MONTANA STATE UNIVERSITY

Hydrogeology and geothermal potential of the Radersburg Valley, Broadwater County, Montana
by Glen Milton Wyatt

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Earth Sciences
Montana State University
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Abstract:
Two major aquifers occur in the Radersburg valley in southwest Montana. Irrigation and other wells discharge water from a Cenozoic basin-fill aquifer composed of up to 1000 feet of sand and gravel. During the 1979 growing season, irrigation wells withdrew about 300 million cubic feet of water from the aquifer. The Madison Group, composed of Mississippian-age limestone, is the other major aquifer. The yearly discharge from springs flowing from the Madison Group at Plunket Lake, in the southern part of the valley, approximately equals the total amount of water withdrawn from the basin-fill aquifer for irrigation during 1979.

Much of the irrigation water is withdrawn from an area north of Plunket Lake. This area is near the projected north-south subcrop trend of the Madison Group aquifer and suggests that the aquifer recharges the basin-fill. Ground-water temperatures and chemistry further support this idea. Ground-water temperatures in this area vary from less than 10°C to about 20°C. The highest temperatures are colinear with the Madison Group subcrop trend, implying deep circulation within the limestone aquifer before the water discharges to basin fill.

Ground-water compositions are generally dominated by calcium, sodium, bicarbonate, and sulfate ions. Sulfate to bicarbonate ratios are highest north of Plunket Lake and indicate that a sulfate-rich source of ground water is locally present. A possible source of such ground water could be remnant evaporite beds within the upper part of the Madison Group. Another indicator of Madison Group discharge to the basin fill is the ratio of strontium to total dissolved solids. High values for the ratio may indicate recharge by a carbonate or evaporite aquifer to an aquifer composed of clastic materials. The highest values for the ratio were obtained from water samples in the vicinity of the projected subcrop trend of the Madison Group.

A possible consequence of the use of an aquifer for irrigation may be the permanent depletion of the ground-water resource. This depletion from aquifer storage is indicated by a continuing drop in ground-water levels. Irrigation by ground water from the basin-fill aquifer began in. the early 1960s. Since that time, water levels have generally declined by about 0.4 foot per year. The amount of water annually lost from aquifer storage as indicated by this average decline is less than one percent of the amount pumped for irrigation in 1979.

Ground water in the valley has a slight geothermal potential, as indicated by ground-water temperatures above 16°C. Water from such low-temperature thermal wells or springs may be used for space heating, greenhouses, or other low-temperature applications.

HYDROGEOLOGY AND GEOTHERMAL POTENTIAL OF THE RADERSBURG VALLEY

BROADWATER COUNTY, MONTANA

by

GLEN MILTON WYATT

A thesis submitted in partial fulfillment
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MONTANA STATE UNIVERSITY
Bozeman, Montana

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ABSTRACT

Two major aquifers occur in the Radersburg valley in southwest Montana. Irrigation and other wells discharge water from a Cenozoic basin-fill aquifer composed of up to 1000 feet of sand and gravel. During the 1979 growing season, irrigation wells withdrew about 300 million cubic feet of water from the aquifer. The Madison Group, composed of Mississippian-age limestone, is the other major aquifer. The yearly discharge from springs flowing from the Madison Group at Plunket Lake, in the southern part of the valley, approximately equals the total amount of water withdrawn from the basin-fill aquifer for irrigation during 1979.

Much of the irrigation water is withdrawn from an area north of Plunket Lake. This area is near the projected north-south subcrop trend of the Madison Group aquifer and suggests that the aquifer recharges the basin-fill. Ground-water temperatures and chemistry further support this idea. Ground-water temperatures in this area vary from less than 10°C to about 20°C. The highest temperatures are colinear with the Madison Group subcrop trend, implying deep circulation within the limestone aquifer before the water discharges to basin fill.

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A possible consequence of the use of an aquifer for irrigation may be the permanent depletion of the ground-water resource. This depletion from aquifer storage is indicated by a continuing drop in ground-water levels. Irrigation by ground water from the basin-fill aquifer began in the early 1960s. Since that time, water levels have generally declined by about 0.4 foot per year. The amount of water annually lost from aquifer storage as indicated by this average decline is less than one percent of the amount pumped for irrigation in 1979.

Ground water in the valley has a slight geothermal potential, as indicated by ground-water temperatures above 16°C. Water from such low-temperature thermal wells or springs may be used for space heating, greenhouses, or other low-temperature applications.

INTRODUCTION

Shortly after the valley in the vicinity of Radersburg, Montana (Figure 1) was settled by farmers, a drought occurred. This drought lasted from about 1917 until the early 1920s and demonstrated the need for irrigation of farmlands in the area. Irrigation water was first supplied to farms by diverting water from springs and creeks. Beginning in 1941, irrigation water was supplied to the Crow Creek area by a system of canals from the Missouri River. Operation of another canal system supplying water to land south of Crow Creek began in 1955. In 1955, approximately twenty-one square miles of land were irrigated by canal systems in this area.

Irrigation water has also been supplied to part of the valley by high-capacity wells. By 1979, irrigation wells were supplying water to about seven square miles of land which had previously been irrigated by canals and to about three square miles which had been dryland farmed. Twenty-seven irrigation wells pumped from 1,000 to 3,000 gallons per minute (gpm) or 134 to 401 cubic feet per second (cfs) during the summer months and discharged over three hundred million gallons of water. Future withdrawals must be balanced by adequate recharge to the aquifer to allow irrigation by wells to continue.

The irrigation wells have been completed in basin-fill sediments which overlie older bedrock units. Many of the wells have been installed in the Warm Springs Creek area north of Plunket Lake. Plunket

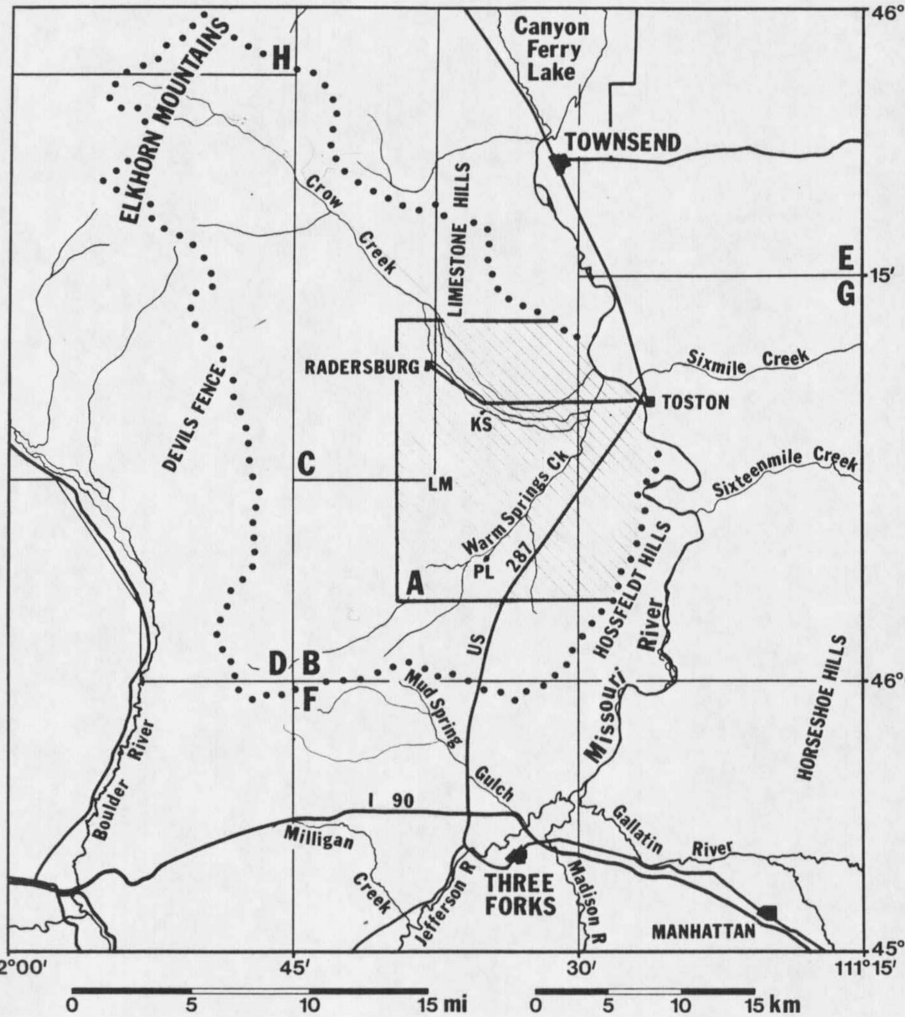
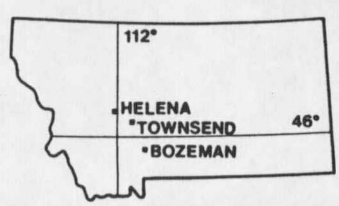


Figure 1. Location of the study area (A, shaded), Plunket Lake (PL), Kimpton Spring (KS), Lone Mountain (LM), the Radersburg drainage basin (dotted), other areas delineate published maps at a scale of 1:62,500 or greater:
(B) Freeman and others (1958),
(C) Klepper and others (1971),
(D) Klepper and others (1957),
(E) Nelson (1963),
(F) Robinson (1963),
(G) Robinson (1967), and
(H) Smedes (1966)



Lake and Warm Springs Creek are fed by water issuing from Mississippian Madison Group limestone at Plunket Spring. The presence of the spring flowing from the limestone indicates that water from the Madison Group is recharging the basin-fill sediments in at least part of the valley.

The temperature of ground water flowing from Plunket Spring and Kimpton Spring was about 16°C to 18°C. This range was at least 10°C above the mean annual air temperature of 43°F (6°C), which enabled these springs to be classified as low-temperature thermal springs (Muffler, 1979). The presence of thermal springs in the Radersburg valley suggested that ground water in the area may be a potential source of energy.

Purpose and Scope

This study has three objectives. The first was to describe the hydrogeology of the basin-fill and Madison Group aquifers in the Radersburg drainage basin. The second objective was to determine if the amount of ground water discharged from the basin-fill aquifer is significantly greater than recharge to the aquifer. The final objective was to determine the temperature, areal extent and geologic source, and potential for development of the geothermal resource in the valley.

Work was divided into two phases. In the first phase of the investigation a compilation was made of water-well logs, ground-water appropriation documents, and land ownership for the study area. The second phase was a field investigation consisting of locating wells; measuring water levels, discharge rates, and water temperatures; assessing the hydraulic properties of the basin-fill materials; and water-quality sampling.

The main study area consisted of land west of the Missouri River, within Townships 4 and 5 North and Ranges 1 and 2 East, in Broadwater County, Montana. The study area was limited to land west of the Missouri River because the river is a master river and because extensive development of irrigation by canal systems and high-capacity irrigation wells has occurred in this area. Warm-water springs also occur here.

Site Numbering System

Wells and other sampling sites were numbered according to the U.S. Bureau of Land Management land subdivision system. All sites are north of the Montana baseline and east of the principal meridian, in quadrant "A" of Figure 2, except the Crow Creek sampling site which is west of the principal meridian in the "B" quadrant. Therefore, the letter designation for quadrant was not used. The township in which a site is located is designated by the first number, either a 4 or 5, for Township 4 or 5 North. The second numeral identifies the range, either a 1 or 2, for Range 1 or 2 East. The third numeral identifies the section. Lower-case letters after the numerals indicate in which part of a section the site is located. The letters begin in the northeast quarter of a section with "a" and end with "d" in the southeast quarter. Parts of quarter sections are similarly divided. If two

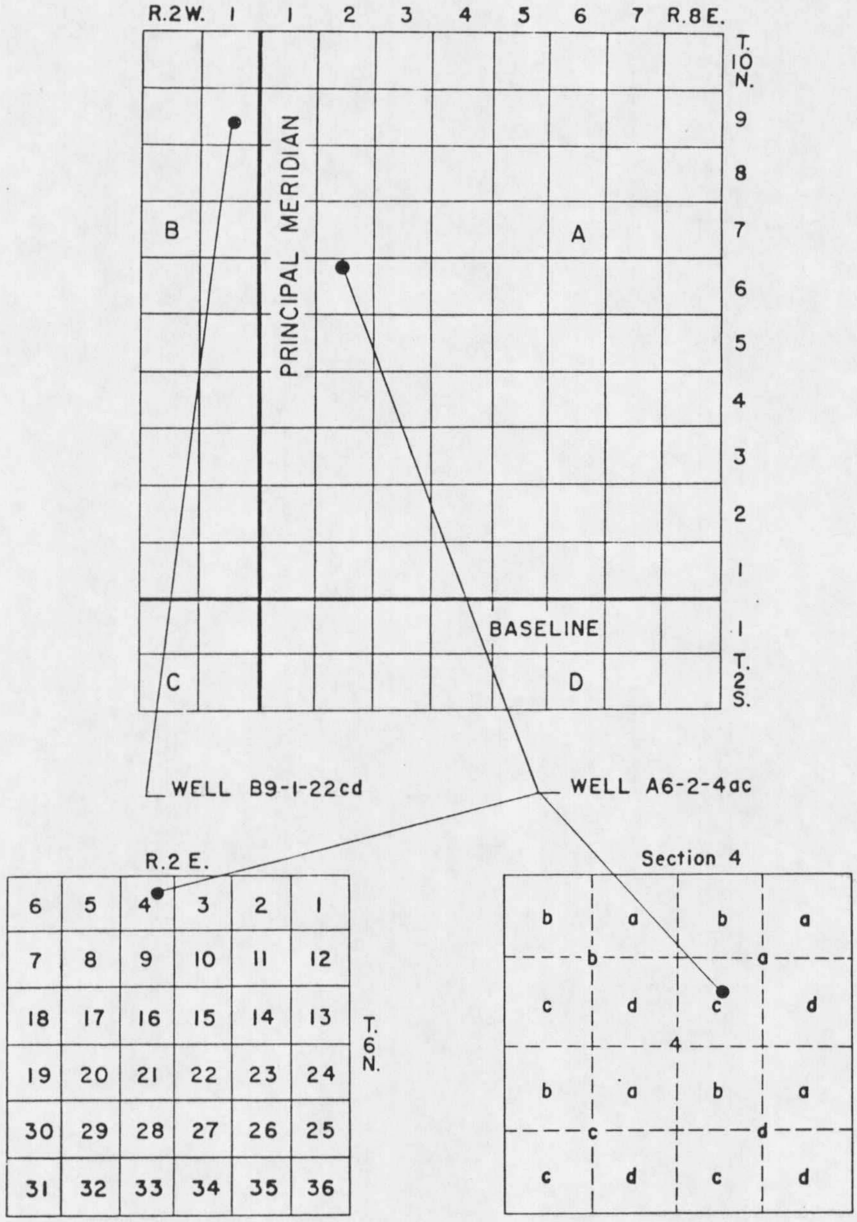


Figure 2. U.S. Bureau of Land Management Land Subdivision System
(Source: Lorenz and McMurtrey, 1956)

wells are located within the same quarter-quarter-quarter section, a serial number is added after the last letter. This system of site location should be considered as accurate to within 0.1 mile, as locations were determined from section corners using a car odometer or by pace.

DRAINAGE BASIN DESCRIPTION

Introduction

To assist with identification of possible recharge sources to irrigation wells, the drainage basin containing irrigated land in the Radersburg valley was identified on the basis of topography. The drainage basin outlines tributary areas of the Missouri River between the Limestone Hills and Hossfeldt Hills and includes the Crow Creek and Warm Springs Creek drainages. The identification of the drainage basin provided limiting boundaries of surface-water and ground-water flow through the Radersburg valley.

Regional Setting and Topography

The Radersburg valley is defined for the purposes of this report as the area below elevation 5000 feet west of the Missouri River and Hossfeldt Hills, south of the Limestone Hills, and east of the Devils Fence in the Radersburg drainage basin. The valley has an undulating surface which slopes to the east. Local relief is generally less than 150 feet, although 5024-feet-high Lone Mountain rises over 600 feet above the neighboring lowlands which slope to about 3880 feet elevation near the Missouri River. The study area is entirely within the Radersburg valley and is the only part of the valley where high-capacity irrigation wells have been drilled.

The Radersburg drainage basin, as shown on Plate 1 and Figure 1, occupies about 400 square miles in the north-central part of the Northern Rocky Mountains physiographic province of Fenneman (1931). The basin is outlined by a drainage divide and the Missouri River. The divide is a broad low rim which extends south from the Missouri River through the Hossfeldt Hills, curves to the west, and turns north through the Devils Fence area. The divide continues along the ridges of the Elkhorn Mountains, extends through the Limestone Hills to the east, and reconnects with the Missouri River. The highest point along the divide and in the basin is 9414-foot Crow Peak, and the lowest point is at about 3870 feet where the divide meets the Missouri River east of the Limestone Hills.

The surface-water and ground-water boundaries of the basin are assumed to be coincident. This assumption appears to be valid because the surface-water divide extends along the tops of mountain ranges or hills throughout much of its length. Drainage divides are usually ground-water boundaries (Toth, 1963), therefore it is unlikely that large volumes of ground water cross the divide. A small amount of ground water may locally pass under the divide because of geologic structure. In such local areas, bedrock under the divide dips into the drainage basin. This condition is of limited areal extent compared to the area of the drainage basin. The amount of ground water which may cross divides was assumed to be much smaller than the volume of ground water in the basin. It is also unlikely that the Missouri River is a source of recharge water to the basin because the river lies below the rest of the basin and ground water generally discharges into master rivers (Toth, 1963).

Drainage

All streams in the Radersburg basin flow into the Missouri River. Crow Creek and Warm Springs Creek, fed by Plunket Spring, form the only perennial system (Table 1). Crow Creek and irrigation ditches from it lose water by seepage into outwash fan sand and coarse gravel in the vicinity of Crow Creek and Indian Creek north of Radersburg (Klepper and others, 1971; Straw, 1979).

Other streams in the basin are ephemeral and show evidence of recent flash floods and mud flows (Straw, 1979). Some of the ephemeral streams have local perennial spring-fed sections. These springs are small and the flow disappears into the streambed alluvium after a short distance. An example of this was seen in Johnny Gulch in May, 1979 where a spring 0.5 mile west of the Kimpton Cow Camp discharged 0.05 cfs of water, flowed downstream for about one-quarter mile, and then was absorbed into alluvium.

Climate

The Radersburg valley has a modified continental climate. The frequent invasions of Pacific Ocean air masses, flow of cool air from surrounding mountains into the valley, and nearness of the mountains combine to make temperature changes less extreme than in a true continental climate. About 40 inches per year of precipitation (as

Table 1. Discharge Measurements from Crow Creek and Springs

Date	Flow Rate (cfs)	Remarks
	Crow Creek	
1919-1922 average	60.0	Colby and Oltman, 1938, in Lorenz and McMurtrey, 1956
1919-1929 average	48.3	Measured above Slim Sam Creek (Montana State Engineer, 1956)
7-14-20	1000	Do.
1-10-22	1.4	Do.
6-28-79	132.7	Measured at Helena National Forest boundary
7-9-79	60.4	Measured at Helena National Forest boundary
7-9-79	45.1	1.5 miles north of Radersburg
9-15-79	17.2	Measured at Helena National Forest boundary
	Plunket Spring	
May, 1922	9.7	Pardee, 1925
August, 1949	8.7	Lorenz and McMurtrey, 1956
5-26-79	0.3	Southeast spring, not including discharge below water surface
5-26-79	0.2	Culvert 0.25 mile north of Plunket Lake
6-26-79	0.3	Culvert as above
10-13-79	10.3	Culvert as above
11-10-79	11.5	Culvert as above
	Kimpton Spring	
July, 1921	0.2	Pardee, 1925
September, 1949	<0.2	Lorenz and McMurtrey, 1956
6-16-79	0.7	In stream 150 feet east of spring
	Spring (SW1/4 sec. 17, T. 5 N., R. 1 W.)	
5-19-79	0.05	0.5 mile west of Kimpton Cow Camp

rain and snow) falls on the local mountains, but the mountains create a rain shadow and less than 12 inches per year falls on parts of the valley floor. Snow usually begins accumulating in the mountains in November, reaches maximum depth and water content in April or early May, and melts mostly in late May or in June. About 200 inches of snow falls each year on the mountains but only approximately 30 inches of snow falls on the valley floor (Olsen and others, 1977). The length of the frost-free growing season in the valley varies from about 110 to 120 days from May to September (Montana State Engineer, 1956). Temperature and precipitation data for the area are shown in Table 2.

An estimate of the amount of evaporation from near-surface soils plus transpiration by plants (evapotranspiration) was made using Thornthwaite's (1948) formula for calculating potential evapotranspiration (PET). The estimate is given in Table 3. Potential evapotranspiration estimates the amount of water lost from a vegetated surface to the atmosphere where the soil water supply is unlimited (Dunne and Leopold, 1978). Thornthwaite's (1948) formula based only on mean monthly air temperature and insolation, or received solar energy, was used to calculate PET:

$$PET = 1.6b(10T/I)^a;$$

where b = the insolation factor, based on the 46° 20' N latitude of Townsend;

T = the mean monthly air temperature (°C);

I = the mean annual heat index, $I = (Ti/5)^{1.514}$, (i = 1-12); and

$a = 0.49 + 0.0179I - 0.0000771I^2 + 0.000000675I^3$.

Table 2. Climatic Data for Townsend and Toston
(Source: Olsen and others, 1977)

	Townsend						Toston	
	Average Daily Maximum (°F)	Average Daily Minimum (°F)	Average Monthly Maximum (°F)	Average Monthly Maximum (°F)	Precipitation Total (in)	Precipitation Snow (in)	Precipitation Total (in)	Precipitation Snow (in)
January	31	7	52	-20	0.5	6.6	0.4	6.0
February	38	14	56	-9	0.3	4.1	0.3	3.5
March	44	18	64	-6	0.5	4.2	0.7	4.6
April	57	28	76	14	0.7	1.7	1.0	3.6
May	66	37	84	24	1.8	1.0	1.8	0.2
June	73	44	89	32	2.5	Tr	2.7	0.2
July	84	48	95	35	1.1	0	1.1	0
August	83	46	95	35	1.1	0	1.1	0
September	71	38	88	24	1.2	0.3	1.0	0.4
October	60	29	79	16	0.7	0.8	0.7	1.2
November	44	20	63	-1	0.5	4.1	0.6	3.0
December	36	13	54	-13	0.5	5.7	0.4	7.3
Year	57	29	97	-25	11.6	28.5	11.8	30.0

Table 3. Potential Evapotranspiration

	Mean Monthly Air Temperature (°C)	(°F)	Insolation Factor	Potential Evapotranspiration (cm)	(in)
January	-7	19	0.78	---	---
February	-3	26	0.80	---	---
March	-1	31	1.02	---	---
April	6	43	1.13	3.6	1.4
May	11	52	1.29	7.8	3.1
June	15	59	1.31	10.9	4.3
July	19	66	1.33	14.2	5.6
August	18	65	1.22	12.5	4.9
September	13	55	1.05	7.3	2.9
October	7	45	0.93	3.6	1.4
November	0	32	0.78	---	---
December	-4	25	0.74	---	---
Year	6	43		59.8	23.6

Thorntwaite's method has been shown to underestimate PET in dry regions (Cruff and Thompson, 1967). Lake evaporation is approximately 28 inches per year in the Townsend area (Environmental Sciences Services Administration, 1968). Lake evaporation and PET may not be representative of the actual amount of evaporation from vegetated and non-vegetated soils because the soil dries and restricts evaporation. This drying occurs early in the year and invalidates the assumption of unlimited water supply. Nonetheless, PET and lake evaporation may place upper limits on the amount of water lost to the atmosphere.

Vegetation and Land Use

Vegetation within the basin is quite variable. Coniferous forests cover the mountains, deciduous and coniferous trees grow along streams, and sagebrush and grasses occur on the unfarmed valley floor and rangeland. Where dryland farming practices are used, winter wheat and barley are the main crops. Sugar beets, hay, alfalfa, corn, and sunflowers are the major crops produced on irrigated lands.

Before irrigation wells were drilled in the valley, all irrigation water was supplied by the diversion of streams and springs. Plate 1 shows the locations of some of the canals and ditches in the valley. As a result of a drought which began in 1917, the first irrigation district was formed in 1918 to develop irrigation in the Crow Creek area (Pardee, 1925; Montana State Engineer, 1956). This district was dissolved in 1940 after a period of tax delinquency. In 1941, water was first diverted from the Missouri River at the Toston Dam through the State-funded Toston Canal West Fork (or West Side Canal) to the Crow Creek area west of Toston. Records of water use for the irrigation seasons from 1941 through 1949 indicate that water was diverted at rates ranging from 1.7×10^8 cubic feet per year (ft^3/yr) to 2.6×10^8 ft^3/yr with a mean of 2.3×10^8 ft^3/yr of water diverted for irrigation (Montana State Engineer, 1956). In 1955, approximately three square miles of land were irrigated by water from this canal system (Montana State Engineer, 1956).

The federally-funded Crow Creek Irrigation Unit began operation in 1955 to develop irrigation on an area of land which approximately equaled the total acreage flooded by the creation of Canyon Ferry Lake near Townsend. Water for irrigation is pumped from the Missouri River at a point about two miles upstream of the Toston Dam, through an inverted siphon, and into the Toston Canal. Table 4 is a summary of water use by the Irrigation Unit from 1968 to 1979. During that period, from 6.3×10^8 ft^3/yr to 10.9×10^8 ft^3/yr with a mean of 8.4×10^8 ft^3/yr was diverted from the Missouri River for irrigation use. In 1955, approximately 2.25 square miles of land were being irrigated by water from the Crow Creek Irrigation Unit (Montana State Engineer, 1956). The amount of land irrigated by this system had increased to about nine square miles by the late 1970s.

Privately-owned ditch systems have also been used to divert water from streams and springs to farmlands. In 1955, approximately sixteen square miles were under irrigation by privately-owned systems (Montana State Engineer, 1956).

Table 4. Water Use Data for Crow Creek Irrigation Unit. (Source: Unpublished U.S. Bureau of Reclamation data, furnished by Arnold Kohlberg, Toston, Montana)

Year	1968	1969	1970	1971	1972	1973	1974	1975	1976	1977	1978	1979
Total water diverted from Missouri River ($x10^8$ ft^3)	7.26	7.58	8.17	8.45	9.25	10.10	8.13	6.30	7.58	8.85	10.86	7.96
Main canal waste ($x10^7$ ft^3)	3.41	2.51	3.27	1.81	3.26	4.05						
Main canal losses ($x10^7$ ft^3)	6.73	7.75	8.16	12.68	13.00	17.17						
Delivered to laterals ($x10^8$ ft^3)	6.25	5.12	5.49	5.73	5.97	6.01						
Lateral waste ($x10^7$ ft^3)	7.07	4.41	7.19	8.82	12.65	11.46						
Lateral losses ($x10^8$ ft^3)	1.15	1.30	1.28	1.26	1.29	1.32						
Delivered to farms ($x10^8$ ft^3)	4.52	4.81	5.03	4.86	5.07	5.51	5.93	3.87	3.89	4.39	4.55	4.17
Acres irrigated	4644	4630	4603	4692	4712	5249	5249	5338	5498	5585	5842	5692

Irrigation wells have reduced the amount of land irrigated by water diverted from ditches, streams, and springs. Well water has replaced ditch water on about seven square miles of land in the area between Kimpton Spring and Plunket Lake. Well water has made an additional three square miles of land arable in this area.

Stratigraphy

A generalized stratigraphic column for the drainage basin is presented in Table 5. Ages, names, lithologies, and thicknesses were compiled from Davis and others (1980), Freeman and others (1958), Klepper and others, (1957, 1971), Kuenzi and Fields (1971), Lorenz and McMurtrey (1956), Robinson (1963, 1967), and Smedes (1966). Lithologies for each stratigraphic unit are listed in order of

Table 5. Generalized Stratigraphic Section and Water-Bearing Properties of Rocks in the Radersburg Drainage Basin (Compiled From: Davis and others, 1980; Freeman and others, 1958; Klepper and others, 1957, 1971; Kuenzi and Fields, 1971; Lorenz and McMurtrey, 1956; Robinson, 1963, 1967; Smedes, 1966)

ERA	SYSTEM or SERIES	FORMATION	LITHOLOGY and THICKNESS	WATER-BEARING PROPERTIES
Cenozoic	Quaternary	undifferentiated	Eolian silt; terrace gravel; alluvial fan sand and gravel; glacial deposits. 60' maximum.	Yields adequate supplies for domestic and stock purposes. In some places, yields may be sufficient for garden irrigation.
	Quaternary-Tertiary	undifferentiated	Pediment and fan gravels with clasts of limestone, quartzite, and volcanic rocks. 5-100'.	Extent too small to be considered as a source of water.
			UNCONFORMITY	
	Tertiary	Bozeman Group: Renova Formation	Rhyolitic sedimentary tuff with a thick lens of bentonitic clay. Intraformational gravel. 0-1900'.	Yields adequate supplies for domestic and stock purposes.
			NONCONFORMITY	
Mesozoic	Lower Tertiary to Upper Cretaceous	Intrusive rocks	Granitic to gabbroic rocks of the Boulder Batholith. Dioritic rocks related to the Elkhorn Mountains Volcanics. Thickness unknown.	Yields water to springs through fractures. Not an important source of water.
			ANGULAR UNCONFORMITY	
	Upper Cretaceous	Elkhorn Mountains Volcanics	Andestitc lava and volcanic breccia; some rhyolitic welded tuffs. 5000+'.	Yields water to springs from fractures and breccias. Not an important source.
		CONFORMITY, DISCONFORMITY, AND ANGULAR UNCONFORMITY AT DIFFERENT PLACES		
		Slim Sam Formation	Crystal lithic tuff, andesitic sandstone, and black shale. 0-1200'.	Not water bearing.
	Upper & Lower Cretaceous	Colorado Formation	Black shale, sandstone, and siltstone. 0-1500'	Not water bearing
			DISCONFORMITY	
	Lower Cretaceous	Kootenai Formation	Sandstone, conglomerate, shale and fossiliferous limestone. 0-650'.	Locally contains a little water.
			DISCONFORMITY (?)	
	Upper Jurassic	Morrison Formation	Shale, mudstone, siltstone, limestone, and sandstone. 400-540'.	Not an important source of water in this area.
		Swift Formation	Sandstone and conglomerate. 20-35'.	Not an important source of water in this area.
			DISCONFORMITY	
Paleozoic	Permian	Phosphoria Formation	Chert and quartzite, locally phosphatic. 40-120'.	Not water bearing.
	Pennsylvanian	Quadrant Formation	Quartzitic sandstone and dolomite. 225-325'.	Yields small supplies, but not imprtant as source of water.

Table 5. (continued)

ERA	SYSTEM or SERIES	FORMATION		LITHOLOGY AND THICKNESS	WATER-BEARING PROPERTIES
Paleozoic	Pennsylvanian & Mississippian		Amsden Fm. & Big Snowy Gp.	Red calcareous siltstone, shale limestone, and dolomite. 200-300'.	Not an important source of water.
				DISCONFORMITY	
	Mississippian	Madison Group	Mission Canyon Limestone	Fine- to coarse-grained, massive limestone. Local collapse breccia in upper part. 800-1500'.	Yields large supplies of water to many springs.
		Madison Group	Lodgepole Limestone	Thin- to medium-bedded fossiliferous limestone; mudstone interbeds at base. 300-700'.	
	Upper Devonian		Three Forks Shale	Shale, siltstone, and limestone. 100-400'.	Generally not water bearing or produces only meager supplies.
			Jefferson Dolomite	Dolomite and limestone. 500-735'.	Source of small springs.
			Maywood Formation	Silty dolomite and calcareous siltstone and shale. 40'.	Yields very little water.
				DISCONFORMITY	
	Upper Cambrian		Red Lion Formation	Calcareous siltstone, shale, and sandstone. 0-90'.	Yields very little water.
			Pilgrim Dolomite	Mottled limestone and dolomite. 350-500'.	Yields water to small springs from joints and solution passages.
	Middle Cambrian		Park Shale	Fissile shale with a few limestone, siltstone, and sandstone beds. 200-300'.	Not water bearing.
			Meagher Limestone	Mottled and banded limestone and dolomite. 360-500'.	Yields water to small springs from joints, bedding planes, and solution channels.
			Wolsey Shale	Shale and mudstone, commonly silty. Limestone near top, quartzite at base. 300-450'.	Not water bearing.
			Flathead Quartzite	Quartzite with a few thin units of shale or siltstone. Finely conglomeratic at base. 25-120'.	Yields meager supply of water to small springs from many cracks and fractures.
				ANGULAR UNCONFORMITY	
Precambrian	Precambrian	Belt Supergroup	Empire Shale	Drab siliceous mudstone. 0-800'.	Not water bearing.
		Belt Supergroup	Spokane Shale	Red mudstone, shale, and sandstone. 0-5000'.	Yields small supplies of water to many springs.
		Belt Supergroup	Greyson Shale	Gray and brown mudstone, shale, sandstone. 0-1500+'.	Yields small supplies of water to many springs.

Note: A thickness of "0" indicates that the unit is locally not present decreasing occurrence within a unit. Water-bearing properties of the units shown are discussed in greater detail in the "Hydrogeology" section of this paper.

Where necessary, formational names were modified or replaced to conform with present usage. For example, Lorenz and McMurtrey (1956) considered the Cambrian "Dry Creek Shale" as the formation between the Cambrian Pilgrim Dolomite and Devonian Jefferson Dolomite. Klepper and others (1957) determined that these argillaceous rocks were equivalent to the Red Lion and Maywood formations of Cambrian and Devonian systems. The unit between the Mission Canyon Limestone and Quadrant Formation was called "Amsden Formation" by Klepper and others (1957) and Freeman and others (1958). This unit was split into the Amsden Formation and Big Snowy Group because of work done in the Limestone Hills by Blake (1959) and Dutro and Sando (1963). The lower Tertiary (Oligocene) Renova Formation is equivalent to all of the following: Freeman and others' (1958) Tertiary units 1, 2, and 3 (To, Tu_2, Tu_3); Klepper and others' (1957) rhyolitic tuff and gravel (Ttg); Klepper and others' (1971) conglomeratic or bentonitic rhyolitic sedimentary tuff (To, Tu); and Robinson's (1963, 1967) Dunbar Creek and Climbing Arrow formations of the Bozeman Group (Hughes, 1980; Kuenzi and Fields, 1971).

Structural Geology

The geologic structure in the vicinity of the Radersburg drainage basin has been described in detail by the authors listed on Figure 1.

Their descriptions were field-checked by Straw (1979) during this investigation. The following description of the structural geology of the area shown on Plate 1 was compiled from the available literature.

Pre-Tertiary layered rocks of the drainage basin have been compressed into a series of north-trending folds which have been cut by faults of differing orientations. From the northwest part of the basin to the southeast part of the basin are the Elkhorn Mountains syncline, the Devils Fence dome, the Radersburg syncline, and the Limestone Hills and Hossfeldt Hills anticlines. The axes of these folds generally trend from north to N 20° E (Plate 1).

The Elkhorn Mountains syncline is a very broad, open fold. In the area shown on Plate 1, the fold is about six to ten miles wide. The axial trace of the fold may only be approximately located because of the intrusive bodies and block faults which are present in the area (Smedes, 1966). South of Radersburg Pass, the fold splits into a series of synclines and anticlines.

The north-south elongated Devils Fence dome is made up of nearly concentric belts of post-Precambrian rocks around a core of Precambrian meta-sedimentary rocks. In the north, it is roughly symmetrical and has minor folds in the crest and on its flanks. The southern part is very broad and plunges more steeply than the northern part (Klepper and others, 1957).

The northward-plunging Radersburg syncline is best exposed in the Cold Springs Gulch area just north of Crow Creek, where the fold is ten to twelve miles wide. Here the west limb dips more shallowly than the east limb. The trough of this fold has been partly invaded by intrusive and extrusive rocks so that the axial trace is nearly impossible to locate (Klepper and others, 1971). The location of the axis of the fold is unknown between Crow Creek and the vicinity of the Wheatland School because of intrusive and extrusive rocks and sedimentary cover. The axis of the fold extends to the south from the drainage basin to the area east of Milligan Creek (Robinson, 1963).

East of the Radersburg syncline are the Limestone Hills and Hossfeldt Hills anticlines. The Limestone Hills anticline has a broad crest, steeply dipping west limb, and gently dipping east limb. Minor folds occur on both limbs. The Hossfeldt Hills anticline has been slightly overturned to the east and cut by the Lombard thrust fault (Freeman and others, 1958).

Faulting has offset the axes of the major folds in the drainage basin. Many of the thrust and strike-slip faults which have been mapped in the area were formed at the culmination of folding. Freeman and others (1958, p. 525) suggested that

> stresses in the overriding block near the thrust surface were relieved mainly by folding and minor thrusting, while at a greater distance from the thrust the relief was by strike-slip movement along faults that may have formed earlier in the deformation but that acted as tears during the latest stage of deformation.

Thrust faults occur in the eastern part of the basin. The Hossfeldt Hills anticline is cut by the west-dipping Lombard thrust fault. Minor thrusts are also present between the Hossfeldt Hills and Plunket Lake. These thrust faults in the southern part of the drainage basin may extend to the north, where they have been concealed by basin-fill sediments.

Steep, north-trending faults parallel the axial planes of folds and may have formed during the period of maximum compression as fold limbs slipped over each other (Klepper and others, 1957). Stratigraphic displacement on these faults is 200 to 3,000 feet. Conjugate northwest and northeast-trending faults and lineaments probably are younger than north-trending faults (Klepper and others, 1957). Displacement on the northwest-trending faults is mostly left-lateral and from 1,000 to 3,000 feet. The northeast-trending faults show mostly vertical offset.

Northwest-trending left-lateral faults have been mapped in the Limestone Hills. The valleys of Crow Creek and Indian Creek are partly controlled by northwest-trending faults. A concealed extension of the Crow Creek fault may have caused the southern termination of the Limestone Hills (Straw, 1979). The aeromagnetic map of the Devils Fence, Radersburg, and Toston 15-minute quadrangles (Kinoshita and others, 1965) shows left-lateral offset of the buried continuation of a Precambrian (?) diabase sill which is exposed in the Limestone Hills. Straw (1979) contends that the extension of the left-lateral Crow Creek fault into the Radersburg valley has offset the axial trace of the Radersburg syncline.

Other faults of varying orientations and offset of a few feet to hundreds of feet occur throughout the basin. Some may have been formed during emplacement of intrusives, collapse of magma chambers and uplift of the Elkhorn Mountains. Pre-Tertiary bedding-plane faults and faults in the Tertiary and younger sedimentary rocks and sediments were mentioned by previous workers (Freeman and others, 1958; Klepper and others, 1957), but they did not map these faults.

Geologic History

The geologic history of the area has been described by the authors credited in Figure 1 and is summarized below. The history of the Elkhorn Mountains-Radersburg valley area may be divided into three stages. The first stage was a time when shallow-water sediments were deposited on a relatively stable surface. Crustal instability and igneous activity marked the second stage. The final, most recent stage has been one of uplift and erosion (Klepper and others, 1957).

The earliest stage lasted from Late Precambrian to Cretaceous. Belt Supergroup rocks were deposited in a shallow sea, then arched and eroded prior to deposition of the marine Middle Cambrian Flathead Quartzite. A conformable sequence of marine carbonates, shales, and sandstones was then deposited (Table 5). The sea regressed at the end of the Cambrian and erosion resulted. Transgression occurred in Late Devonian, when carbonates, sandstones, and shales were deposited without angular unconformity on a surface of negligible relief. Transgression continued until Permian time, with one minor withdrawal at the end of the Mississippian. After deposition of the Phosphoria Formation, the sea withdrew and there was a hiatus until brief transgression when the marine Swift Formation was laid down. Continental deposits of the Morrison and Kootenai formations conformably overlie the Swift. The period of relative stability ended in Colorado time when the area was alternately above and below sea level (Freeman and others, 1958; Klepper and others, 1957).

The second stage began with volcanism and folding, recorded in the locally folded volcaniclastic rocks of the Slim Sam formation (Klepper and others, 1971). Andesitic breccias, tuffs, and flows of the Elkhorn Mountains volcanics were erupted. Some folding and faulting took place during this period, and concordant or partly concordant dioritic intrusives, compositionally similar to the volcanic rocks, were squeezed into weak zones (Klepper and others, 1971). The main episode of Laramide deformation began after volcanism. The volcanic rocks dip less steeply than the underlying units and the dioritic intrusives have been folded (Klepper and others, 1971). Eastward thrusting of the Elkhorn Mountains block occurred near the end of the period. Late Cretaceous to Paleocene (?) rocks related to the Boulder batholith were intruded after this faulting. Many of these rocks occupy fractured zones in synclinal troughs (Klepper and others, 1957).

Uplift and erosion took place during the third stage. This stage began after the intrusion of the Boulder batholith and has continued to the present. Prior to the now Eocene Oligocene, the Elkhorn Mountains were uplifted relative to the valley floor, and erosion of the mountains produced a mature topography. None of these pre-Oligocene sediments were preserved in the valley. During the Oligocene, sedimentary tuff and interbedded gravel were deposited in fluvial, deltaic, and lacustrine environments in the valley (Freeman and others, 1958).

From late Oligocene to Miocene time, the Limestone Hills and Elkhorn Mountains continued to be uplifted and to shed sediments to the ancestral Townsend valley, north of the Radersburg valley. The Radersburg valley was relatively stable, as suggested by thinning of Oligocene sediments in the study area (Freeman and others, 1958). During Pliocene time, there was a halt in uplift, the Missouri River maintained a nearly constant base level, and an eastern-sloping pediment formed on the west side of the valley (Pardee, 1925). Uplift then began again, which deformed the Tertiary rocks and rejuvenated the Missouri River. With this rejuvenation, tributaries of the Missouri River incised their drainages, cut into the pedimented surface, and formed alluvial fans. Quaternary time has been one of continued erosion, except during times when local glacial deposits have accumulated (Freeman and others, 1958). Minor uplift of the mountains has occurred to the west of the Devils Fence and along range front faults north of the Limestone Hills (Klepper and others, 1957).

HYDROGEOLOGY

Introduction

Geologic formations which are identified by distinctive lithologic characteristics may be grouped on the basis of water transmitting properties into aquifers and aquitards (Maxey, 1964). Aquifers are rock units which will transmit significant (or economically important) amounts of water to wells or springs under normal hydraulic gradients. Aquitards are less permeable rock units which will not easily transmit water to a well or spring, but may be highly porous and store great amounts of water and allow water to be transmitted to or from adjacent aquifers (Miller, 1980, p. 496).

There are two types of aquifers (Figure 3). Unconfined or water-table aquifers are bounded by an upper surface known as the water table, where water pressure is equal to atmospheric pressure. Confined or artesian aquifers contain water under pressure and are bounded above and below by aquitards. Water levels in wells emplaced in confined aquifers will usually rise above the top surface of the aquifer. Water level data for confined aquifers, which relate to hydraulic head or potential in the aquifer, may be plotted on a map and contoured to define the potentiometric surface of the aquifer.

The hydraulic conductivity, transmissivity, storage coefficient, and hydraulic gradient are used to characterize the amount of water stored in an aquifer and the rate of movement of water in an aquifer.

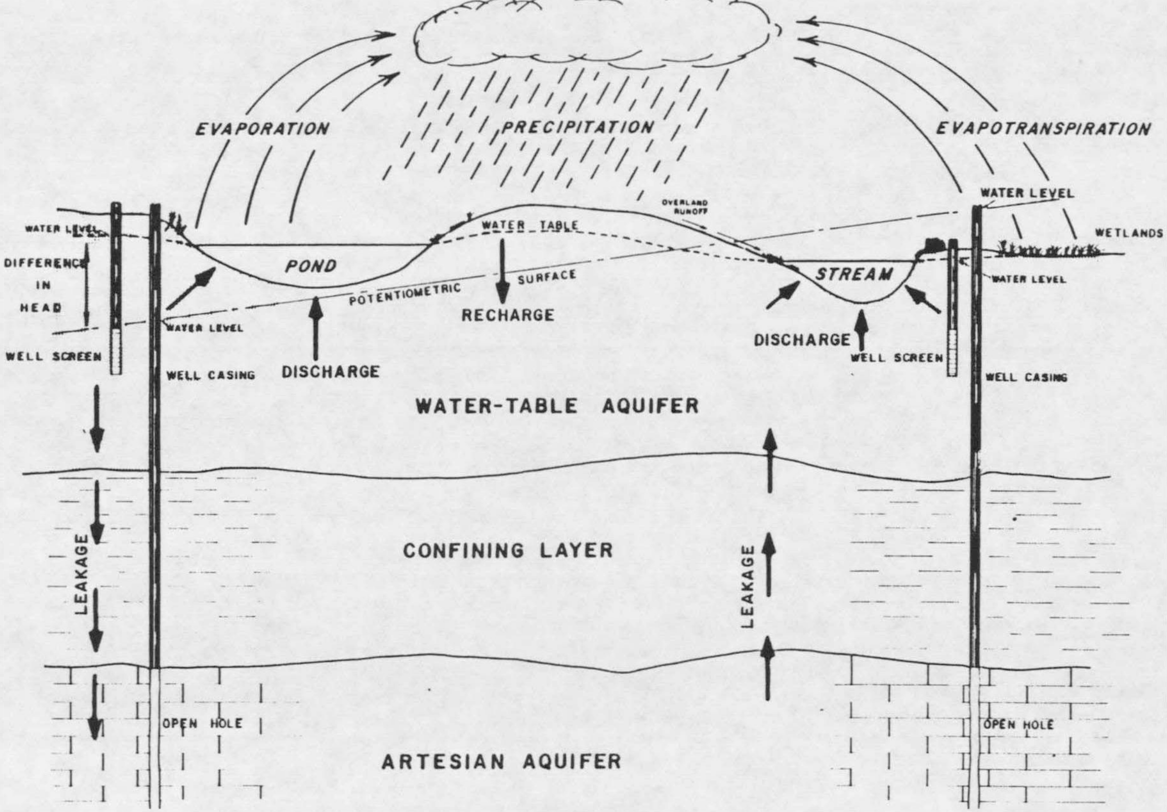


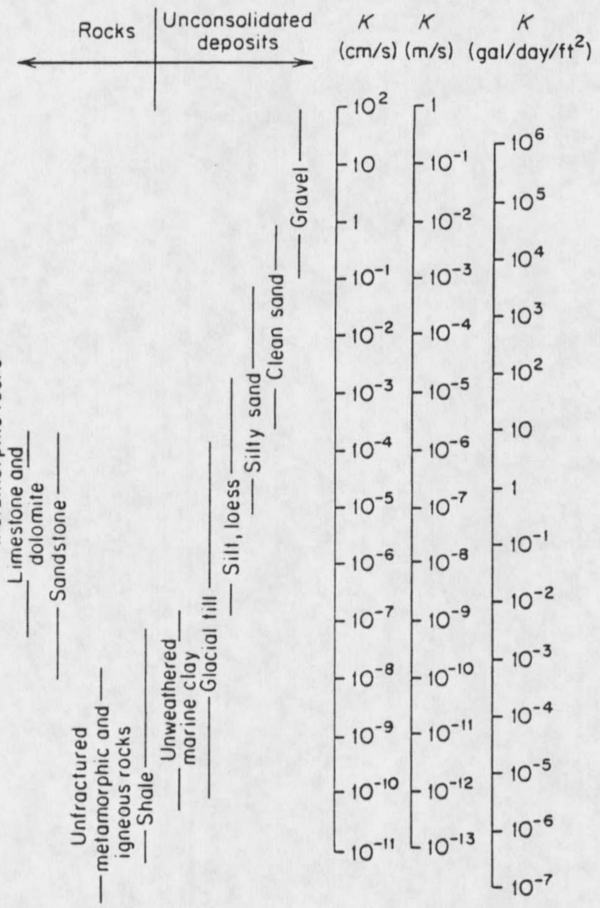
Figure 3. Aquifer Types (Source: Miller, 1980)

Hydraulic conductivity is a measure of permeability and describes the rate at which a unit volume of water will flow through a unit cross-sectional area of material under a unit hydraulic gradient. Transmissivity is the rate at which water will move through a unit width of an aquifer at a unit hydraulic gradient and is the product of the hydraulic conductivity multiplied by the aquifer thickness. The storage coefficient describes the volume of water an aquifer will lose or gain to storage per unit surface area per unit change in head (Lohman, 1972). The hydraulic gradient is the change in static head or potential per distance along a path of flow (Miller, 1980).

The formations listed in Table 5 have been grouped and classified as aquitards or aquifers based on descriptions given in Lorenz and McMurtrey (1956), Taylor (1978), and Bergeron (1979); typical values for hydraulic conductivity of unconsolidated sediments and rocks (Table 6); specific capacity tests (Table 13 in Appendix A); and hydraulic properties determined by my field work. Aquifers in the area are the Mississippian Madison Group and the Tertiary and Quaternary basin-fill and alluvial sediments. These aquifers yield great amounts of water to wells and springs in the basin and have hydraulic conductivities from about one hundred to over one thousand gallons per day per square foot (gpd/ft^2). These aquifers are discussed in greater detail below.

Other formations in the basin have hydraulic conductivities much lower than the aquifers (Miller, 1976) and yield little or no water to wells or springs or are not water bearing (Lorenz and McMurtrey, 1956).

Table 6. Hydraulic Conductivities of Rocks and Unconsolidated Materials (Source: Freeze and Cherry, 1979)



These units (Table 5) are very poor aquifers or aquitards and include Precambrian through Upper Devonian formations underlying the Madison Group, Mississippian-Pennsylvanian through Upper Cretaceous sedimentary rocks overlying the Madison Group, and Upper Cretaceous to lower Tertiary extrusive and intrusive igneous rocks. The estimated hydraulic conductivities of these rocks are less than one hundred gpd/ft^2 (Table 13) and may be as low as 10^{-6} gpd/ft^2 (Table 6). These units have low hydraulic conductivities because the rocks are generally fine-grained, are cemented with silica or calcite, or have intergranular spaces which are filled with micaceous or argillaceous material. Small fractures, joints, bedding planes, and solution channels in these units probably act as zones through which water is transmitted. Evidence for the transmission of water through joints was seen at Hunsaker Spring, where water discharges from joints in Precambrian bedrock.

Mississippian Madison Group Aquifer

The most productive pre-Tertiary aquifer in the region is the Mississippian Madison Group. The aquifer has been extensively studied in Montana, Wyoming, and the Dakotas in connection with oil, coal, uranium, and geothermal resource development (Balster, 1974; Blankennagel and others, 1977; Brown and others, 1977; Head and Merkel, 1977; Hopkins, 1976; Miller, 1976; Swenson and others, 1976; Taylor, 1978). Well yields range from 20 to 9,000 gpm and water temperature varies from about 10°C near land surface to about 120°C where the unit is very deep (Taylor, 1978).

The Madison Group aquifer is composed of two formations in southwest Montana: the Mission Canyon Limestone and the Lodgepole Limestone (Table 5). The Mission Canyon Limestone in the Radersburg drainage basin is a medium-gray to light-gray, finely to coarsely crystalline, thick- to massively bedded limestone. Nodules and lentils of chert and locally fossiliferous beds occur in the upper part of the unit south of Plunket Lake. Solution breccias and karst features are present in the upper part of the Mission Canyon in the drainage basin (Klepper and others, 1957, p. 19-20; 1971, p. 5; Robinson, 1963, p. 43-44) as well as in other parts of the Northern Rockies (Aram, 1979; Aram and others, 1978; Campbell, 1977; Gries and Crooks, 1968; Roberts, 1966; Sando, 1974). The Mission Canyon Limestone grades downward into the thinly and well-bedded, medium-gray, fossiliferous, fine- to coarse-grained Lodgepole Limestone, in which mudstone interbeds are present.

Water is transmitted through the Madison Group aquifer by primary and/or secondary permeability. Primary permeability is a function of grain size, sorting, shape, and organic or inorganic origin of the sediment at the time of deposition (Moore, 1979). Primary permeability will be high in a carbonate unit made up of sand-sized particles or skeletal remains of reef-building organisms with minor amounts of intragranular carbonate mud. Primary permeability appears to be very low in the Madison Group because much of the unit is microcrystalline with only minor amounts of sand-sized or larger particles. Hydraulic conductivities of 83 Madison Group core samples from a test well 300 miles east of the study area, in Custer County, Montana, ranged from less than 1.8×10^{-4} gpd/ft^2 to 0.33 gpd/ft^2 with a mean of 1.6×10^{-2} gpd/ft^2 (Brown and others, 1977). Twenty-nine of the samples were of unfractured, non-brecciated, microcrystalline to coarse-grained limestone. The samples yielded primary hydraulic conductivity values from 1.8×10^{-4} gpd/ft^2 to 5.8×10^{-3} gpd/ft^2 with a mean of 5.5×10^{-4} gpd/ft^2. Porosities of these samples ranged from 0.3 to 9.4 percent and averaged 3.2 percent (Brown and others, 1977).

Secondary permeability plays a more important role in water transmission in the Madison Group. Possible sources of secondary (post-depositional) permeability include dolomitization, formation of karst and solution breccias, and fracturing. Dolomitization is conversion of calcitic limestone to dolomitic limestone or dolostone. The process results in increased porosity and permeability (Blatt and others, 1972) and may be a possible source of secondary permeability in the Madison Group. Dolostone or dolomitic limestone has been observed in the Mission Canyon Formation near Livingston, Montana (Roberts, 1966); however, dolomitization may be neglected as a source of highly permeable zones in the Radersburg drainage basin. Thin sections of Mission Canyon Limestone samples collected during this study from south of Plunket Lake showed no dolomite, and dolomite is not mentioned in any of the Mission Canyon stratigraphic sections measured in the area (Klepper and others, 1957, 1971; Ruppel, 1950).

Karst features, such as caves, sinkholes, and enlarged joints, cut across strata in the Madison Group. Fossil karst topography in the Toston area and Limestone Hills has been described as an irregular surface with up to 100 feet of local relief which marks the contact between the base of the Big Snowy Group and the upper surface of the Mission Canyon Limestone (Klepper and others, 1971; Robinson, 1963). Breccias within the Mission Canyon Limestone which have red and yellow limy-mud cement between angular limestone blocks occur in the Elkhorn Mountains and are remnants of collapsed caves and sinkholes (Klepper and others, 1971). These breccias, which are discontinuous, fill upward-widening cavities or joints formed between Mission Canyon Limestone and Amsden Formation deposition (Roberts, 1966).

In contrast with the karst features are stratigraphically controlled evaporite-solution breccia deposits. Solution breccia beds have a poorly defined upper surface and well defined lower boundary, are from 10 to 75 feet thick, and occur about 200 feet below the top of the Madison Group in the southern Elkhorn Mountains. Klepper and others (1957, p. 19) reported that the solution breccias consist of

> a chaotic mass of angular blocks, slabs and finer textured rubble, in part cemented by laminated yellowish or reddish calcareous mudstone similar to beds in the lower part of the Amsden Formation and in part by finely crystalline calcium carbonate.

Solution breccia beds are very common in the Mission Canyon Limestone throughout the Northern Rockies. Solution breccias were formed by the leaching of evaporite beds after the Madison Group was uplifted during the Late Cretaceous to early Tertiary Laramide Orogeny (Roberts, 1966). In other places where there was no uplift, the evaporite beds are still present in the subsurface. Anhydrite beds were penetrated in a well near Ringling, Montana, 40 miles east of the study area. The anhydrite beds correlate with exposed solution breccia intervals in the Mission Canyon Limestone near the study area (Roberts, 1966).

Laboratory and field evidence supports the contention that secondary permeability is greater than primary permeability in the Madison Group. Fourty-six of the 83 Madison Group core samples from the Custer County well were of dolomitic limestone, dolostone, and brecciated or fractured limestone. Secondary hydraulic conductivities in this group of samples ranged from 1.8×10^{-4} gpd/ft^2 to 0.33 gpd/ft^2 and averaged 3.6×10^{-2} gpd/ft^2. Porosities ranged from 0.3 to 24 percent and averaged 7.0 percent. In addition to the rock-core permeability tests which indicate high secondary permeability, well drillers have reported lost-circulation zones in the Mission Canyon Limestone in southeastern Montana (Miller, 1976). Loss of drilling fluid circulation between the drill bit and land surface occurs because of voids in a formation. This may be an indication of karst or solution brecciated areas at depth.

Stress-caused fractures, such as faults and joints, also contribute to the secondary permeability of the Madison Group. Plunket Spring is formed by a series of north-south trending springs in the southern end of Plunket Lake. The springs are colinear with the north-south trend of joint sets in Mission Canyon Limestone outcrops just south of the lake. The springs issue over a distance of about 50 feet from openings in very close- to closely spaced joints (1 to 4 inches apart) in the Mission Canyon Limestone.

Plunket Spring and Big Spring show that the Madison Group is a significant source of water in and adjacent to the Radersburg basin and that water is confined or under pressure in the aquifer. Water flowing from Plunket Spring causes churning of the surface of Plunket Lake when about 1.5 feet of water covers the openings. The churning indicates a pressure head above land surface. Discharge from Plunket Spring is impounded to fill Plunket Lake. The lake level varies as water is stored or discharged into Warm Springs Creek for irrigation or stock watering. The amount of water discharged into Warm Springs Creek is controlled by a gate valve and varied from 0.2 to 11.5 cfs during 1979 (Table 1). Previous workers reported the spring flow to be constant throughout the year and gaged the discharge from the spring at 9.7 and 8.7 cfs (Pardee, 1925; Lorenz and McMurtrey, 1956). Using the latter figure, the discharge to the lake is about 3×10^8 cubic feet per year (ft^3/yr) or 2×10^9 gallons per year.

approx 3858 gpm

Big Spring, on the east bank of the Missouri River, east of the drainage basin, flows from the basal part of the Amsden Formation, but is probably connected by a fault zone to a "cavernous reservoir" in the Madison Group (Lorenz and McMurtrey, 1956, p. 216). The flow rate of Big Spring in May of 1922 and 1951 was 64.4 and 56.7 cfs (Pardee, 1925, p. 46; Lorenz and McMurtrey, 1956, p. 216) or about 2×10^9 ft^3/yr. Water from this spring is diverted to the area north of Toston for irrigation.

under pressure

Basin-Fill Aquifer

Geology

Thick Tertiary-age basin-fill deposits and thin Quaternary alluvium along stream channels occur in intermountain basins in southwestern Montana. These poorly consolidated to unconsolidated sediments of clay, silt, sand, and gravel form the principal aquifers in the area (Taylor, 1978).

Poorly consolidated Tertiary-age sediments in the Radersburg valley consist of rhyolitic or bentonitic sedimentary tuff. The rhyolitic part of this unit is made up of beds of clay- to fine sand-sized ash fragments and coarser sand grains and granules interbedded with beds of very-fine to coarse-grained conglomeratic sandstone with pre-Tertiary clasts up to one foot in diameter. The rhyolitic tuff beds are thin and distinct to thick and indistinct or lenticular, horizontal to gently dipping, and typically contain over fifty percent tuff with interbeds of fine- to coarse-grained sand and gravel. This lens is about 2.5 miles long and hundreds of feet thick, and crops out about 4 miles south of Lone Mountain. Thickness of the Tertiary unit ranges from a featheredge overlapping older rocks to about 1,000 feet in the southernmost part of the drainage basin (Klepper and others, 1957, p. 42; Klepper and others, 1971, p. 12-13; Freeman and others, 1958, p. 508-510; Robinson, 1967).

Unconsolidated Quaternary deposits overlie the Tertiary material. A ten-foot thick veneer of Late Tertiary to Quaternary pediment gravel, consisting of locally derived gravel with a sand and clay matrix, is present near the Elkhorn Mountains. The pediment deposits thin to the east and are less than 5-feet thick near Plunket Lake (Freeman and others, 1958). Quaternary sediments in the valley are also locally derived and include glacial outwash near the Elkhorn Mountains, alluvial fan sand and gravel, alluvium of the present drainage systems, and eolian silt. The maximum thickness of these sediments is probably less than 60 feet (Lorenz and McMurtrey, 1956). Some domestic and stock wells along Crow Creek and Warm Springs Creek withdraw water from this aquifer.

Although many wells have been drilled in the Radersburg valley, the depth to the subsurface boundary between Tertiary basin-fill deposits and Quaternary alluvium is unknown. Logs of water wells (Table 12 in Appendix A) are generally not descriptive enough to allow distinction between Quaternary and Tertiary beds of clay, sand, and/or gravel. However, the contact between basin-fill sediments and alluvium probably is less than one hundred feet below the land surface. At depths between 100 and 400 feet are Tertiary sand and/or gravel beds from two to twenty feet thick and as much as 60 feet thick. These beds are interbedded with clay beds usually one to four feet thick and as much as 40 feet thick. Overlying the Tertiary beds are thinner beds of Tertiary and/or Quaternary sand and gravel interbedded with clay. Because of the inability to distinguish between the Quaternary and Tertiary sediments, and the relative thinness of the Quaternary sediments, the aquifer formed by both Tertiary and Quaternary unconsolidated sediments is referred to as the basin-fill aquifer in this report.

Hydrology

Many wells have been drilled into the basin-fill aquifer in the Radersburg valley to supply water for domestic, stock, and irrigation uses. The hydrologic properties of the aquifer may be directly measured because of these wells.

Well Design. Water wells which have been drilled into the basin-fill aquifer were identified using existing drillers' well logs and ground water appropriation forms. Information from these documents is summarized in Appendix A. Water wells in the study area have been drilled for domestic, stock watering, and irrigation use. Table 11 in Appendix A lists well construction information. Domestic and stock wells are usually six inches in diameter and are equipped with submersible pumps. Most of the domestic and stock wells in the valley have been drilled to a maximum depth of less than two hundred feet using cable-tool drilling methods. Well construction methods vary; some wells draw water from torch-cut slots, some use commercially avaliable slotted well screens, and others intake water from open-ended casing. The wells have been completed by natural development. This process involves over-pumping or rapid bailing after installation which creates a natural gravel pack around the intake zone. Over-pumping or rapid bailing removes fine-grained materials from water-bearing sand gravel around the well and increases the permeability of the aquifer around the intake zone (Johnson Division, 1975). Domestic and stock wells generally yield less than 50 gpm (Table 13). Although Lorenz and McMurtrey (1956, p. 189) found that "in some places yields [from the near-surface sediments] may be sufficient for irrigation," they were discussing garden irrigation using yields of 8 to 10 gpm (p. 218). Garden irrigation requires much less water than does the irrigation of large fields, which uses 1,000 to 3,000 gpm of water.

The large volumes of water are produced by high-capacity irrigation wells drilled into the basin-fill aquifer. High-capacity irrigation wells have greater maximum depths, are larger in diameter, and have been designed to draw water from greater thicknessess of sands and gravels than domestic or stock wells. Line-shaft turbine pumps supply from 1,000 to 3,000 gpm of water to side-roll sprinkler, center-pivot sprinkler, and flood irrigation systems. Cable-tool drilled wells are generally equipped with 16-inch diameter casing and have been naturally developed. Wells drilled by reverse rotary techniques are constructed of 16-inch or larger casing, centered in artificially gravel-packed boreholes up to 32 inches in diameter. Torch-cut slots, slotted well screens, and louvered screens have been used in the high-capacity irrigation wells (Table 11). Irrigation wells tested by water well drillers have yielded 600 to 3400 gpm in the basin-fill sediments (Table 13).

<u>Transmissivity and Hydraulic Conductivity</u>. The transmissivity and hydraulic conductivity of the basin-fill aquifer were estimated

throughout the study area using specific capacity data and were measured in the vicinity of two wells using aquifer test methods. The specific capacity of a well is a measure of well yield divided by the difference between static and pumping water levels and is expressed in gallons per minute per foot of drawdown (gpm/ft). Specific capacity data are affected by factors such as variations in well design, well efficiency, and test duration. Additionally, many of the specific capacity values reported in Table 13 were calculated from discharge and drawdown measurements reported on drillers' well logs. The accuracy of these measurements is unknown. Meyer's (1963) method was used to estimate transmissivity from specific capacity data. Estimated hydraulic conductivities of the aquifer were determined by dividing the estimated transmissivity by the total length of perforated or screened interval of each well. The test procedures, methods, and results are described in Appendix A.

Specific capacity tests of domestic, stock, and observation wells (Table 13) usually yielded estimated transmissivity values of less than 5000 gpd/ft but ranged up to 100,000 gpd/ft. Estimated hydraulic conductivity averaged 540 gpd/ft^2. Tests of larger diameter irrigation wells yielded basin-fill sediment transmissivity and hydraulic conductivity estimates averaging 50,000 gpd/ft and 430 gpd/ft^2.

Aquifer tests provide more accurate assessment of hydraulic conductivity and transmissivity than specific capacity tests. This is because aquifer tests are generally conducted under more controlled conditions than specific capacity tests and because data from aquifer tests are analyzed by methods that best fit the aquifer geometry and test conditions (Stallman, 1971).

Aquifer tests consisted of a pumping test and a recovery test. Data from the tests and analyses of the tests are described in detail in Appendix B. The pumping test was performed by pumping irrigation well 4-1-10bcb at 1600 gpm for 44 hours. Wells 4-1-9aac and 4-1-10bbc were used as observation wells. Drawdown and recovery measurements were made during the pumping period and for nine hours after pumping ceased. The methods of Chow (1952), Cooper and Jacob (1946), Hantush (1956), and Theis (1935) were used to analyze the data. Because of equipment failures, variation in pumping rate during the test, and the relatively low efficiency of the pumping well, observation well data were used to determine aquifer properties. The four analytical methods used with recovery data consistently indicated that the transmissivity of the aquifer in the test area is 300,000 gpd/ft. The hydraulic conductivity of the aquifer is about 1800 gpd/ft^2.

The transmissivity of the aquifer in the vicinity of well 4-1-9ca was determined by a recovery test. The well was pumped for 52 minutes at 12 gpm. Insufficient drawdown measurements were made during the pumping period to allow analysis of these data. After the pumping period ceased, recovery measurements were made for 49 minutes. The recovery data were analyzed using the method of Cooper and Jacob (1946) and Theis (1935). The transmissivity determined was approximately 10,000 gpd/ft.

A comparison of the estimated and measured transmissivities shows that the estimated transmissivities derived from specific capacity tests may underestimate the aquifer's actual transmissivity. Estimated transmissivities of the aquifer in the areas of wells 4-1-9ca and 4-1-10bcb are 5000 gpd/ft and 80,000 gpd/ft, which are about 50% and 73% lower than the measured transmissivities of 10,000 gpd/ft and 300,000 gpd/ft, respectively. Order of magnitude approximations of the transmissivity of the aquifer near small-diameter and irrigation wells are 10,000 and 100,000 gpd/ft, respectively.

The estimated transmissivity values from tests of small and large diameter wells indicate that the aquifer's transmissivity is variable throughout the study area. This variability appears to be caused by the total thickness of the water-bearing beds of sand and gravel which yield water to each well and not because of variations in the hydraulic conductivity of the aquifer materials.

Because transmissivity is affected by aquifer thickness, the hydraulic conductivity appears to better describe the rate of water movement through the aquifer. There was no significant difference in estimated hydraulic conductivities derived from irrigation and small-diameter wells. Estimated hydraulic conductivities averaged 540 gpd/ft^2 from tests of small diameter wells and 430 gpd/ft^2 from tests of irrigation wells. There was no statistical difference between the means of the two types of wells when compared with the two-tailed t-test at a 10 percent confidence level using methods given in Davis (1973). The estimated hydraulic conductivity of the producing zones in the basin-fill aquifer is therefore on the order of 500 gpd/ft^2.

The estimated hydraulic conductivity is probably lower than the actual hydraulic conductivity of the aquifer. This may be because the same specific capacity data used to estimate transmissivities which are at least 60 percent lower than the measured transmissivities were used to compute estimated hydraulic conductivities. An order of magnitude estimate is that the hydraulic conductivity of the aquifer is 1000 gpd/ft^2.

Storage coefficient. Accurate determination of the storage coefficient of an aquifer requires the use of observation wells during a pumping test. The storage coefficient determined from the pumping test of well 4-1-10bcb indicates that ground water is confined in at least part of the basin. The value of 0.0016 obtained from the pumping test is within the 0.005 to 0.00005 storage coefficient range for confined aquifers given in Freeze and Cherry (1979). Other indicators of confined ground water are the presence of many clay layers in the valley sediments and the presence of at least one well (5-2-31dc) where the static water level is above the ground surface.

Although no pumping tests have been performed to confirm the presence of unconfined ground water, ground water is probably unconfined in some places. Lorenz and McMurtrey (1956, p. 208) state that "part of the ground water in Quaternary and Tertiary deposits is under water-table conditions." Drillers' logs listed by Lorenz and McMurtrey (1956) and Pardee (1925) indicate that unconfined ground water is present near Crow Creek and Warm Springs Creek. Near the creeks, the aquifer is composed of near-surface stream channel alluvial deposits of sand and gravel with minor clay interbeds. The lack of low permeability confining beds and the shallow water levels near the creeks suggest water in the sand and gravel is not under pressure and is therefore unconfined. The storage coefficient has not been measured in unconfined parts of the aquifer. Storage coefficient values where the aquifer is unconfined may range from 0.01 to 0.30 (Freeze and Cherry, 1979). In similar unconfined deposits of the Gallatin Valley to the southwest, the measured storage coefficient was at least 0.05 (Hackett and others, 1960).

Ground-water movement. Ground-water levels in existing wells were measured from May, 1979 to April, 1980. The methods used to measure water levels are given in Appendix A. Measurements of water levels are presented in Table 15 in that appendix. Water levels were measured in wells which may penetrate multiple water-producing zones in both confined and unconfined parts of the basin-fill aquifer. For this reason, the water levels do not necessarily show the absolute water table elevation or potentiometric surface elevation. However, the water level measurements allow the determination of the general direction of ground-water flow and the approximate hydraulic gradient in the aquifer.

Figure 4 is a water-level contour map drawn from static water level measurements made during June to August, 1979. Water level elevations in Figure 4 are to plus or minus five feet because ground surface elevations were estimated from topographic maps. The water level surface shown sloped from above 4080 feet in the west to less