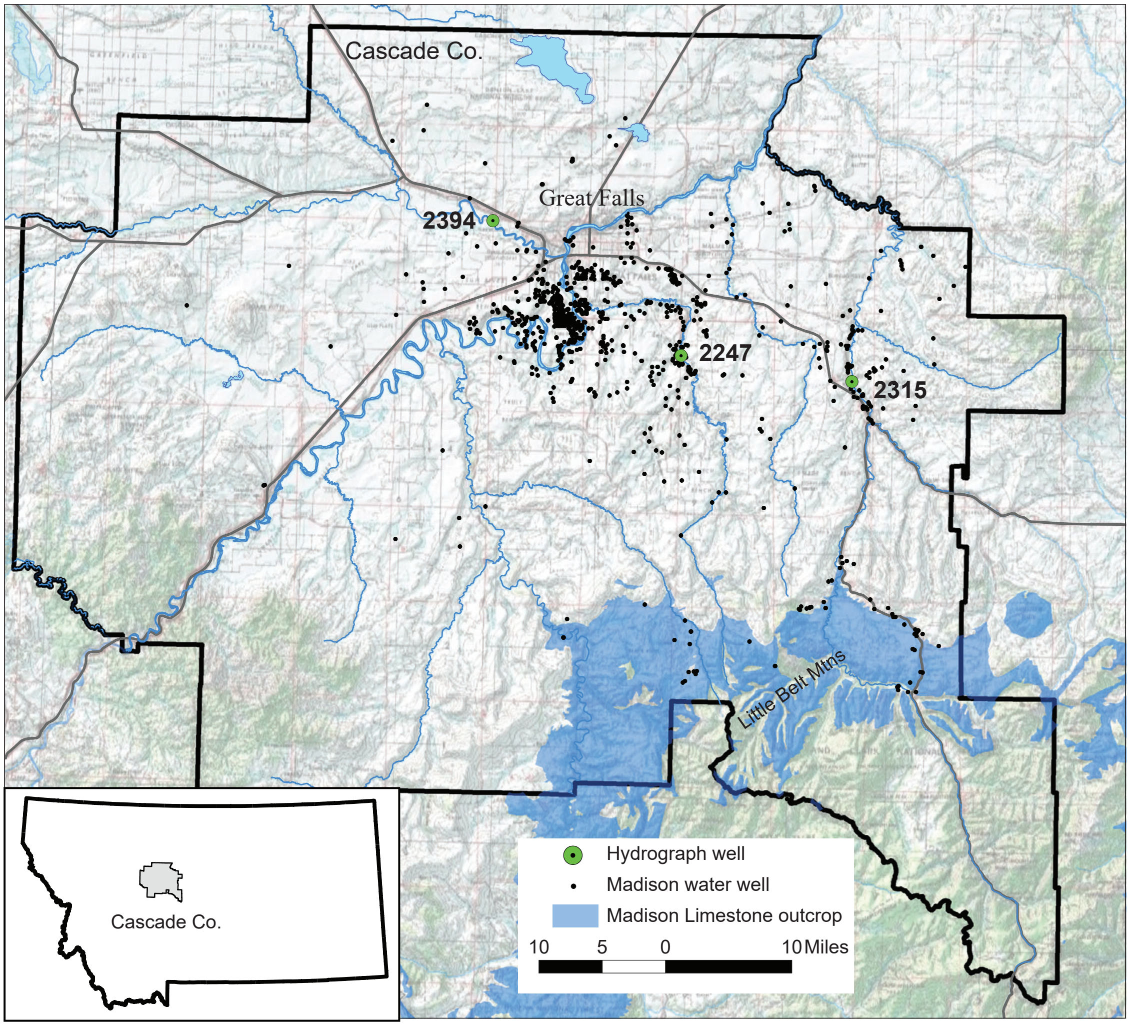


Figure 16. More than 900 wells (black dots) obtain water from the Madison Limestone near Great Falls. The Madison Limestone is exposed at the surface in the Little Belt Mountains (blue area on map), but is more than 400 ft below the surface at Great Falls.

Data from the Ground Water Information Center database show that in Cascade County, between about 1995 and 2006, the number of wells completed in the Madison aquifer went from about 400 to 800 (fig. 17). During this same time, water levels in Madison aquifer observation wells near Great Falls dropped about 30 ft (fig. 17). Based on these data, it appeared that water was being removed from the aquifer faster than it could be replenished, and many water users began to question the aquifer's sustainability.

Since 2006, however, even though new wells continued to be drilled into the Madison aquifer, water levels have climbed to levels higher than those measured in 1995. Although there may be a development impact, the data suggest that groundwater withdrawals are not driving water-level changes. Rather climate, or more specifically precipitation, appears to be the primary water-level control. Comparing water-level trends to departures from average annual precipitation shows that the period when water levels declined between 1995 and 2005 coincided with below-average precipitation (fig. 17). Recovery occurred between 2006 and 2013 when the climate was relatively wet.

This example highlights the importance of long-term water-level data with regard to developing a comprehensive understanding of groundwater storage change. The Madison aquifer system is dynamic and is strongly impacted by short- and long-term climate variability.

The Madison aquifer is also the source of several large springs; four of these springs supply water for fish hatcheries. Giant Springs in Cascade County, Big Spring in Fergus County, Hatchery Spring in Gallatin County, and Blaine Spring in Madison County all supply a consistent flow of relatively low-TDS water with very similar water chemistry despite the geographic distance between them (fig. 18).

Most of the wells completed in the Madison aquifer are for domestic and stockwater uses and are less than 1,000 ft deep; the median reported Madison

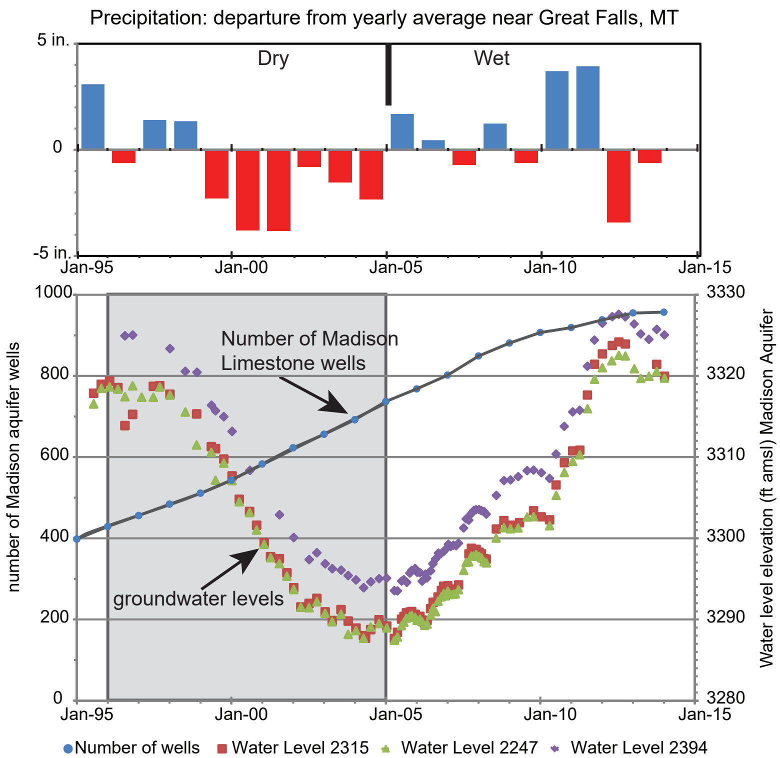
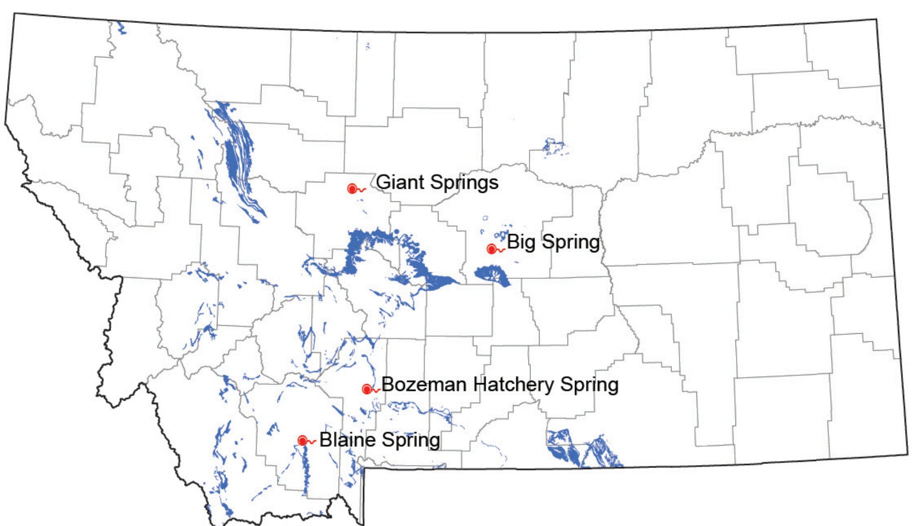


Figure 17. Between 1995 and 2005, the number of wells drilled into the Madison Limestone around Great Falls nearly doubled. During the same time period, water levels in the aquifer dropped 30 ft. However, this was also a very dry period, as indicated by the departure from average precipitation plot.above. Water levels recovered following several "wet" years even though wells continued to be drilled into the aquifer. Location of the hydrograph wells is shown in figure 16.



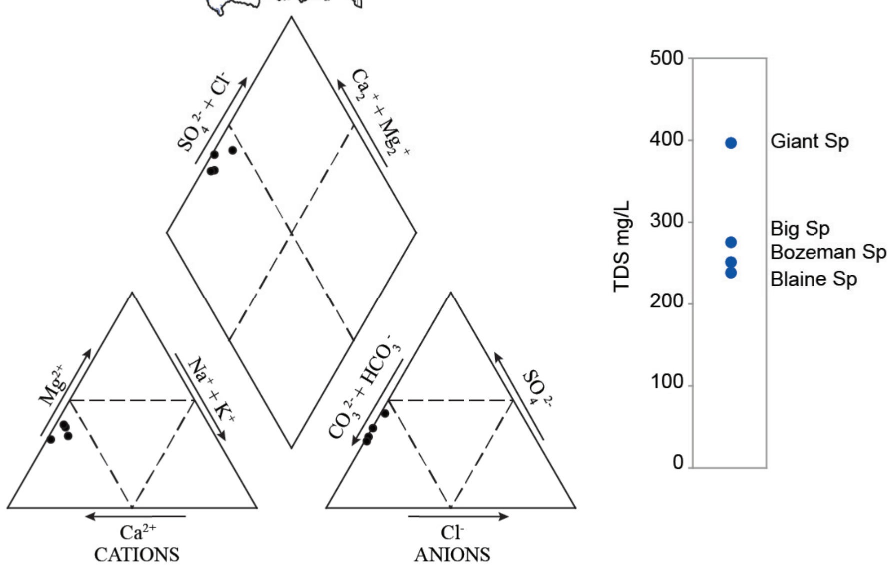


Figure 18. The Madison limestone feeds springs that supply chemically uniform water to four fish hatcheries.

water-well depth is 369 ft. Typical reported well yields are greater than other bedrock aquifers, ranging from 10 to over 100 gpm (10th and 90th percentile of reported yields); a few wells (eight) have reported yields greater than 1,000 gpm; the median yield is 30 gpm (fig. 15).

Water quality in the Madison aquifer is generally good, especially near outcrop areas. The TDS concentrations (from 130 samples) range from about 40 to 5,300 mg/L, with a median of 412 mg/L. The water is generally a Ca-Mg-HCO_3 type water; however, water with greater TDS concentrations (greater than 500 mg/L) is typically enriched in SO_4. The water generally becomes more mineralized with distance from the outcrop-recharge areas.

The Madison is an oil and gas reservoir in north-central and eastern Montana. In the North Cut Bank and Red Creek oil fields the Madison yields saline water and commercial quantiles of oil; the saline water is used for secondary recovery in oil wells (Zimmerman, 1967).

Kootenai Aquifer

The Kootenai Formation is the basal Cretaceous unit in eastern Montana and marks the beginning of the early Cretaceous sea-level rise; equivalent units include the subsurface Lakota and Fusion Formations. The Kootenai is composed of lenticular fluvial or conglomeratic sandstone, siltstone, limestone, and shale that was deposited on an eroded Jurassic surface in a fluvio-deltaic environment. The Kootenai's thickness ranges from 100 to 600 ft (Condon, 2000; Levings, 1983).

The Kootenai is divided into three informal units. The lower (basal) unit is marked by a hard, coarse-grained arkosic "salt and pepper" sandstone and chert-pebble conglomerate. The basal sandstones, which form the primary aquifer, are informally referred to as the Third Cat Creek Sandstone (central Montana); the Sunburst Sandstone or the Cutbank Sandstone (northwestern plains); and the Pryor Conglomerate or the Lakota Sandstone (eastern Montana; Levings, 1983).

The middle unit, informally called the Second Cat Creek sandstone, is a brown-gray sandstone interbedded with siltstone and shales. The upper unit, the variegated argillaceous member, is composed of mottled red-maroon and grayish green shale and siltstone with a few thin beds of light gray friable sandstone. In places, a reddish color change distinguishes the Kootenai from the overlying shale units.

In central Montana, overlying the Kootenai Formation is a gray to tan, fine- to medium-grained sandstone known as the First Cat Creek sandstone (named by oil well drillers of the Cat Creek field in the 1920s) that yields water to wells. This is the lateral equivalent of the Fall River Sandstone in southeastern Montana, and the Dakota Sandstone of South Dakota and far eastern Montana. Although this unit is part of the Colorado Group, for this report it is considered part of the Kootenai aquifer.

In much of eastern Montana the Kootenai is too deep or its water too saline to be used. Only where the Kootenai crops out or is close to the surface does it yield potable water. The Kootenai is most widely used in Cascade County off the north flank of the Little Belt Mountains, and in the Judith Basin off the northeast flank of the Big and Little Snowy Mountains and the Judith Mountains (fig. 19). About 90 percent of the wells are in Cascade, Fergus, Judith Basin, and Petroleum Counties. The Kootenai is utilized to some extent off the north flank of the Pryor Mountains in south-central Montana and in the Big Sky area in southwest Montana.

Recharge is mainly from infiltration of precipitation and stream loss on upland outcrop areas. In places, the Kootenai also receives recharge from the underlying Madison Limestone. Regional groundwater flow in the Kootenai is to the east and north away from the Little Belt and Big Snowy Mountains. Where the Kootenai is buried, away from the outcrop areas, water is generally under artesian conditions, and flowing wells are common in the Judith Basin.

Most of the wells completed in the Kootenai aquifer are for domestic and stockwater uses. Reported well depths from 1,600 wells range from less than 30 to more than 4,000 ft. The median well depth is 189 ft, but about 260 wells exceed 1,000 ft. Well depths increase with distance from outcrop areas. Most reported well yields are between 5 and 60 gpm (10th and 90th percentile of reported yields), with a median of 15 gpm (fig. 19).

Water in the Kootenai generally is suitable for most uses. The TDS concentrations (from 140 samples) range from about 110 to 3,410 mg/L with a median of 455 mg/L. In Cascade County and in the Judith Basin, the water quality typically evolves with depth and distance from outcrop (recharge areas) from a

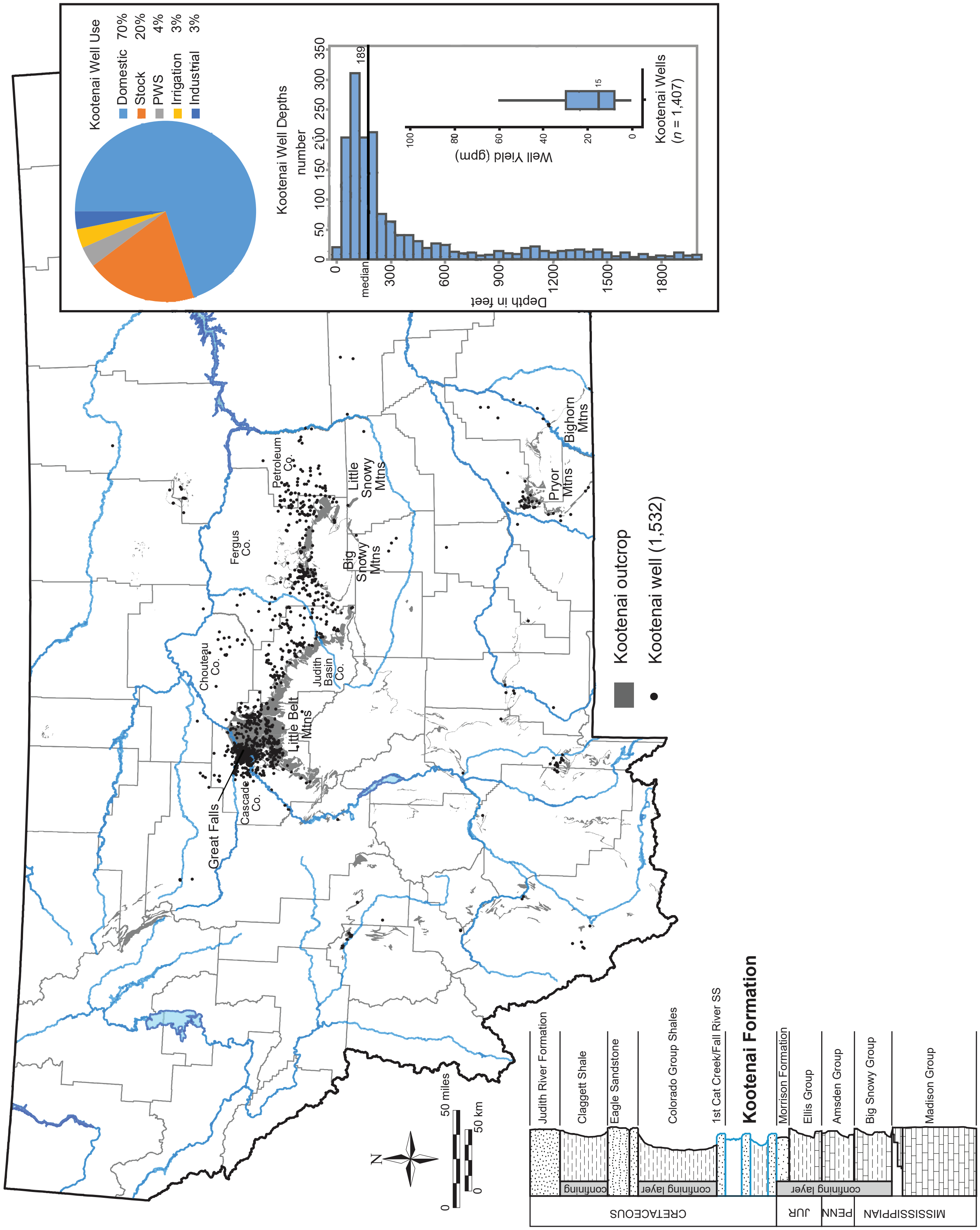


Figure 19. Distribution of wells in the Kootenai sandstone aquifer.

Ca-HCO_3 type water with TDS less than 500 mg/L to a Na-HCO_3 water with TDS greater than 500 mg/L. In the Big Sky area, Kootenai water quality is good, with a median TDS concentration of about 330 mg/L.

Cretaceous Shale (Colorado Group)

The Cretaceous Colorado Group shales (Kshale) serve as a regional confining layer that separates the Kootenai from the overlying Eagle Sandstone. The Colorado confining layer is about 2,000 ft thick in north-central Montana (Zimmerman, 1967), and is widely exposed in the northwestern and central part of the Great Plains region in Montana (hydrogeologic map, fig. 1). Although some sand layers within the Colorado Group yield water to wells, it is typically highly mineralized and not suitable for most uses.

Eagle Aquifer

The Upper Cretaceous Eagle Formation is an eastward-thinning wedge of non-marine to shallow-water marine strata that underlies most of the Great Plains region in Montana. The Eagle Sandstone is composed of alternating beds of gray, thick-bedded sandstone, sandy shale, and interbedded coal. The Eagle grades completely to shale (the Gammon Shale) from the Bowdoin and Porcupine domes east. In central Montana the basal Virgelle Sandstone member is 115 to 180 ft thick and consists of light gray, medium-grained, cross-bedded, massive sandstone; the upper part of the formation consists of shale, siltstone, and sandstone. A chert pebble lag deposit commonly marks the contact between the Eagle and overlying Claggett Shale (Vuke and others, 2007).

The Telegraph Creek Formation, a gray sandy shale and fine-grained sandstone, is a transitional unit between the underlying Colorado Group and the Eagle Formation. In eastern and east-central Montana the basal Virgelle Sandstone and the underlying Telegraph Creek cannot be differentiated in well logs (Feltis, 1982a). For this report the Telegraph Creek is included as part of the Eagle aquifer.

The thickness of the Eagle/Telegraph Creek increases from about 200 ft in the northwest to 900 ft in the northeast part of the State; near the Montana–Wyoming border it is as much as 1,200 ft thick (Feltis, 1982a). Sandstones in the Eagle are thickest in the western part of the Great Plains region; both the Eagle and the Telegraph Creek grade eastward into the Gammon Shale (Condon, 2000). The Eagle Formation is overlain, and confined, by the Claggett Shale.

In north-central Montana the Eagle crops out around the Sweet Grass Hills and on both sides of the Sweetgrass Arch. On the west side of the Sweetgrass Arch, the Eagle is represented only by the Virgelle Sandstone Member (Mudge, 1972). In central Montana it is exposed along the east flank of the Little Snowy Mountains and, in south-central Montana, the Eagle forms the prominent “Rim Rocks” surrounding the Billings area.

The distribution of water wells in the Eagle is shown in figure 20. The Eagle is recharged by infiltration of precipitation and by stream loss across outcrops. Vertical leakage into the Eagle across the overlying and underlying aquitards is probably minimal. Water generally flows from recharge areas along the flanks or upland/highland areas downgradient/down dip (Levings, 1982a). Water is generally under confined conditions except in outcrop areas. The Eagle is widely used between the Judith Mountains and the Little Rocky Mountains, and southeast of the Little Belt Mountains. Flowing wells, demonstrating artesian conditions, are common in this area. Unrestricted flow from uncapped and abandoned wells has resulted in localized loss of artesian pressure (Reiten, 1993).

In south-central Montana the Eagle Formation ranges from about 600 to 800 ft thick. Most of the wells completed in the Eagle aquifer are within 300 ft of the surface and used for domestic and stockwater purposes. Reported well depths range up to more than 2,300 ft (153 Eagle wells are more than 1,000 ft deep); however, the median well depth is 170 ft. Most reported well yields are between 3 and 50 gpm (10th and 90th percentile of reported yields), with a median of 10 gpm.

Groundwater in the Eagle aquifer typically has elevated TDS concentrations and is only suitable for use near outcrop areas. The total dissolved solids concentrations (from 188 samples) range from about 115 to over 7,000 mg/L with a median of 987 mg/L. The potable to semi-potable water (TDS less than 2,000 mg/L) occurs mostly in outcrop areas where the aquifer is unconfined or where alluvial deposits overlie the aquifer. Low TDS water is generally Ca-Mg-SO_4 type water. Where the aquifer is confined by the Claggett Shale, the water becomes more mineralized and the chemistry changes to Na-SO_4-HCO_3 type water.

The Eagle Formation is an important gas producer south of Havre (Condon, 2000), and near the Bowdoin Dome (Noble and others, 1982).

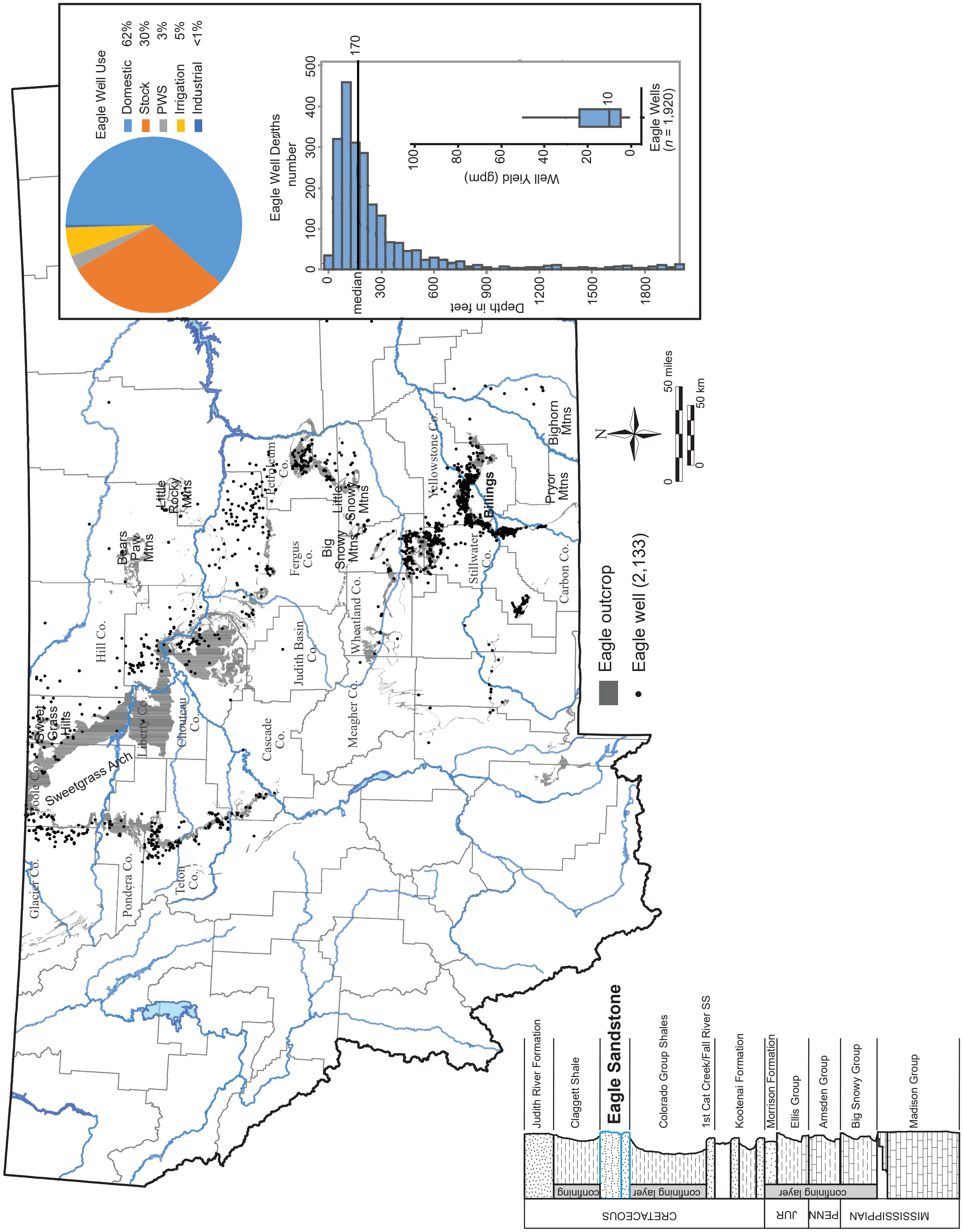


Figure 20. Distribution of wells in the Eagle sandstone aquifer.

The Claggett Shale (Kshale) disconformably overlies the Eagle Formation and serves as a regional confining unit. It is a dark marine shale 200–500 ft thick (Condon, 2000).

Judith River Aquifer

The Judith River Formation conformably overlies the Claggett Shale and crops out over a large part of north-central Montana (hydrogeologic map, fig. 1). The formation consists of eastward-thinning wedges of coastal-plain deposits of interbedded sandstone, shale, and coal. The strata are enclosed by the westward-pointing wedges of the Claggett and Bearpaw Shales (Feltis, 1982b). The sandstone beds are brown and light gray. Overall thickness ranges from less than 100 to 500 ft. Individual sandstone beds are generally less than 60 ft thick; the thickest sandstone beds occur near the base (Tuck, 1993).

Recharge is through infiltration of precipitation on outcrop and sub-crop areas, and stream loss across outcrops. Recharge may also occur through upward discharge from underlying, deep aquifers. Water in the Judith is unconfined in outcrop areas, but is generally under confined conditions at depth. In north-central Montana, between the Little Rocky Mountains and the Judith Mountains, the Judith River aquifer is strongly artesian. Flowing artesian wells are common in topographically low areas along the Musselshell and Missouri Rivers, and Telegraph Creek in southern Phillips County, and northern Fergus and Blaine Counties (Osterkamp, 1968).

The Judith River Formation is present throughout most of south-central Montana, but it is typically greater than 1,000 ft below the surface at distances more than 2 mi from the outcrop (Madison and others, 2014). Wells are concentrated in an arc paralleling the outcrop north and east of Billings (fig. 21). In north-central Montana most of the Judith River wells are between the Milk and Missouri Rivers, and peripheral to the Bowdoin Dome. Most wells are for domestic (55 percent) or stock use (41 percent). Reported well yields are typically between 3 and 30 gpm, with a median of 10 gpm; a few wells report yields up to 500 gpm. The overall median well depth is 156 ft; however, near the Missouri River in northern Fergus County and southern Phillips and Valley Counties, wells are typically more than 600 ft deep and range up to 1,900 ft deep. Flowing artesian wells are common in this area.

Water quality in the Judith River is generally mineralized; TDS concentrations range from 123 to more than 8,000 mg/L, and the median concentration from 183 samples is 1,770 mg/L. Most of the sampled water is a Na-HCO_3-SO_4 type. In general the more mineralized water (higher TDS) is enriched in Na and SO_4. Higher quality, lower TDS water occurs near recharge areas.

The Judith River Formation is overlain by the upper Cretaceous Pierre/Bearpaw Shale that ranges from 2,000 to 3,000 ft thick (Smith, 1999a; Miller, 1979). The Pierre/Bearpaw Shale (KSHLE) is a marine claystone and shale, and forms a regional confining unit that separates the Judith River Formation from the Fox Hills Sandstone.

Fox Hills–Hell Creek Aquifer

Sandstone beds of the Upper Cretaceous Fox Hills Sandstone and the lower part of the Hell Creek Formation are hydraulically connected and form an extensive aquifer that is widely used in eastern Montana.

The Fox Hills Formation contains interbedded fine- and medium-grained sandstone, sandy shale, siltstone, and minor carbonaceous shale. The unit was deposited as the last inland sea retreated northeastward out of Montana during the closing of the Cretaceous period. A white sandstone bed in the upper part of the formation, the Colgate Member, forms a distinctive cliff along the flanks of the Cedar Creek Anticline southeast of Glendive (Smith, 1999b). Sandstone units within the Fox Hills Formation range from less than 100 to up to 350 ft thick. They are the most extensive in southeast Montana and thin to the north and west (Smith, 1999a; Feltis 1982c). The Fox Hills is exposed along the margins of the Powder River and Williston basins and in narrow bands around the Cedar Creek Anticline, Poplar and Porcupine Domes, and the Black Hills Uplift (fig. 1).

The overlying Hell Creek Formation is composed of silty shale, mudstone, fine- and medium-grained sandstone, and a few thin coal seams. The Hell Creek accumulated by stream deposition in laterally migrating channel belts and on floodplains and is as much as 600 ft thick. The upper two-thirds of the formation is primarily composed of mudstone and acts as a confining bed; the lower third contains sandstone beds up to 100 ft thick (Smith, 1999a). In south-central Montana, the Lance Formation is fine-grained sandstone with interbeds of shale and a few thin coals (Wilde and Por-

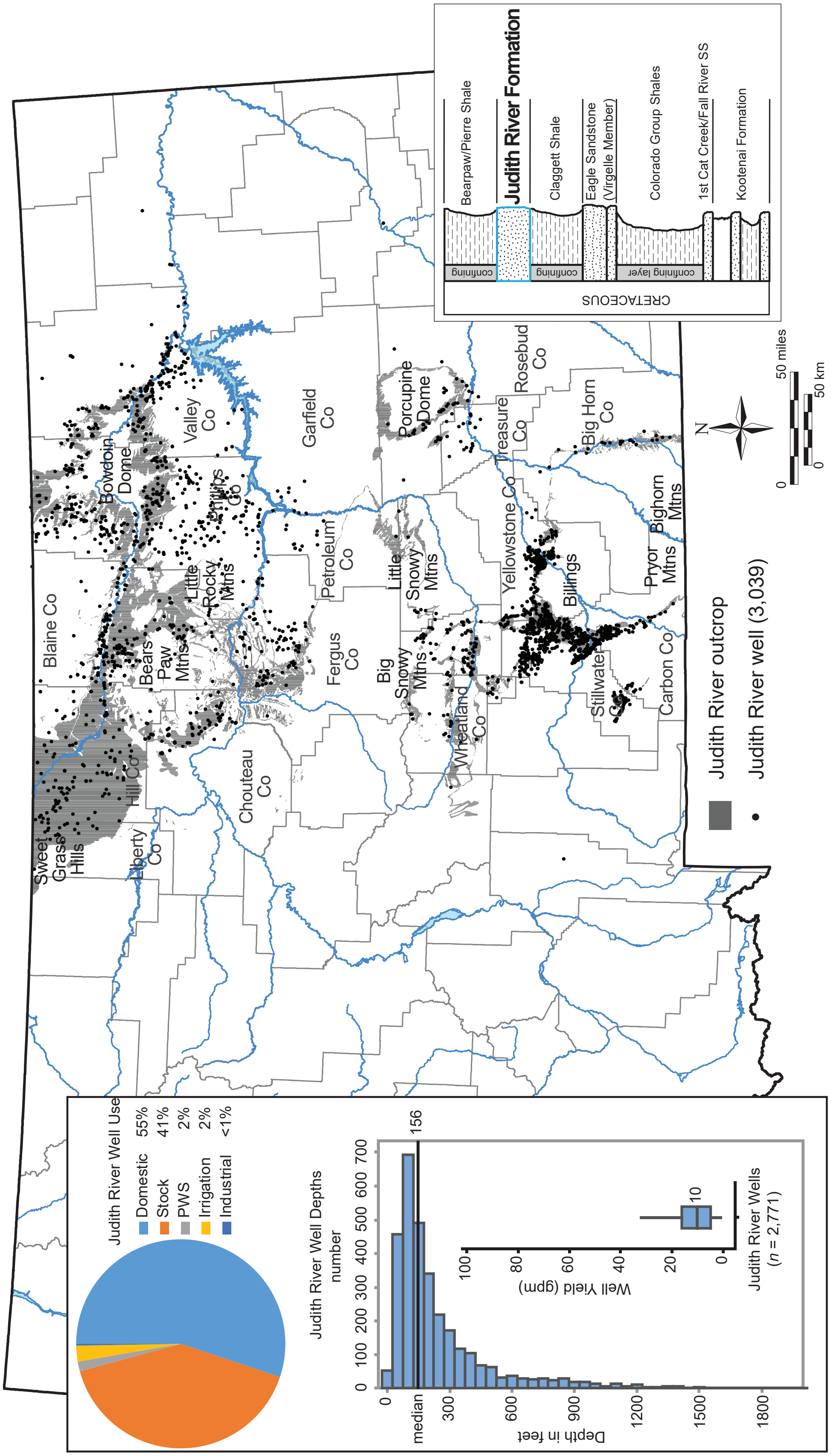


Figure 21. Distribution of wells in the Judith River sandstone aquifer.

ter, 2000; Lopez, 2001). The Lance is defined in Wyoming and southern Montana, and is equivalent to the Hell Creek Formation of central and eastern Montana (Fox, 1993). In general, the Lance Formation contains more sand and less shale than Hell Creek, and thickens to the west. The Lance Formation is considered part of the Hell Creek Formation for this report.

The hydraulically connected sandstone beds in the Fox Hills and overlying Hell Creek Formations form one aquifer that underlies the eastern quarter of Montana. Wells in the Fox Hills–Hell Creek aquifer are concentrated near outcrop areas and in the lower Yellowstone River valley (fig. 22). Flowing wells are common in the Yellowstone River valley between Glendive and Sidney, and the area off the southwest limb of the Cedar Creek anticline (Smith and others, 1999).

The Fox Hills–Hell Creek aquifer occurs at depths of 600 to 1,600 ft below land surface throughout most of eastern Montana except near the Cedar Creek Anticline, the Poplar Dome, and the Larb Hills in northern Blaine County (Smith, 1999b). Mudstones in the Hell Creek Formation confine the top of the aquifer, and the Bearpaw/Pierre Shale confines the base of the aquifer. In the lower Yellowstone River valley, regional flow in the Fox Hills–Hell Creek aquifer is toward the Yellowstone River, generally parallel to the axis of the Cedar Creek Anticline (LaFave, 1999). In northeastern Montana, regional flow is toward the Missouri River (Levings, 1982b). In topographically high areas, recharge occurs by downward leakage from overlying aquifers and through the fine-grained part of the Hell Creek Formation. Groundwater discharges from the aquifer to wells and in topographically lower areas, and by upward leakage to shallower aquifers. Industrial withdrawals in the early 1960s and uncontrolled discharge from flowing wells have resulted in a persistent loss of artesian pressure in the aquifer.

In the lower Yellowstone River area, about 1,500 wells are completed in the Fox Hills–Hell Creek aquifer (fig. 23). The widespread use has resulted in persistent water-level declines, especially in the lower Yellowstone River valley. The hydrograph from an observation well near Terry (fig. 23, well 1846) shows declining water levels—about 25 ft during the past 33 yr. Long-term declines occur when more water is removed from the aquifer than is recharged. At some point these declines can create undesirable effects such as increased pumping-lift costs, decreased yields, and causing flowing wells to cease flowing.

Overpumping the Fox Hills–Hell Creek aquifer resulted in Montana's first controlled groundwater area. In the early 1960s at the South Pine oil field near Baker, the Fox Hills–Hell Creek aquifer was pumped at about 450 gpm to support secondary oil recovery. The withdrawals caused water-level declines that affected surrounding stock and domestic wells. In response to landowner complaints, the South Pine Controlled Groundwater Area was created in 1967 to limit Fox Hills–Hell Creek aquifer pumping and slow the rate of water-level decline. Between 1975 and 1977 the industrial water supply wells were phased out of production, and water levels in the aquifer began to recover.

The long-term hydrograph from an observation well (fig. 23, well 136642) located within the controlled groundwater area shows that between 1962 and 1967 pumping caused the water level to drop more than 130 ft. After the controlled groundwater area was established and industrial pumping reduced, the rate of water-level decline slowed considerably—dropping about 20 ft after 1967. After the pumping was completely phased out, the water level rose about 110 ft, but stabilized about 40 ft below the 1962 level. The failure to fully recover may likely be related to the other overdrafts that are creating the declines observed near Terry.

Well depths in the Fox Hills–Hell Creek aquifer range to more than 1,800 ft deep; the median well depth is 200 ft. Well yields are generally less than 20 gpm; however, 38 wells reported yields between 200 and 800 gpm.

The water in the Fox Hills–Hell Creek aquifer is slightly mineralized but consistent in composition. The TDS concentrations range from about 600 to 1,400 with a median of 1,000 mg/L (399 samples); Na and HCO_3 are the major ions in solution.

Fort Union Aquifer

The early Tertiary (Paleocene) Fort Union Formation is exposed over the eastern third of Montana and is an important source of domestic and stockwater. During the late Cretaceous–early Tertiary, streams draining upland areas to the west deposited sediments in a fluvial to deltaic/estuarine environment (Downey and Dinwiddie, 1988). The resulting formation contains alternating beds of fine- to medium-grained sandstone, siltstone, mudstone, lignite coal, and a few thin limestone beds. The deposits are characterized

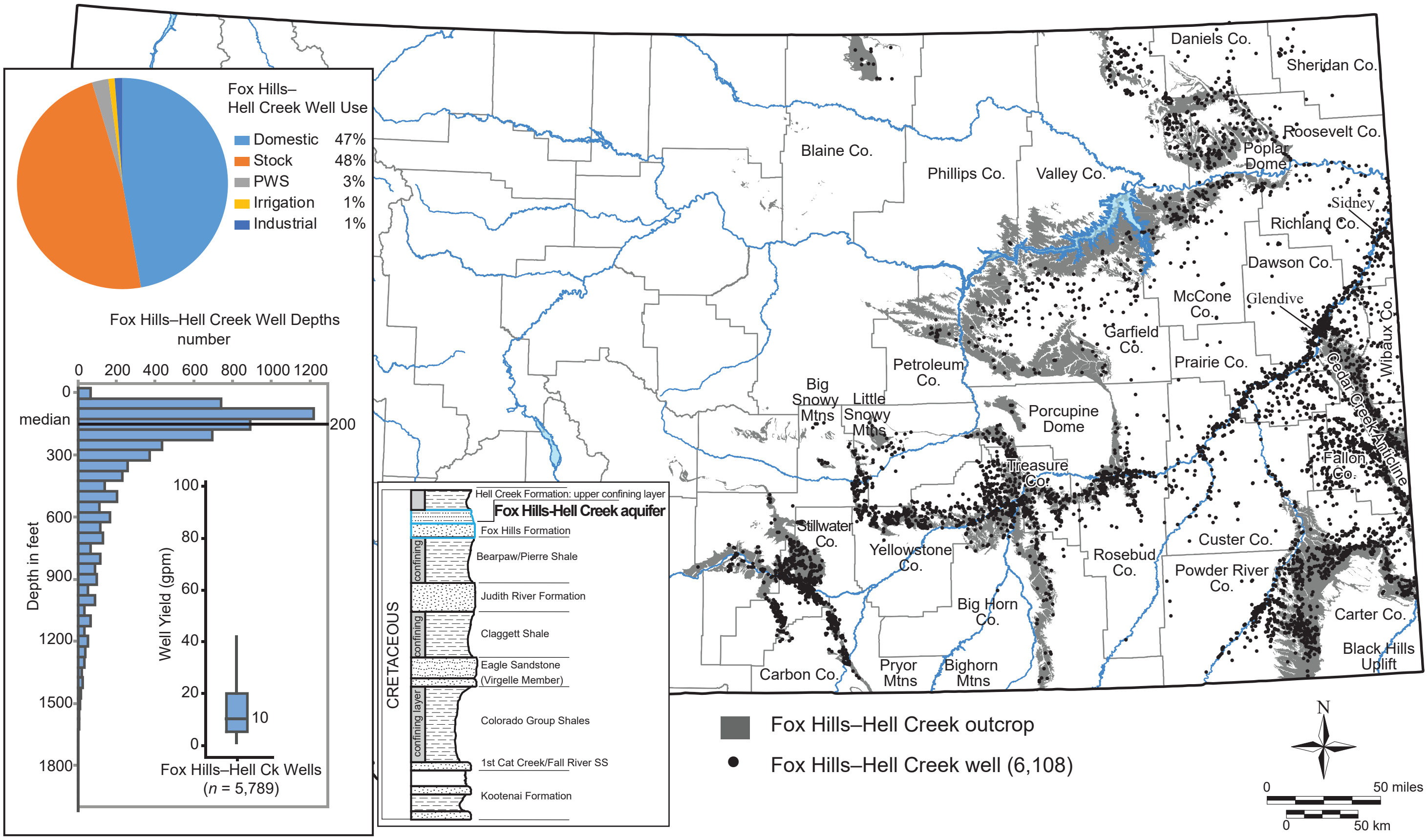


Figure 22. Distribution of wells in the Fox Hills–Hell Creek aquifer.

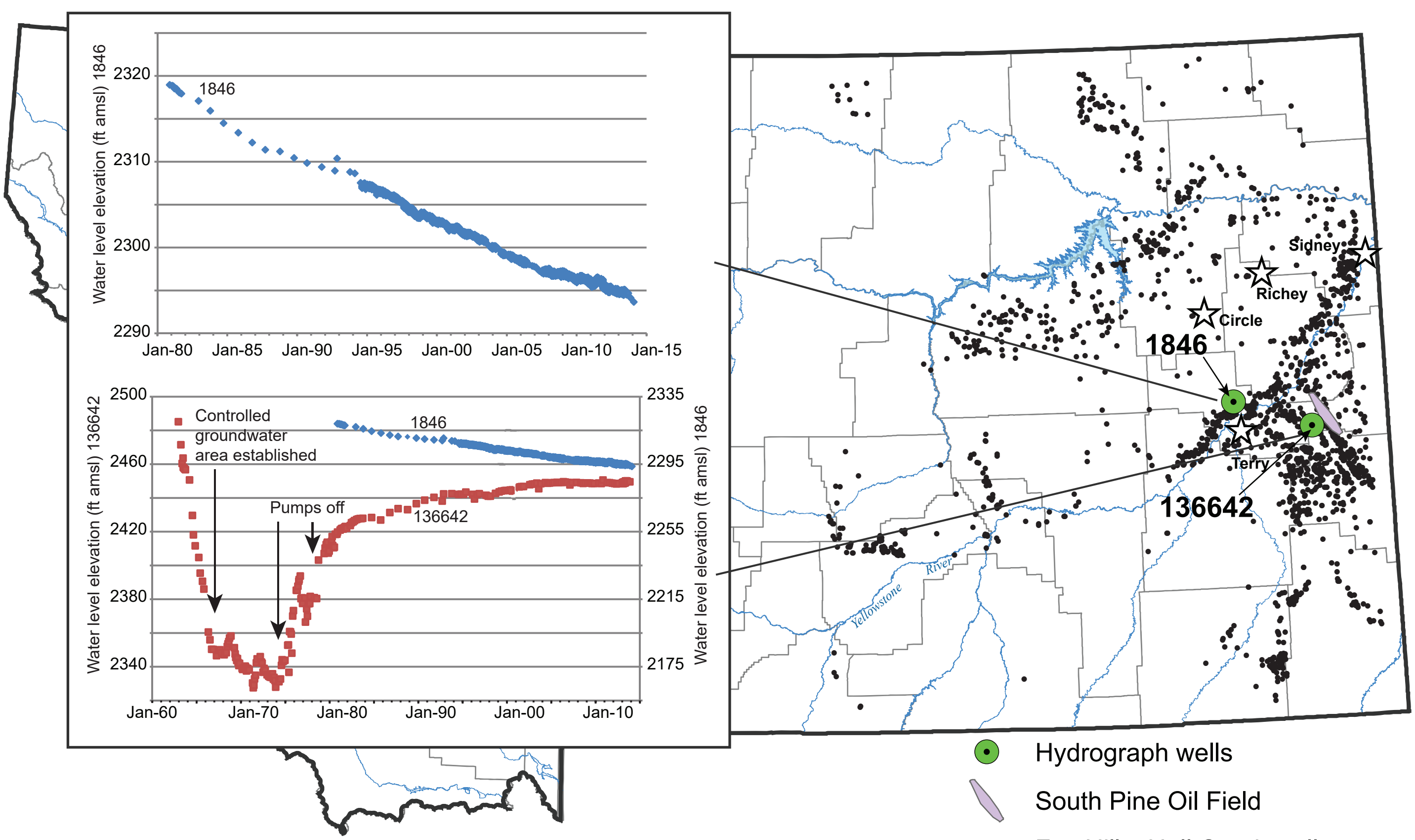


Figure 23. Water levels in the Fox Hills–Hell Creek aquifer near Terry are declining at a rate of about 1 ft per year. Near the South Pine field, water levels have recovered since industrial pumping ceased; they are still 40 ft lower than 1960s levels.

by lenticular beds, truncated units, and abrupt facies changes. Near major stream drainages, the Fort Union has been eroded into distinctive badlands topography.

The Fort Union Formation has been divided into three primary members; from oldest to youngest, they are the Tullock, Lebo Shale, and Tongue River (Vuke and others, 2007). The basal Tullock Member is composed of sandstone with dark shale, and interbeds of siltstone and thin coalbeds; the Lebo Member is a gray shale with interbeds of siltstone and coal; and the Tongue River Member is composed of alternating sandstone, siltstone, shale, and numerous thick and extensive coalbeds. Many Tongue River coal seams have burned to form red-orange, highly fractured clinker beds. Clinker is resistant to erosion and highly permeable and may extend as much as a mile into the subsurface (Lee, 1981), making it a ready conduit for groundwater recharge. The average thickness of the Fort Union is more than 1,000 ft.

Aquifers within the Fort Union Formation occur in sandstone beds and coal. The water-bearing sandstone and coal units are interbedded with shale and mudstone, resulting in a great deal of vertical and horizontal anisotropy. At a regional scale, the Fort Union Formation includes two aquifers: (1) a shallow unit including unconfined or semi-confined aquifers within about 200 ft of the land surface; and (2) a deeper unit, below about 200 ft, that consists of a locally confined aquifer(s). The shallow system generally coincides with the upper part of the Tongue River Member, whereas the deep system is associated with the Tullock Member. Differences in groundwater flow and water quality distinguish the two units. Groundwater flow paths in the shallow, unconfined system are relatively short and extend from local drainage divides to nearby valley bottoms. Flow paths in the deeper unit typically extend from regional drainage divides to regional topographic lows (Smith and others, 1999; Patton and others, 1999; LaFave, 1999; Thamke and others, 2014).

Recharge areas generally coincide with topographic highs; infiltration of rainfall and snowmelt recharges the shallow water-table aquifers. Deep aquifers within the Fort Union Formation are recharged by downward leakage from overlying aquifers. Clinker beds, because of their high permeability and position along ridge tops, serve as important recharge areas.

Groundwater use from the Fort Union is widespread in eastern Montana (fig. 24). Most of the wells are reportedly used for stock watering (52 percent) and

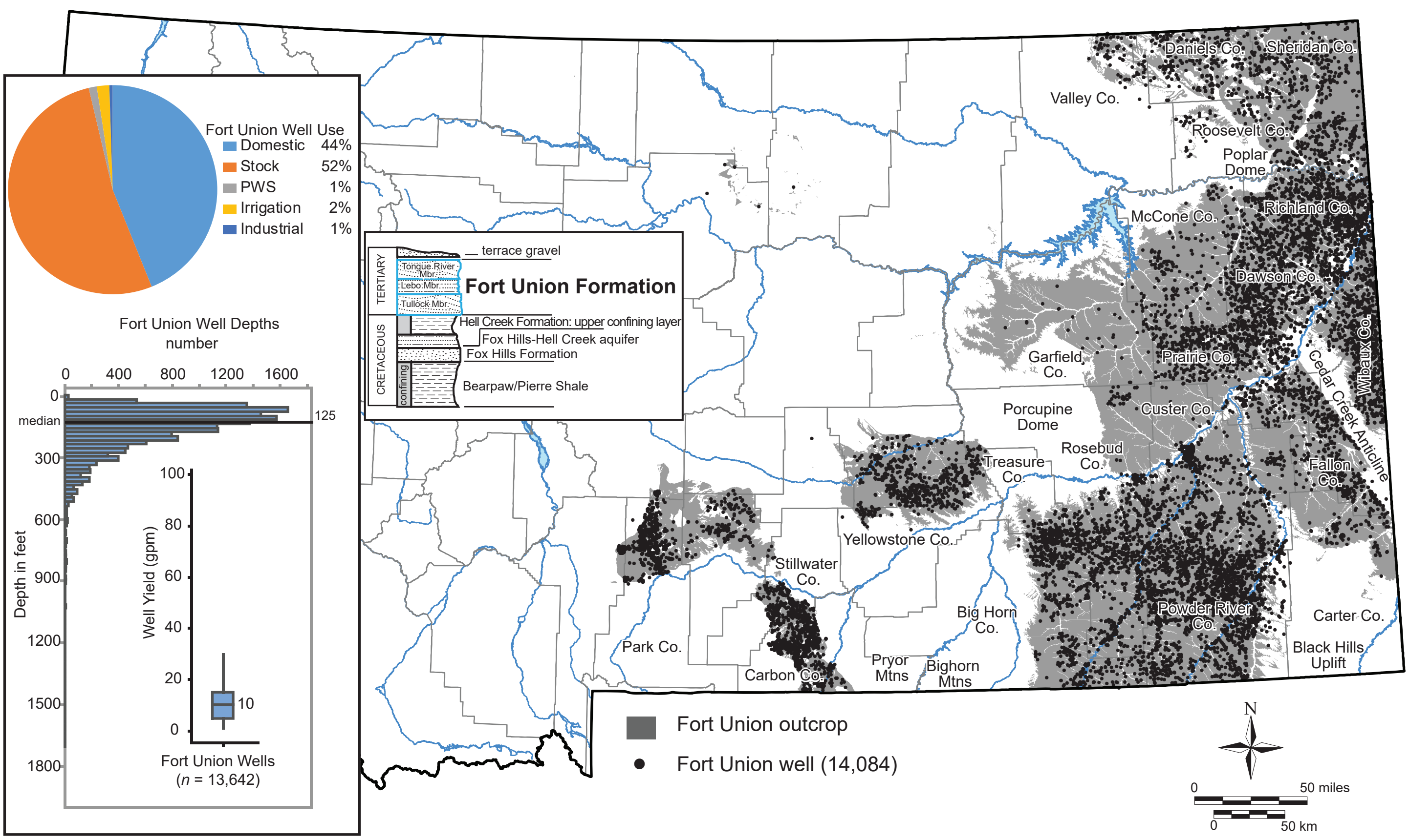


Figure 24. Distribution of wells in the Fort Union aquifer.

domestic (44 percent) purposes. Well depths generally are between 70 and 300 ft deep with a median of 125 ft. Well yields are typically low. Although a few wells have reported yields greater than 100 gpm, most are between 5 and 15 gpm, with a median of 10 gpm.

Water quality in the Fort Union is mineralized and varies as a function of depth, distance from the outcrop, and lithology (Lee, 1981; Smith and others, 1999). Water quality in the shallow Tongue River unit is variable, with TDS concentrations less than the deep unit. The median TDS from wells less than 200 ft deep is 1,050 mg/L (334 samples); the water is a mixed Na-Ca-Mg-HCO_3 type. As water moves deeper into the formation, dissolution and cation-exchange reactions change the composition to a Na-HCO_3 type water (Smith and others, 1999); the median TDS concentration for wells greater than 200 ft deep is 1,300 mg/L.

Alluvial Aquifers—Eastern Montana

Alluvial valleys along the Yellowstone and Missouri Rivers and their tributaries are a major source of water in the Great Plains region. These aquifer systems are generally narrow, restricted to the width of the alluvial valley. The alluvial aquifers consist of Quaternary and Tertiary sand and gravel deposits interbedded with silt and clay that occur in the river valley floodplains, and discontinuous low-level terrace gravels that flank the rivers and many mountain fronts. Alluvium is thickest along the Missouri and Yellowstone Rivers and is present along major tributary streams. Aquifer thickness generally increases in the downstream direction. Across most of the Great Plains region, alluvial valley aquifers are incised into and are underlain by relatively impermeable units—mostly Cretaceous shales and, in places, Cretaceous and Tertiary sandstones. The width of alluvial valleys is commonly controlled by the bedrock lithology as seen in the Yellowstone River valley. Upstream from Park City (20 mi southwest of Billings), the Yellowstone has cut through relatively resistant Cretaceous sandstones (Judith River and Eagle) and the valley width is less than 1 mi wide (hydrogeologic map, fig. 1). Downstream from Park City, where less-resistant Colorado shale is present, the valley is as much as 9 mi wide and the river is flanked by a series of well-established terraces (Lopez, 2000; Gosling and Pashley, 1973).

Alluvial aquifers are also developed in Wisconsinan glacial sediments north of the Missouri River (fig. 25). The major aquifers occur in glacial outwash and buried valley deposits. Buried valley aquifers are the product of Pleistocene glaciation. The most productive of these aquifers occur in areas where the pre-glacial Missouri River channel was overridden by glacial ice and subsequently "buried." The main buried channel aquifers in Montana occur near Sidney and in Sheridan County in northeastern Montana, south of Great Falls in central Montana, and from Big Sandy to nearly Havre in north-central Montana.

In the alluvial aquifers associated with the Yellowstone and Missouri Rivers, the rivers are the principal control on the groundwater flow systems. The rivers are gaining and represent regional groundwater discharge zones. Groundwater discharges from the alluvium principally as baseflow and evapotranspiration; minor quantities are discharged to wells. Recharge to the alluvial aquifers is principally from precipitation, irrigation water—either as canal leakage or return flows from applied irrigation water—and high-stage flows. Adjacent bedrock aquifers and tributary gravels contribute lesser amounts of recharge.

Wells completed in the Quaternary–Tertiary alluvium of eastern Montana are used primarily to meet domestic or stockwater needs (70 and 20 percent, respectively; fig. 25). Reported well yields are typically between about 15 and 40 gpm, with a median of 23 gpm. About 250 of 18,300 wells have reported yields greater than 500 gpm and are used for irrigation or public water supply. The overall median well depth is 35 ft.

Water in the eastern alluvial aquifers is generally more mineralized than in western basin-fill aquifers; TDS concentrations mostly range between 350 and 1,300 mg/L with a median of 711 mg/L. The water is a mixed Ca-Mg-Na-HCO_3 type (1,902 samples). Where alluvial deposits are in contact with Cretaceous shales, the TDS concentrations are generally higher and the water is more enriched in Na and SO_4.

TRACKING MONTANA'S GROUNDWATER

The Montana Bureau of Mines and Geology's (MBMG) Ground Water Monitoring Program collects water-level measurements from strategically located wells across the State. Long-term groundwater-level records are the only direct measure of how aquifers respond to seasonal, climatic, developmental, or land-use changes. Long-term groundwater hydrographs are similar to long-term records of streamflow and precipitation, and must be evaluated at decadal scales (LaFave, 2014).

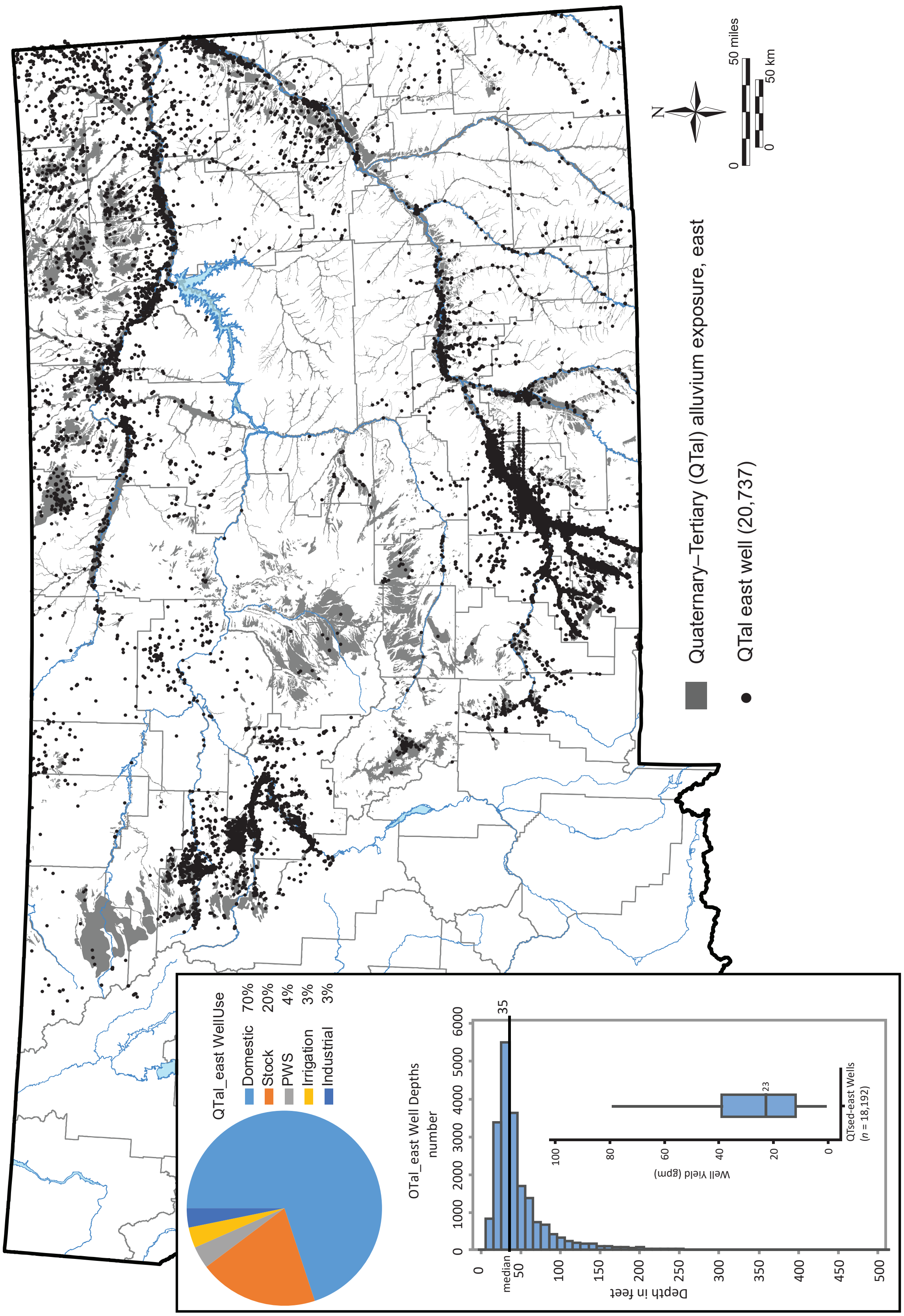


Figure 25. Distribution of wells in the eastern alluvial aquifers.

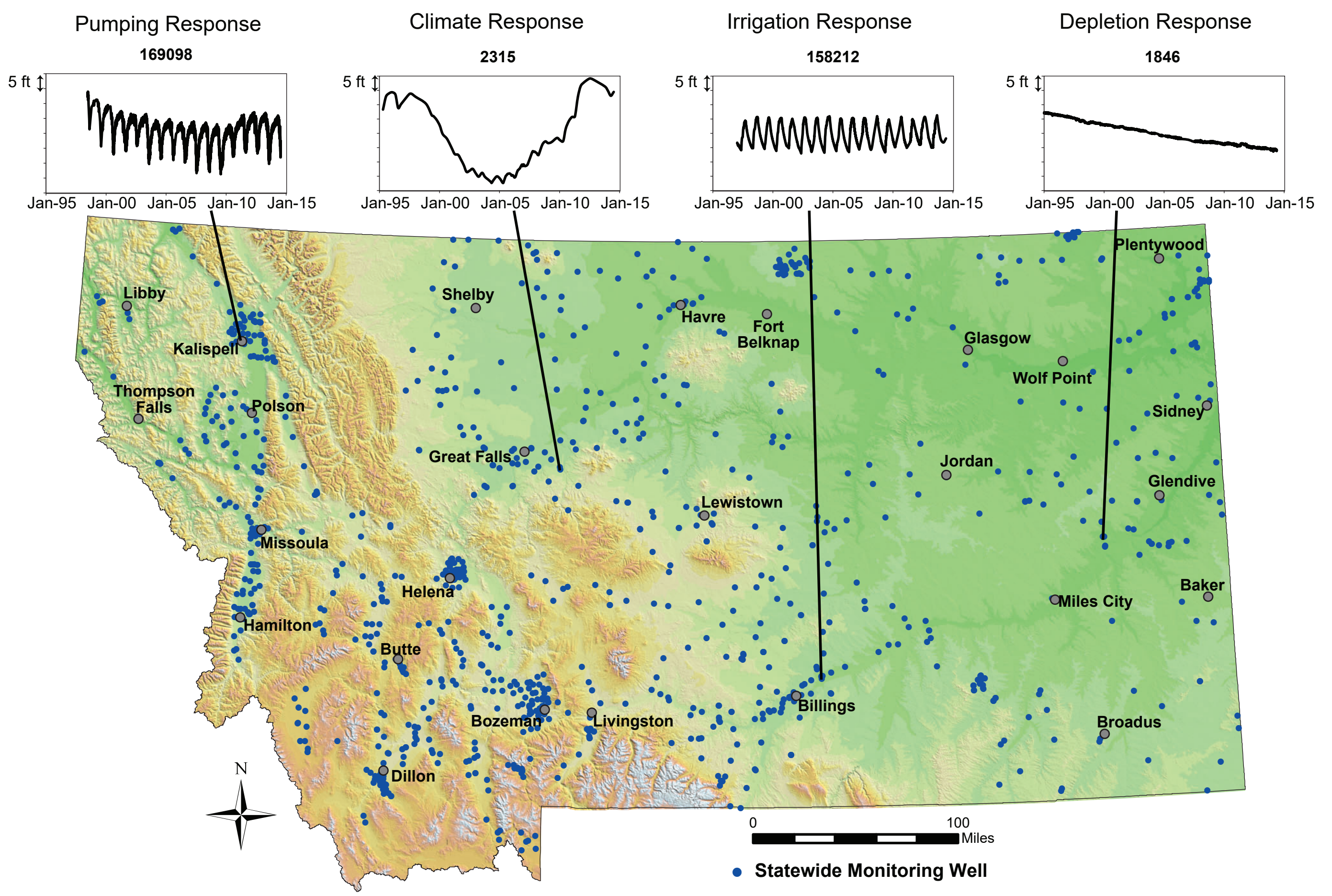


Figure 26. Groundwater levels are tracked in more than 800 wells across Montana. Differences in recharge, development, and land use result in different water-level responses. Systematic long-term water-level measurements provide a basis to assess groundwater storage changes and how different stresses may impact aquifer systems.

Since 1993, the MBMG has been collecting groundwater-level data systematically from more than 800 wells (fig. 26); some have been regularly monitored since the 1950s. The monitoring network covers the State's major aquifers and includes wells that range from less than 10 to more than 1,600 ft deep.

Water levels in most Montana aquifers follow natural seasonal patterns, typically rising each spring and early summer, and declining during the late summer and fall. In addition to the seasonal changes, water levels respond to other stresses such as pumping (response in hours or days), climate variability (response in years to decades), and widespread development (response occurs at varying time scales). Montana's long-term monitoring network is now showing where and which aquifers are impacted by these different stresses, highlighting the value of long-term, decadal-length records. Without continued monitoring, Montanans would have only limited, antiquated data to address these important issues.

For information on Montana's groundwater resources, visit the Ground Water Information Center website: http://mbmggwic.mtech.edu/

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