Hydrogeology of the Armstrong and Nelson Springs, Park County, Montana

by William D Clarke III

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Earth Sciences
Montana State University

Abstract:
Armstrong and Nelson Springs, Park County, Montana, are nationally recognized trout fishing areas. The hydrogeology of the springs, however, has never been formally investigated. The stratigraphic and structural characteristics of the spring area suggest the springs may issue from any one of three aquifers present in the spring area and may be fault, stratigraphically, and/or topographically controlled. A hydrogeologic study was conducted to determine the source aquifer(s) of the springs, the controls on spring formation, and related groundwater flow systems.

Spring discharges and temperatures were periodically measured throughout the field study. The positions of individual springs were identified and mapped. Local domestic-water wells were inventoried, measured, and tested. Previously published geologic maps of the area were field checked and modified as necessary. Finally, groundwater samples were collected from springs and water wells for field and laboratory analysis.

Three aquifers are present in the study area: the Cenozoic Yellowstone and alluvial fan aquifers, and the Mississippian Madison Group aquifer. The three aquifers discharge chemically distinct water. Armstrong and Nelson Springs are made up of individual spring areas which peak in discharge at different times of the year and are positioned at different elevations with respect to the Yellowstone River. Some spring temperatures are seasonally dependent and some are seasonally independent. Each sampled Armstrong and Nelson Spring discharges chemically similar water which is equivalent to the chemical quality of the Yellowstone aquifer.

The Yellowstone aquifer is interpreted to be the source aquifer of Armstrong and Nelson Springs. The Yellowstone aquifer is a wedge-shaped alluvial body which thins from approximately 820 m in thickness south of the study area to 15.5 m in thickness immediately north of the study area. Chemical characteristics, groundwater flow patterns, and groundwater temperature patterns indicate the alluvial fan aquifer and the Madison Group aquifer are not source aquifers for the springs. Armstrong and Nelson Springs are topographically controlled and form in response to the thinning of the Yellowstone aquifer. Groundwater is forced to the surface to form the springs. Spring discharges, temperatures, and positions indicate both local and regional groundwater flow systems discharge in the area.

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MONTANA STATE UNIVERSITY
Bozeman, Montana

August 1991

APPROVAL

of a thesis submitted by

William D. Clarke III

This thesis has been read by each member of the thesis committee and has been found to be satisfactory regarding content, English usage, format, citations, bibliographic style, and consistency, and is ready for submission to the College of Graduate Studies.

15 August 1991
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Date August 16, 1991

ACKNOWLEDGMENTS

First and foremost, I would like to thank Dr. Stephan G. Custer for his guidance, patience and aid in obtaining funding for the project. Suggestions and constructive criticisms from Dr. William W. Locke and Dr. David R. Lageson were greatly appreciated and helped considerably in the preparation of this thesis. Special thanks are also extended to my colleagues. Your comments were greatly appreciated.

Funding for the chemical analyses was provided through the Donald L. Smith Memorial Scholarship and the Montana Bureau of Mines and Geology. Seismic data was provided by the Montana Power Company, and in particular Mr. Dave King. Special thanks is extended to the owners of the springs: the Nelson's, the O'Hair family and Mr. Dana. Finally, I would like to thank my wife, Lore Holt Clarke and my parents, Ginny Wilson and Bill Clarke. Without their support, patience, and understanding, completion of this project would not have been possible.

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ABSTRACT

Armstrong and Nelson Springs, Park County, Montana, are nationally recognized trout fishing areas. The hydrogeology of the springs, however, has never been formally investigated. The stratigraphic and structural characteristics of the spring area suggest the springs may issue from any one of three aquifers present in the spring area and may be fault, stratigraphically, and/or topographically controlled. A hydrogeologic study was conducted to determine the source aquifer(s) of the springs, the controls on spring formation, and related groundwater flow systems.

Spring discharges and temperatures were periodically measured throughout the field study. The positions of individual springs were identified and mapped. Local domestic-water wells were inventoried, measured, and tested. Previously published geologic maps of the area were field checked and modified as necessary. Finally, groundwater samples were collected from springs and water wells for field and laboratory analysis.

Three aquifers are present in the study area: the Cenozoic Yellowstone and alluvial fan aquifers, and the Mississippian Madison Group aquifer. The three aquifers discharge chemically distinct water. Armstrong and Nelson Springs are made up of individual spring areas which peak in discharge at different times of the year and are positioned at different elevations with respect to the Yellowstone River. Some spring temperatures are seasonally dependent and some are seasonally independent. Each sampled Armstrong and Nelson Spring discharges chemically similar water which is equivalent to the chemical quality of the Yellowstone aquifer.

The Yellowstone aquifer is interpreted to be the source aquifer of Armstrong and Nelson Springs. The Yellowstone aquifer is a wedge-shaped alluvial body which thins from approximately 820 m in thickness south of the study area to 15.5 m in thickness immediately north of the study area. Chemical characteristics, groundwater flow patterns, and groundwater temperature patterns indicate the alluvial fan aquifer and the Madison Group aquifer are not source aquifers for the springs. Armstrong and Nelson Springs are topographically controlled and form in response to the thinning of the Yellowstone aquifer. Groundwater is forced to the surface to form the springs. Spring discharges, temperatures, and positions indicate both local and regional groundwater flow systems discharge in the area.

INTRODUCTION

Armstrong and Nelson Springs in Park County, Montana (Figure 1), are nationally recognized for their prime trout habitats, spawning areas, and superb fishing challenges (Decker-Hess, 1986). The springs represent a natural resource which may become threatened by the increased residential expansion in the spring creek area. Paradise Valley is a richly scenic area of southwestern Montana and as such the valley is an attractive location for people to live. According to water-well-drilling records obtained from the Montana Department of Natural Resources and Conservation, Montana Bureau of Mines and Geology, and the Park County Courthouse, the number of domestic groundwater wells in the study area has increased from 22 in 1970 to 115 in 1987, a 523 % increase (Figure 1). There is concern that increased groundwater withdrawal may affect the springs. Additionally, with the increase of residential expansion into the area is an increase in associated septic system effluent which could potentially degrade the quality of groundwater discharged from Armstrong and Nelson Springs. Despite the obvious importance of the springs as a natural groundwater and surface water resource and concern regarding continued groundwater development of the area, no detailed hydrogeologic investigation of the origin of the springs has been conducted.

Armstrong and Nelson Springs issue from Cenozoic valley-fill deposits which overlie Paleozoic and Mesozoic rocks including faulted Mississippian Madison Group limestone. Cenozoic valley fill, including both Tertiary and Quaternary deposits, and the Paleozoic Madison Group represent major aquifers throughout much of Montana (Hackett and others, 1960; Groff, 1962a, 1962b, 1965; McMurtrey, Konizeski, and Brietkrietz, 1965; Zimmerman, 1966, 1967; Feltis, 1973, 1977; Sondregger and others, 1982; Wilke, 1983; Donovan, 1985; Downey, 1986; Levings, 1986). The valley-fill and Madison Group aquifers are also often associated with spring development.

The Madison Group often forms springs or discharges water to overlaying alluvium, especially where the Madison Group is significantly fractured and faulted (Feltis, 1973; Huntoon 1976a, 1976b, 1985a, 1985b, 1985c; Krothe and Bergeron, 1981; Wilke, 1983; Wyatt, 1984). Faults and fractures in the Madison Group as well as in other limestone aquifers provide a zone of higher permeability as compared to the primary permeability in the aquifer and so, often serve as a preferrential path of groundwater migration (Huntoon, 1976a, 1981, 1985c; Wilke, 1983; Huntoon and Coogan, 1987). With time, groundwater flow through the trace of faults and fractures may increase as the limestone wall rock dissolves to form well developed solution channels which can eventually carry a large volume of groundwater flow and therefore support a large spring network (Huntoon 1976a, 1981, 1985c; Wilke, 1983; Huntoon and Coogan, 1987). Armstrong and Nelson Springs originate in the immediate vicinity of

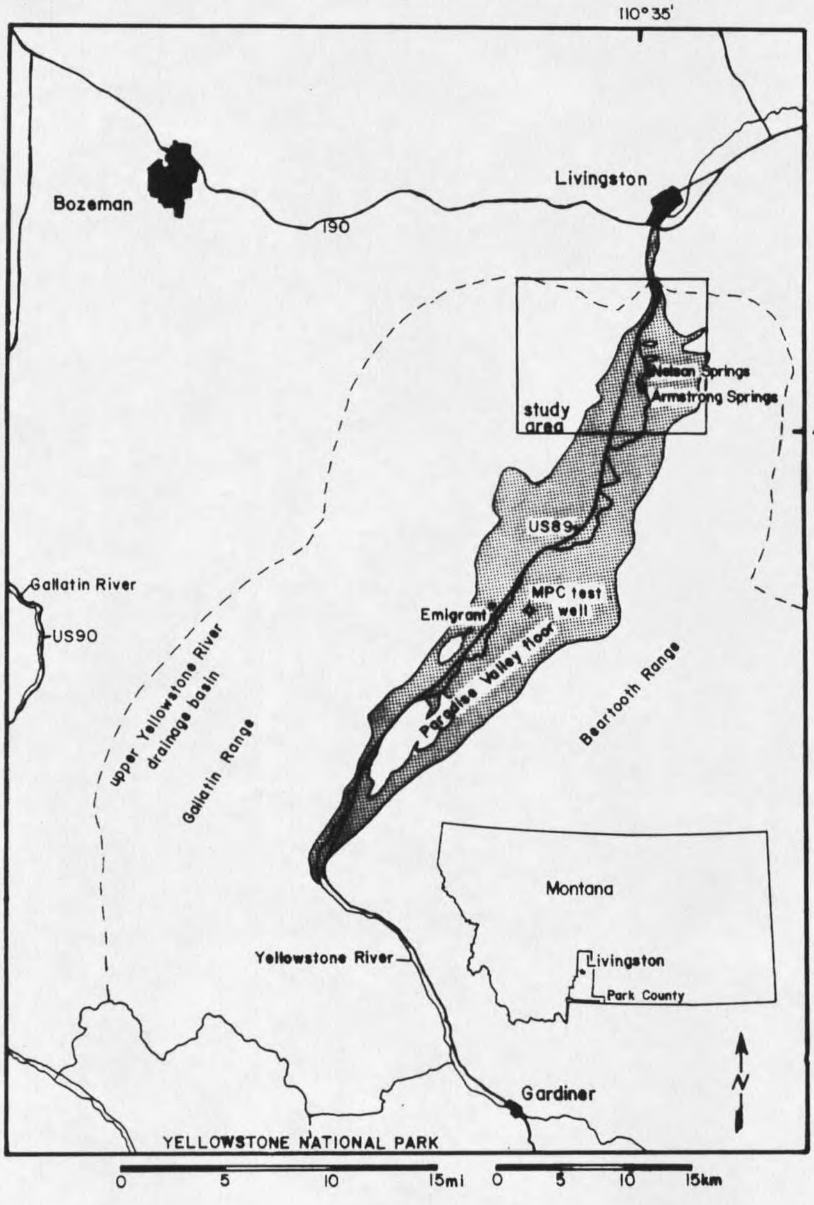


Figure 1. Paradise Valley and the study area. Dotted line denotes the regional Yellowstone River drainage basin. Stippled pattern denotes the distribution of Cenozoic alluvium within the valley floor.

faulted Madison Group rocks which suggests a possible hydrologic connection between Armstrong and Nelson Springs and the Madison Group. In other studies in Montana and Wyoming, leakage to springs and overlaying formations from the Madison Group at depth has been identified based upon groundwater chemical compositions, anomalous groundwater temperature patterns, groundwater gradient patterns, and inference based upon geographic, stratigraphic, and structural relationships between springs and the Madison Group (Feltis, 1973; Huntoon, 1976a, 1976b, 1976c, 1985a, 1985b; Downey, 1986; Sondregger and others, 1982; Wilke, 1983; Wyatt, 1984; Huntoon and Coogan, 1987).

Alternatively, the springs may not be related to the Madison Group. The springs may be stratigraphically controlled, formed at a zone of lesser hydraulic conductivity within the Cenozoic valley fill. Faults which cut the Cenozoic valley fill may influence development of the springs. Also, the springs may simply issue from topographic depressions in the Cenozoic valley fill or where the valley fill is thin due to local subsurface bedrock patterns. Cenozoic valley fill has been determined to yield significant volumes of groundwater to wells and springs. Cenozoic valley-fill groundwater has been identified by the location of wells and springs with respect to geographic, stratigraphic and structural relationships, chemical composition of groundwater, groundwater temperatures, and groundwater gradients (Hackett and others, 1960; Groff, 1962a, 1962b; McMurtrey, Konizeski, and Brietkrietz, 1965; Feltis, 1973; Sondregger and others, 1982; Wilke, 1983; Donovan, 1985; Levings, 1986).

Purpose

A study was undertaken to determine the hydrogeology of Armstrong and Nelson Springs using conventional methods. More specifically, the study addresses the following questions:

(1) Are the springs controlled structurally, stratigraphically, or topographically or do they reflect some combination of these controls?

(2) Which aquifer(s) is (are) the probable source aquifer(s) for Armstrong and Nelson Springs?

(3) Do the springs originate from a single or multiple groundwater flow system?

General Setting

The study area is approximately 16 km south of Livingston, Montana, near the northern terminus of the Paradise Valley (Yellowstone Valley) (Figures 1 and 2). Springs occur on both sides of the Yellowstone River, which flows north through the valley. The study area lies within Townships 3 and 4 South and Ranges 9 and 10 East, and includes all springs known to discharge to the spring creeks (Figure 2). Field investigations were confined to the valley floor.

The Beartooth and Gallatin mountain ranges respectively form the eastern and western margins of the study area. A narrow canyon at the northern terminus of Paradise Valley forms the northern boundary. The

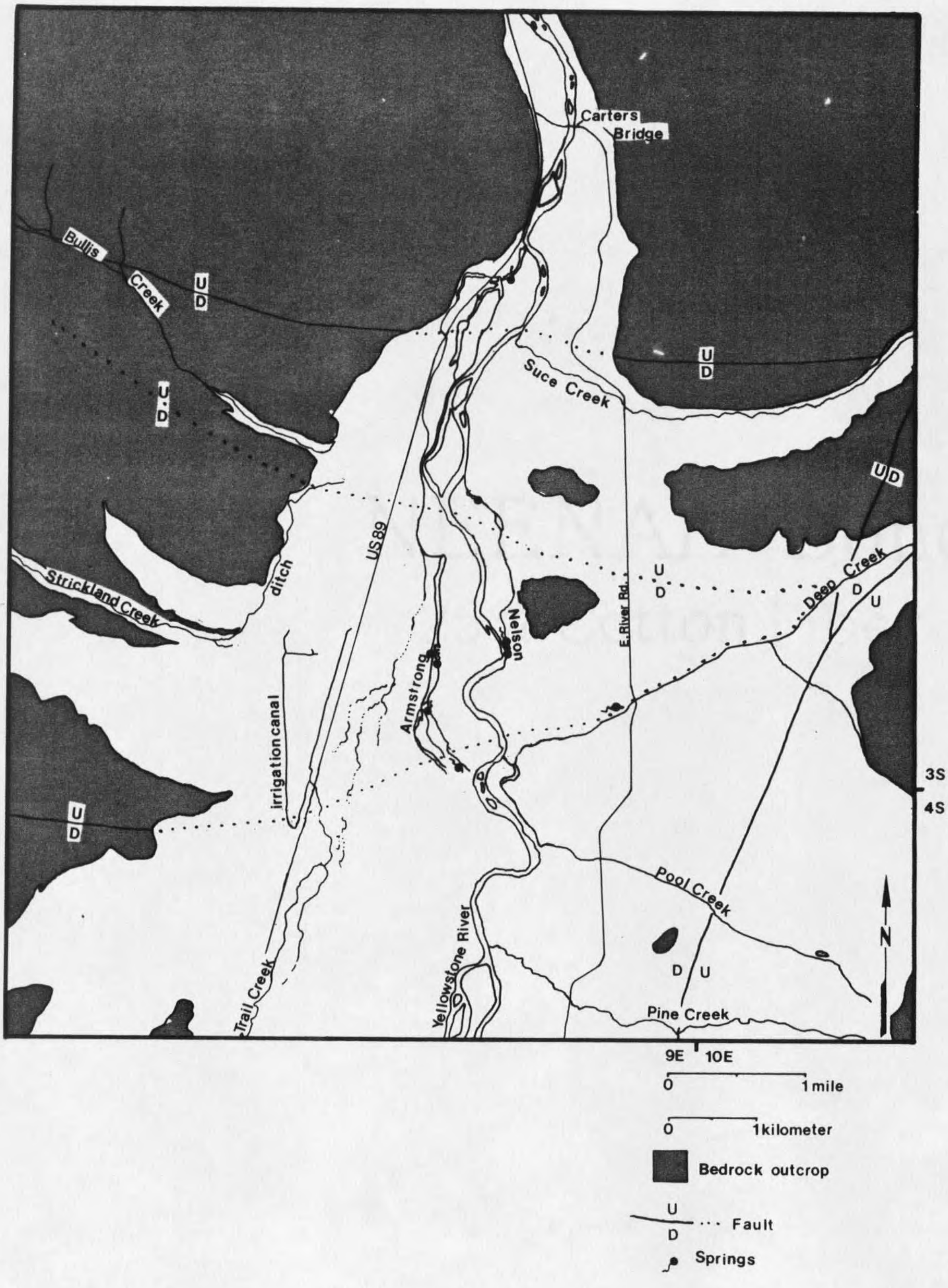


Figure 2. Generalized geologic map of the study area (modified from U.S.G.S. Brisbin topographic 7.5 minute quadrangle, 1981, and Roberts 1964a, 1964b).

southern boundary was arbitrarily drawn upstream of the spring creeks at the approximate position of Pine Creek (Figure 2).

Paradise Valley is a wide, linear fault-block basin which trends northeast for approximately 55 km (Figure 1) (Bonini and others, 1972). The valley is terminated at the north and south ends by narrow canyons cut through bedrock by the Yellowstone River. The valley varies from 0.5 to 10 km in width. Within the study area, the valley ranges from 0.5 to 8 km in width. Topography within the valley floor dips gently to the north dropping from an elevation of 1,520 m above sea level at the southern terminus to 1,390 m above sea level near the northern terminus. The local relief between the bordering mountain ranges and valley floor is approximately 1,500 m. Seven peaks within the Beartooth and Gallatin Ranges border the valley and exceed 2,900 m above sea level.

Mean monthly temperatures for the Paradise Valley area range from a maximum of 13.5°C to a minimum of 1.6°C. The mean annual temperature is 7°C (Ruffner, 1985). The valley floor receives between 36 and 51 cm of precipitation (rain and snow) annually (United States Department of Agriculture, 1971). The surrounding peaks receive between 102 and 127 cm of annual precipitation (United States Department of Agriculture, 1971).

Drainage Systems

The Yellowstone River is the master stream for the region (Figure 1). The river receives water from the many perennial streams which drain the Gallatin and Beartooth highlands and the Yellowstone Plateau. The average discharge for the Yellowstone River based on a 60 year record measured near Livingston is 107 m^3/s (United States Department of the

Interior, 1985). Over this period the river discharge ranged between 1,030 and 17 m^3/s (United States Department of the Interior, 1985). Between October 1986 and February 1988 the Yellowstone River flow varied between 156 and 28 m^3/s (M. White, United States Geological Survey, written communication, 1988).

Four streams in the study area flow from the east and are confluent with the Yellowstone River (Figure 2). In contrast, the flow from only one of three stream channels on the west side of the valley in the study area continues to the Yellowstone River (Figure 2). Strickland and Bullis Creeks lose their entire flow to the alluvial groundwater system prior to reaching the Yellowstone River. Strickland Creek is partially diverted for irrigation purposes, but the lack of any abandoned stream channels on aerial photos or topographic maps indicates the stream has not reached the Yellowstone River in post-glacial time. Trail Creek is the only creek which flows from the west that discharges into the Yellowstone River. Trail Creek is diverted twice for irrigation. Below the second diversion, flow is ephemeral, but the channel is clearly defined on aerial photographs and the topographic map of the area. Thus, flow in the western tributaries is commonly lost to the groundwater system before reaching the Yellowstone River, whereas the eastern tributaries flow directly into the river.

Stratigraphic Setting

The rocks are the framework in which groundwater occurs and moves. In the study area, Precambrian through Tertiary-age bedrock is exposed

(Roberts, 1964a, 1964b) (Table 1, Plate 1). The Paleozoic section is composed primarily of carbonate rocks with some fine grained clastic units (Table 1). The Mesozoic section consists of fine to coarse grained clastic rocks (Table 1). Tertiary volcanic rocks are exposed in outcrop in the southwestern portion of the study area, but no Tertiary-age basin-fill deposits have been formally recognized in outcrop or in the subsurface in the study area (Roberts, 1964a, 1964b). Only a thin veneer (approximately 16 m in thickness) of Quaternary deposits and possibly some Tertiary deposits are present in the central portion of the study area (Bonini and others, 1972).

This thesis focuses on large-discharge springs; therefore, aquifers which characteristically are capable of producing such springs must be identified. Due to the dearth of hydrogeologic information pertaining to the area, the general hydrogeologic properties of the stratigraphic units in the area are inferred from other studies in the region (Table 1). This data will be supplemented later in the thesis with data from the field study. In general, rocks in the study area older than the Madison Group have not been recognized to yield large quantities of groundwater or produce large springs (Groff 1962a, 1962b, 1965; Roberts 1964a, 1964b, 1966; Zimmerman 1967; Konizeski, Brietkrietz, and McMurtrey, 1968; Swenson, 1968; Feltis, 1977; Head and Merkel 1977; Taylor 1978; Miller 1979; Downey 1982, 1986).

Table 1. Generalized stratigraphic section and water-bearing properties of rocks found in the study area. (After Groff, 1962a, 1962b, 1965; Roberts, 1964a, 1964b, 1966; Zimmerman, 1966, 1967; Konizeski, Brietkrietz, and McMurtrey, 1968; Swenson, 1968; Feltis, 1977; Head and Merkel, 1977; Taylor, 1978; Miller, 1979; Downey, 1982, 1986; Wyatt, 1984).

Series	Formation	Maximum Thickness (m)	General Character	Water-Bearing Properties
Quaternary	Recent Alluvium	16+	Modern stream and river deposits of silt, sand, and gravel.	Principal aquifer. Can yield adequate water for small to large irrigation wells.
	Alluvium and Colluvium	42+	Unconsolidated alluvial fan and slope wash deposits of silt, sand, and gravel.	Yields enough to meet domestic and stock needs.
	Glacial Outwash	22+	Unconsolidated sands and gravels.	Yields appreciable domestic supplies.
	Till and Pediment Veneer	?	Unconsolidated and semiconsolidated clay, silt, sand, and gravel.	Too thin and/or localized to be a significant aquifer. Possible small local yield.
---UNCONFORMITY---				
Tertiary	Volcanic	1220	Andesite and basalt flows, breccias, tuffs, conglomerates, sandstones.	May yield small to moderate water quantities where extensively fractured.
Upper Cretaceous	Cody Shale and Frontier Formations	159	Dark shale with some arkosic sands.	Not an aquifer.
Lower Cretaceous	Mowry and Thermopolis Shale	354	Dark gray shale and sandstones.	Yield some water to local wells.

Table 1. Generalized stratigraphic section and water-bearing properties of rocks found in the study area-Continued.

Series	Formation	Maximum Thickness (m)	General Character	Water-Bearing Properties
Lower Cretaceous	Kootenai Formation	85	Sandstone, mudstone, fossil bearing limestone.	Probably not an important aquifer here due to limited presence. Yields water to some minor springs.
---UNCONFORMITY---				
	Morrison Formation	137	Varicolored shale, siltstone, mudstone, and limestone.	Yields water to some small springs. Not generally an aquifer.
Upper Jurassic	Swift Formation	31	Calcareous glauconitic oyster bearing sandstone.	May yield small amounts of water. Not an important aquifer.
	Rierdon Formation	29	Olive gray shale to bluish limestone.	Generally not an aquifer.
	Piper Formation	104	Olive gray shale to argillaceous limestone.	Generally not an aquifer.
---UNCONFORMITY---				
Pennsylvanian	Quadrant Formation	88	Quartzite to well sorted sandstone and dolomite.	Yields water to some small springs. Not an important water source.
------------	Amsden Formation	46	Dolomite, calcareous sandstone, siltstone, and limestone.	Yields some water to local wells. Poor quality water here.
---UNCONFORMITY---				

Table 1. Generalized stratigraphic section and water-bearing properties of rocks found in the study area-Continued.

Series	Formation	Maximum Thickness (m)	General Character	Water-Bearing Properties
Upper Mississippian	Mission Canyon Limestone	61	Massive gray limestone and dolomite.	Yields 1.3 x $10^{-3} m^3/s$ to a well in the area. Generally considered to have appreciable aquifer potential.
Lower Mississippian	Lodgepole Limestone	146	Well bedded chert bearing limestone and dolomite.	
	Three Forks Shale	15	Dark fossiliferous shale with basal chert.	Not an aquifer.
Upper Devonian	Jefferson Dolomite	107	Massive gray limestone and dolomite.	Possibly yields small quantities to local wells.
---UNCONFORMITY---				
Ordovician	Bighorn Dolomite	27	Massive mottled gray and yellow dolomite.	Not an important water source.
Upper Cambrian	Grove Creek and Snowy Range Formations	70	Limestone and shale with limestone pebbles.	Not an important water source.
	Pilgrim Limestone	40	Massive magnesian and dolomitic limestone with flat pebble conglomerates.	May yield small amounts of water if fractured.
Middle Cambrian	Park Shale	67	Multicolored shale and argillaceous limestone, siltstone, and mudstone.	Not water bearing.

Table 1. Generalized stratigraphic section and water-bearing properties of rocks found in the study area-<u>Continued</u>.

Series	Formation	Maximum Thickness (m)	General Character	Water-Bearing Properties
	Meagher Limestone	40	Thin bedded medium gray limestone and yellow gray dolomite.	Not an important aquifer.
Middle Cambrian	Wolsey Shale	85	Green gray calcareous shale and mudstone.	Not water bearing.
	Flathead Formation	30	Gray to red quartz sandstone and quartzite.	Yields small water amounts from fractures.
---UNCONFORMITY---				
Archaen	Undiff.		Gneiss, schist and granite. Pegmatite dikes locally.	Yields small amounts of water if fractured.

Mississippian Madison Group Aquifer

The Madison Group is considered to be an important aquifer throughout much of Montana, Wyoming, North Dakota, and South Dakota (Groff, 1966; Swenson, 1968; Anderson and Kelley, 1976; Huntoon, 1976a, 1976b, 1985a, 1985b; Head and Merkel, 1977; Krothe and Bergeron, 1981; Downey, 1982, 1986; Wyatt, 1984). The Madison Group is composed of the Mission Canyon and Lodgepole Formations in southwestern Montana (Table 1) (Roberts, 1966). The Mission Canyon Formation conformably overlies the Lodgepole Formation (Roberts, 1966). In the study area, the Mission Canyon Formation is composed of massive, very fine to medium crystalline limestone, dolomitic limestone, calcitic dolomite, and dolomite with interbedded solution breccias of dolomite and kaolinite clay (Roberts, 1966). Near the top, the Mission Canyon Formation contains discontinuous karst deposits infilled with red illite clay (Roberts, 1966). The Lodgepole Formation consists of massive to thinly bedded, fine to coarsely crystalline limestone, dolomitic limestone, calcitic dolomite, and dolomite (Roberts, 1966).

The Madison Group aquifer is defined as the Mission Canyon and Lodgepole Formations in the study area. However, from a groundwater standpoint, the Mission Canyon Formation is much more important than the Lodgepole Formation because the Mission Canyon Formation contains a greater percentage of thick-bedded limestones, dolomites, collapse breccias, dissolution features, and less insoluble residues than does the underlying Lodgepole Formation (Roberts 1966, Aram 1979). The fact that the Mission Canyon Formation is the principal component is attested to in the Lewis and Clark Caverns of southwestern Montana. The extensive cavern network is concentrated at the base of the Mission Canyon Formation due to the stoppage of downward groundwater flow by the underlaying relatively insoluble Lodgepole Formation (Aram, 1979).

The Madison Group aquifer yields water to wells throughout Montana at rates which range from 1.2 $m^3/s \times 10^{-3}$ to 5.6 $\times 10^{-1}$ m^3/s (Taylor, 1978). In the Madison Group aquifer primary porosity and permeability is of lesser importance than secondary porosity and permeability (Huntoon, 1976, 1985a, 1985b; Miller, 1976; Head and Merkel, 1977; Budai and others, 1987). Limited primary porosity and permeability is present in the Madison Group as relict depositional intercrystalline porosity. Secondary porosity and permeability in the Madison Group results from dolomitization, tectonic fracturing, and dissolution by migrating meteoric waters. Secondary dolomitization of calcitic limestone to dolomitic limestone increases porosity and permeability (Blatt and others, 1972). Accordingly, permeabilities are greater in the dolomites than in the limestones of the Madison Group (Budai and others, 1987). Fractures within the dolomites are more open than in the limestone beds due to smaller amounts of carbonate cement in the dolomites (Budai and others, 1987).

Karst, solution breccias, and tectonic fractures are also significant to groundwater migration in the Madison Group (Sando, 1974; Aram, 1979; Huntoon, 1985a, 1985b, 1985c, 1987). Tectonic fractures in particular are important groundwater pathways. Dissolution of the carbonate aquifer promotes karst and solution breccia formation near these fractures (Sando, 1974; Aram, 1979). Roberts (1966) noted solution collapse breccias and discontinuous karst deposits within Madison Group outcrops of the study area. These features are pervasive in uplifted Mission Canyon rocks throughout Montana and Wyoming (Keefer, 1963; Sando, 1974; Aram, 1979; Huntoon, 1985a, 1985b). Paleozoic units younger than the Madison Group and Mesozoic rocks which are present in the study area generally produce only small groundwater yields in the region (Table 1) (Groff, 1962a, 1962b, 1965; Zimmerman, 1966, 1967; Konizeski, Brietkrietz and McMurtrey, 1968; Feltis, 1977; Miller, 1979; Wyatt, 1984; Downey, 1986). In some parts of southwestern Montana, however, Tertiary and Quaternary deposits do produce springs or large well discharges (Hackett and others, 1960; Groff, 1962a; McMurtrey, Konizeski, and Brietkrietz, 1965; Sondregger and others, 1982; Wyatt, 1984; Levings, 1986).

In order to provide a better interpretation of the source of the springs, the Cenozoic valley fill is separated into the alluvial fan aquifer and the Yellowstone aquifer (glacial outwash and recent alluvium) (Plate 1). These stratigraphic divisions are similar to Roberts' (1964a, 1964b) divisions. The glacial outwash and recent alluvium of Roberts (1964a, 1964b) have been grouped as the Yellowstone aquifer to reflect interpretation as a single hydrostratigraphic unit.

Alluvial Fan Aquifer

In the study area alluvial fans coalesce and descend to the Yellowstone River which truncates the toe of the fans to form bluffs. The alluvial fans consist of mixed silt, sand, pebble, cobble, and boulder size gravels which were probably deposited by fluvial and glacio-fluvial activity (Roberts, 1964a). Studies which have been conducted in the study area by other researchers have not identified the presence of

Tertiary-age deposits (Roberts, 1964a, 1964b; Bonini and others, 1972; and Personius, 1982). For the purpose of this study, the alluvial fan aquifer is considered to include all alluvial materials between the Beartooth mountain front and the Yellowstone aquifer below material mapped as alluvial fans.

The maximum thickness of the alluvial fan deposits is unknown, but well reports indicate the maximum thickness of the deposits exceeds 42 m. In the adjacent Gallatin Valley, the Bozeman alluvial fan, which is underlain by both Tertiary and Quaternary deposits, may yield small to moderate quantities of groundwater to wells according to the degree of sorting and the amount of silt and clay present (Hackett and others, 1960). In general, stringers and lenses of gravel and sand are the source of most of the groundwater within the Bozeman fan (Hackett and others, 1960). Transmissivities of the Tertiary Bozeman fan deposits range from 4×10^{-7} m^2/s to 3.8×10^{-4} m^2/s. Transmissivities of the Quaternary Bozeman fan deposits range from 6.4×10^{-4} m^2/s to 9.3×10^{-3} m^2/s.

Yellowstone Aquifer

Glacial outwash deposits (Pinedale) and Recent Yellowstone alluvium compose what is referred to as the Yellowstone aquifer (Plate 1). This aquifer may directly overlie Paleozoic bedrock if Tertiary basin-fill is absent in the study area. The aquifer is composed of semiconsolidated to unconsolidated interbedded clay, silt, sand, and gravel (Roberts, 1964a). The maximum thickness of the aquifer is unknown; however, Montana Power Company (MPC) RP624-Hobbs test well drilled approximately 12 km south of the study area encountered 243 m of alluvium (Montagne and Locke, 1989). The alluvium at the MPC test well site includes Pleistocene glacial outwash deposits, Holocene fluvial deposits, and possibly extensive pre-last-glacial (pre-Pinedale) gravel deposits (Pierce, 1979; Montagne and Chadwick, 1982; Locke, 1986; Montagne and Locke, 1989). The maximum thickness of the Pinedale glacial outwash deposits may be as much as 45 m, 9 km south of the study area based upon relationships between the height of the head of the glacial outwash above the current level of the Yellowstone River (Locke, 1986). The Pinedale glacial outwash deposits and Recent fluvial deposits are also present in the study area (Roberts, 1964a; Montagne and Chadwick, 1982; Locke, 1986; Montagne and Locke, 1989). The Yellowstone aquifer alluvium exceeds 22 m in thickness near the southern limit of the study area as indicated by domestic water well reports and is no more than 15.5 m thick at the northern terminus of Paradise Valley (Kirby, 1940). The presence of pre-last-glacial alluvium in the study area is unknown, but if present, it too must be greatly thinner than at the MPC test well site.

Yellowstone River alluvium is an important aquifer north of the study area near Livingston (Groff, 1962a). The thickness of the alluvium there is reported by Groff (1962a, Table 1) to be 16 m. The deposits are thickest adjacent to the Yellowstone River and thin laterally away from the river (Groff, 1962a). Groundwater is pumped from the recent alluvium for municipal, domestic, and industrial purposes. Reported well yields range from $1.2 \times 10^{-3} m^3/s$ to $57 \times 10^{-3} m^3/s$ (Groff, 1962a).

In the adjacent Gallatin Valley, Quaternary stream channel deposits yield copious amounts of groundwater to wells (Hackett and others, 1960). The Gallatin River channel deposits consist of cobbles and gravels with intermixed sand, clay, and silt. The deposits are generally thickest toward the center of the Gallatin Valley. Hackett and others (1960) calculated a transmissivity of 2.4×10^{-3} m^2/s for Tertiary-age stream channel alluvium in the Gallatin Valley. Transmissivities of the Quaternary stream channel alluvium in the Gallatin Valley range from 7.2 $\times 10^{-3}$ m^2/s to 9.6×10^{-2} m^2/s.

Structural Setting

Geologic structures often influence groundwater movement, occurrence, and spring development. Geologic structures have been recognized as particularly important in the Madison Group (Kiersch and Hughes, 1952; Maxey, 1968; Williams, 1970; Feltis, 1973; Huntoon, 1975, 1976a, 1976b, 1981, 1985a, 1985b, 1987; Miller, 1976; Stringfield and others, 1979; Trudeau and others, 1983; Mayo and others, 1985; Downey, 1986; Huntoon and Coogan, 1987). Both thrust faults and extensional faults are known to control spring development through: (1) juxtaposition of rock units with contrasting hydraulic conductivities so as to create a barrier to flow, (2) development of a semi-impervious zone or gouge layer along the fault zone which retards groundwater flow across the fault, and (3) development of fault breccias which increase the vertical/horizontal hydraulic conductivity ratio and which may induce greater vertical flow (Williams, 1970; Huntoon, 1981, 1985a, 1985b; Trudeau and others, 1983; Mayo and others, 1985). The presence of faults in the immediate vicinity of Armstrong and Nelson Springs suggests the structures may control groundwater migration. Therefore, an understanding of the local structures is critical to interpretations with respect to spring formation.

In southwestern Montana, Sevier, Laramide foreland, strike-slip systems, and Basin and Range structures are present (Reynolds, 1979; Lowell, 1983; Schmidt and Garihan, 1983) (Figure 3). In the study area,

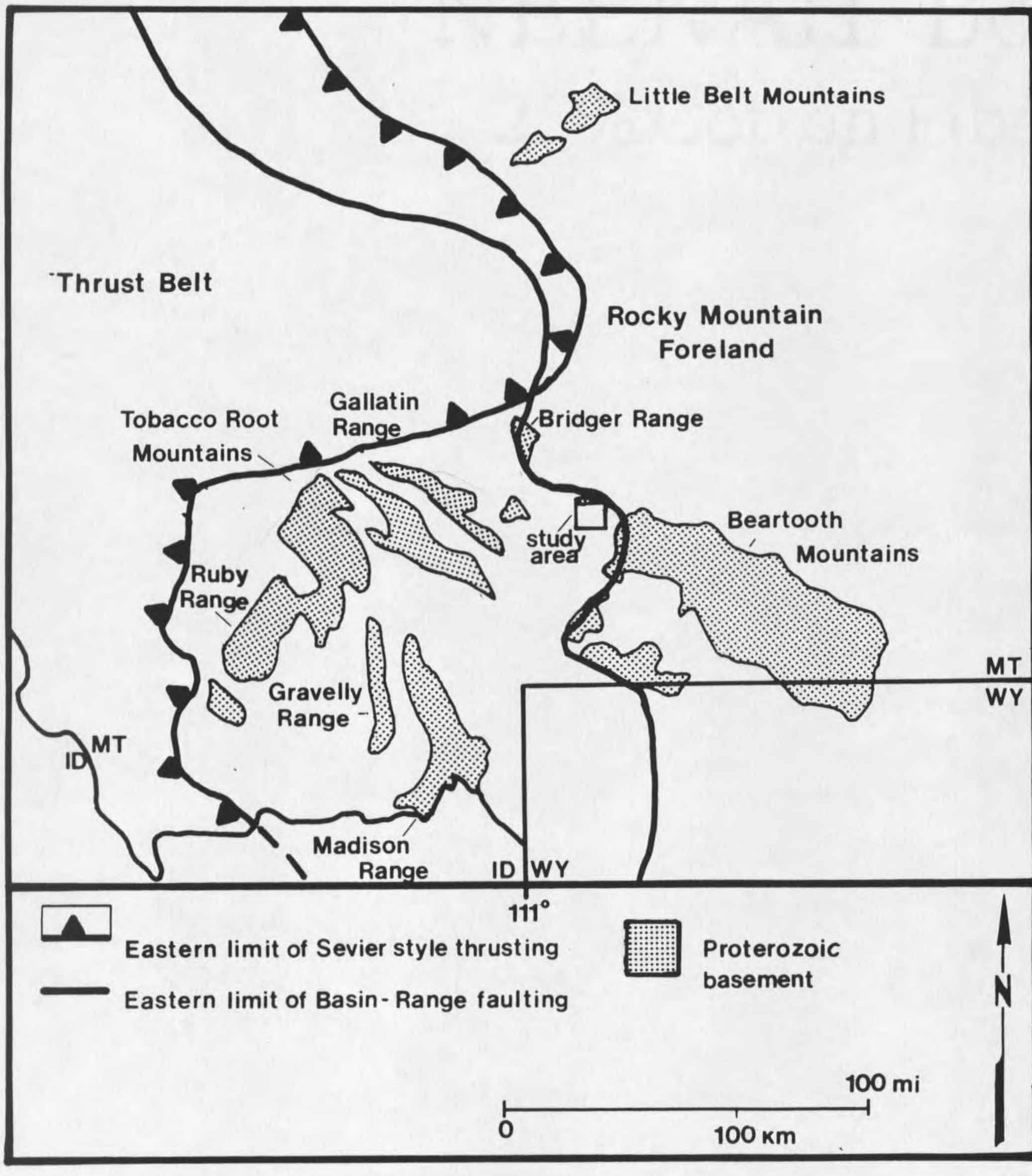


Figure 3.Extent of Sevier style and Basin and Range tectonism (modified from Personius, 1982; and Miller, 1987).

both Laramide foreland and Basin and Range structures exist and are thought to represent a reactivation of Proterozoic basement weaknesses (Foose and others, 1961; Reid and others, 1975; Reynolds, 1979; Brewer and others, 1981; Erslev, 1982; Montagne and Chadwick, 1982; Personius, 1982; Schmidt and Garihan, 1983; Robbins and Erslev, 1986) (Plate 1).

Laramide Foreland Structures - Northern Paradise Valley

At least two Laramide foreland structures trend northwest across the axis of Paradise Valley (Roberts, 1964a, 1964b; Miller, 1987) (Plate 1). Downplunge projections were used to constrain the cross-section interpretation of reconstructed outcrop patterns (Plates 1 and 2). The northernmost structure consists of the Canyon Mountain anticline within the hanging wall of the Suce Creek thrust fault. The Canyon Mountain anticline plunges to both the northwest and the east-southeast and is overturned to the southwest (Plates 1 and 2). The northern limb of the fold dips to the northeast (Plate 2, section A-A'). The fold contains Precambrian through Mesozoic-age rocks. The Mississippian Madison Group occupies the topographic crest of the anticline. The Suce Creek thrust fault has displaced Archean basement southwest over the Cretaceous Kootenai Formation (Roberts, 1964a, 1964b, 1972). The axial trace of the anticline parallels the trace of the Suce Creek fault. The Suce Creek fault dips northeast at 50° - 60° based upon a three point solution of the fault from Roberts, 1964a and 1964b (Plate 2).

The Center Hill anticline and Center Hill fault are the southernmost Laramide structures in the study area (Plate 1). The Center Hill anticline is an asymmetrical, eastward-plunging anticline with a shallow dipping northern limb and a near vertical southern limb (Plate 2, section A-A') (Roberts, 1964a, 1964b). Mississippian Madison Group rocks occupy the topographic crest of the anticline. Center Hill anticline lies within the hanging wall of the Center Hill fault, a northeasterly dipping fault, which places Cretaceous Kootenai and older Mesozoic and Paleozoic formations over the Cretaceous Mowry and Thermopolis Shales (Plates 1 and 2).

South of the Suce Creek fault, in the northwest corner of the study area, a series of klippen of Paleozoic formations (Cambrian-Mississippian) overlie Cretaceous Mowry and Thermopolis shales (Plate 1). These klippen have been interpreted to be gravity-slide blocks which moved downward from structural and topographic highs of Canyon Mountain anticline (Skeels, 1939; Foose and others, 1961). Another interpretation (Schmidt and Garihan, 1983) suggests the klippen may represent bedding-plane thrusts resulting from the overcrowding of the synclinal limb of the overturned Canyon Mountain anticline. The rocks of the klippen are extensively shattered. The shattered nature of the rocks suggests the method of emplacement was rapid such as would be expected if the klippen represent a gravity slide. No matter which interpretation is preferred, the Madison Group outcrops in this portion of the study area are not interpreted to continue as subcrops which pass beneath the springs (Plate 2).

Indirect evidence for a third foreland structure located between the Suce Creek and Center Hill faults has been discovered during this study and will be discussed in detail later.

Basin and Range Structures
- Paradise Valley

Following compressional tectonism, tensional forces became dominant and Basin-and-Range-style normal faults were developed throughout southwestern Montana and much of the western United States as well as in Paradise Valley during the Miocene (Reynolds, 1979). Paradise Valley is considered to be a half-graben tilted to the east along the reactivated, previously existing, Deep Creek normal fault (Reynolds, 1957; Bonini and others, 1972; Montagne and Chadwick, 1982; Personius, 1982) (Plate 1). The Deep Creek fault trends northeast, bordering the western flank of the Beartooth uplift (Plate 1), and represents the easternmost extent of Basin and Range extension in Montana (Reynolds, 1979; Personius, 1982). The fault dips to the west-northwest at an angle of between 80° and 90° (Bonini and others, 1982). Dip-slip motion along the fault has offset the Precambrian through Mesozoic-age strata cut by the fault (Plate 1) (Reid and others, 1975).

The amount of displacement along the Deep Creek fault varies greatly throughout Paradise Valley. The displacement along the Deep Creek fault in the northern part of the study area is only approximately 830 m while the amount of displacement increases to the south to as much as 1,800 m between Pool Creek and Pine Creek (Plate 1) (Reid and others, 1975). Further south, beyond the study area, displacement along the Deep Creek fault may be as much as 6,100 m (Bonini and others, 1972; Erslev 1982).

Cenozoic Geologic History

The hydrogeologic framework of the Paradise Valley directly reflects the Cenozoic evolution of the valley. The Cenozoic evolution of the valley is characterized by periods of structural deformation, volcanism and glaciation (Foose and others, 1961; Roberts, 1972; Montagne and Chadwick, 1982). The Tertiary Period is characterized by both compressional and extensional tectonic deformation resulting in compressional uplift and erosion of highland areas and the subsequent formation of the Paradise Valley with the deposition of basin-fill. The Tertiary history of uplift and structural deformation throughout Montana and Wyoming, including Paradise Valley, may have prompted the later development of Cenozoic karst in the Madison Group aquifer (Huntoon, 1976a, 1976b, 1976c, 1985a, 1985b, 1987; Aram, 1979).

Tertiary structural activity in the Livingston area includes Laramide compressional uplift of the Beartooth block (Foose and others, 1961). Laramide compression persisted into late Paleocene or early Eocene (Roberts, 1972; Lageson, 1989). Movement (pre-Basin and Range extension) along Deep Creek normal fault may have preceded the initiation of Laramide compression as suggested by the apparent truncation of the Deep Creek fault at its northern extent by the Laramide Suce Creek fault (Richards, 1957) (Plate 1). The Deep Creek fault was later reactivated during the Eocene. Chadwick (1969) noted a 10° southeastward dip in earliest Eocene lava flows and laharic units to the west of the study area along Eightmile Creek in the northern Gallatin Range. These tilted rocks are overlain by later Eocene horizontally bedded volcanic rocks. The tilted Eocene volcanic rocks may reflect movement along the Deep Creek fault which occurred after deposition of the early Eocene volcanic rocks, but prior to deposition of the overlaying horizontally bedded Eocene volcanic rocks (Chadwick, 1969; Montagne and Chadwick, 1982; Personius, 1982).

During Miocene time, Basin and Range extension occurred throughout southwestern Montana as well as the Paradise Valley resulting in uplifts of the mountain blocks and the development of half-grabens (Reynolds, 1979). Miocene deposits unconformably overlay Eocene volcanic rocks in the southern portion of the valley. The presence of the Miocene deposits indicates that at least the southern portion of Paradise Valley was a depositional basin during the Miocene. Barnosky and Labar (1989) conclude that the Miocene deposits, the Hepburn's Mesa Formation, are perennial saline lake deposits which suggests an internal drainage was developed in the Paradise Valley during the Miocene. Active Miocene deposition to the south suggests that similar Tertiary basin-fill deposits could be present in the study area. However, the gravity surveys of Bonini and others (1972) suggests that such deposits may not be extensive if present at all. The present topographic relief in the Paradise Valley area was created by uplift and subsequent erosion since the Miocene (Foose and others, 1961). Movement along the Deep Creek fault has persisted through the Quaternary (Personius, 1982).

The Quaternary is characterized by tectonism associated with the Deep Creek fault as evidenced by the presence of Recent fault scarps (Personius, 1982). The Quaternary is also characterized by glaciation and subsequent fluvial adjustments after glaciation through erosion and deposition of alluvium (Montagne and Chadwick, 1982). During the Pleistocene Epoch (Bull Lake and Pinedale time), a piedmont outlet glacier of the Yellowstone ice sheet flowed down the valley at least once based upon distributions of scattered glacial erratics, relationships between individual moraines, and various geomorphic evidence (Pierce, 1979; Montagne and Chadwick, 1982; Montagne and Locke, 1989). Multiple advances into the Paradise Valley are suggested by analogy to adjacent outlet and alpine glaciers and local stratigraphy. The extent that the Bull Lake piedmont outlet glacier advanced into Paradise Valley is unknown due to the burial or destruction of any present Bull Lake moraines by the subsequent Pinedale piedmont outlet glacier advance (Pierce, 1979; Montagne and Chadwick, 1982; Locke, 1986; Montagne and Locke, 1989). However, the thickness of alluvium encountered in the MPC test well located approximately 12 km south of the study area (Figure 1) suggests extensive glaciation preceded the final glacial advance in the area (Montagne and Locke, 1989).

The Pinedale piedmont outlet glacier flowed down the Paradise Valley to within 9 km of the study area. The terminus of the Pinedale piedmont outlet glacier is marked by the Eightmile terminal moraines (Pierce, 1979; Montagne and Chadwick, 1982; Locke, 1986; Montagne and Locke, 1989). The Pinedale piedmont outlet glacier deposited an extensive apron of glacial outwash fan deposits on the floor of the Paradise Valley (Plate 1) (Pierce, 1979; Montagne and Chadwick, 1982). Alpine glaciers flowed down the Deep Creek and Pine Creek drainages and deposited terminal moraines and glacio-fluvial outwash fans on the valley floor in the study area (Roberts, 1964a; Pierce, 1979; Montagne and Chadwick, 1982) (Plate 1). Since deposition, the glacial deposits have been incised to near the pre-last glacial level by the Yellowstone River drainage system (Pierce, 1979; Locke, 1986; Montagne and Locke, 1989).

METHODS

The data collection for the thesis consisted of obtaining drillers' well reports wells in the study area to create a well inventory, and the field investigation. The field investigation consisted of field verification and modification of existing geologic maps, reconnaissance and mapping of individual Armstrong and Nelson Springs, collection of spring discharge and temperature measurements, inventory and measurements of selected wells, aquifer tests, and collection of water samples for chemical analysis.

Well Inventory

Drillers' well reports were collected for Townships 3 and 4 South, Ranges 9 and 10 East from the Montana Department of Natural Resources and Conservation, the Montana Bureau of Mines and Geology (MBMG), and from the Park County Courthouse. Well report information is presented as appendices. Well inventory, well measurement, and aquifer testing were conducted simultaneously to maximize efficiency. Not all wells present in the field area were field tested and verified. Permission to use the wells and position of the wells with respect to known aquifers and other wells, determined which wells were included in the study. The locations of wells included in well reports were field checked and verified for field tested wells used in this study. Well reports were not available for some wells. Well information obtained in the field is included in the well report data presented in Appendix A.

Site Numbering System

Wells are assigned numbers in accordance with the U.S. Bureau of Land Management system of land subdivision (Figure 4). The "D" prefix indicates the well is east of the Montana Principal Meridian and south of the base line. For all sites in the study area, "D" is the first character of the description. Therefore, the "D" will not be included in the descriptions. The first numeral indicates the township, the second the range, and the third the section in which the well is located. Lowercase letters that follow the section number indicate the location of the well within the section. The first letter denotes the quarter section (160 acre tract), the second the quarter-quarter section (40 acre tract), and the third the quarter-quarter-quarter section (10 acre tract). These subdivisions are designated a, b, c, and d, in a counter-clockwise direction, beginning in the northwest corner. For example, a well located in the SE1/4 NW1/4 sec. 4, T. 6 N., R. 2 E., would be numbered A6-2-4ac (Figure 4).

Field Mapping

Geologic and topographic maps at 1:24,000 scale and aerial photos at 1:20,000 scale were used for field mapping purposes. Strikes and dips in addition to those in Roberts (1964a, 1964b) were measured at outcrop locations critical to interpretations and cross-sections (Plate 1). Only the dips and strikes measured during this study are presented in Plate 1.

Sections within the Madison Group limestone were critically examined for evidence of faulting and completeness of section.

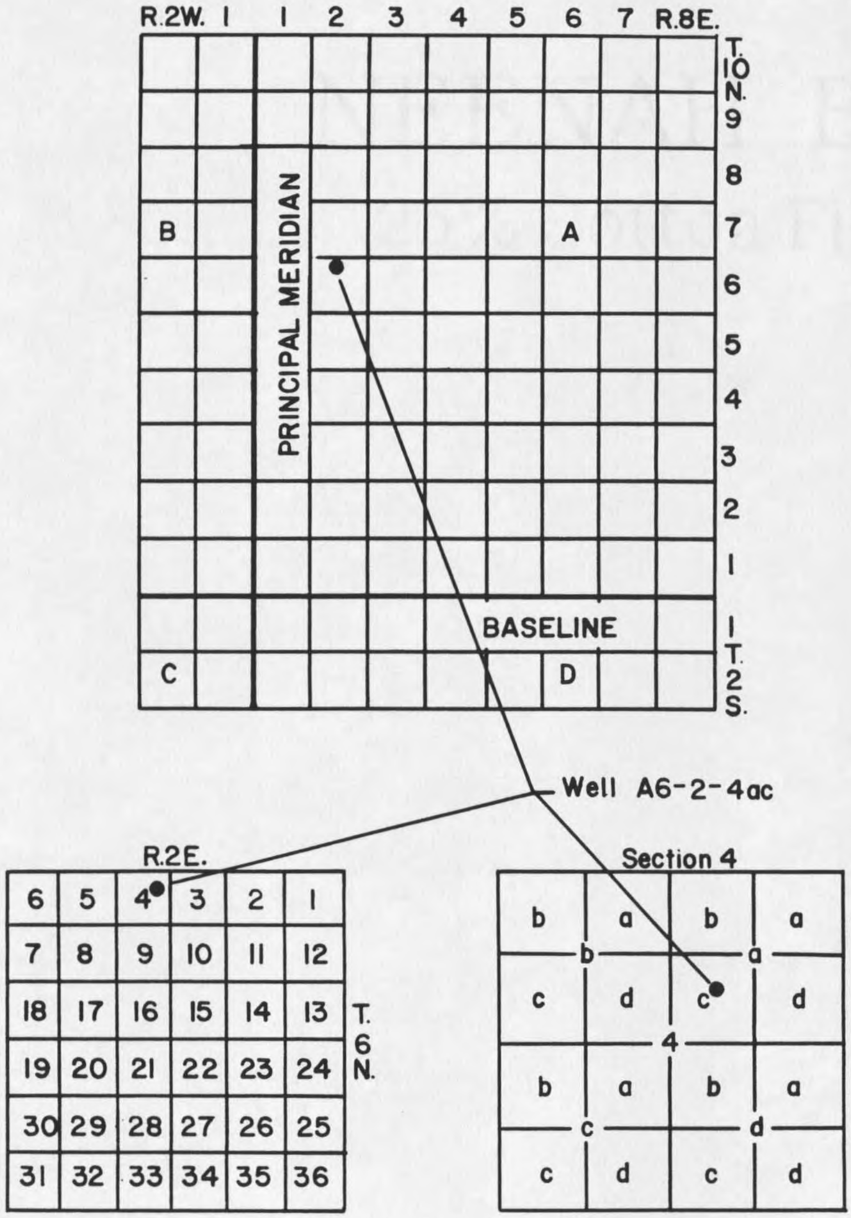


Figure 4. U.S. Bureau of Land Management surveying system. (From Lorenz and McMurtrey, 1956.)

Discharge Measurements

Seasonal changes in discharges of the spring creeks and the master Armstrong Spring (A5) (Plate 3) were measured periodically throughout the study at selected stations (Plate 3). The discharge measurement methods used are prescribed in the National Handbook of Recommended Methods for Water Data Acquisition (United States Department of the Interior, 1982, Mid-Section Method p. 1-24).

A General Oceanics propeller current meter was used for the discharge measurements. Prior to field use, the calibration of the meter was checked using a flume and an electronic current meter belonging to the Montana State University Civil Engineering Department. The calibration of the electronic meter is assumed correct as the calibration of the meter is frequently checked by the faculty of the Civil Engineering Department according to Dr. Al Cunningham (oral communication, 1987). The calibration of the General Oceanics meter was checked over a range of flows from the lowest velocity the General Oceanics meter could detect (10 cm/s) to higher velocities (41 cm/s) (Table 2). Repeated measurement at velocities of 20 cm/s and 41 cm/s with the General Oceanics current meter yielded errors of - 1 to -11 % with respect to the electronic flow meter. The average spring creek velocity measured during the study was 31 $\pm$ 17 cm/s. Given that the average velocity measured in the field and the standard deviation both generally fall within the velocities measured to calibrate the General Oceanics flow meter (Table 2), the mean of the errors estimated by the calibration tests (-4.5 %) suggests the General Oceanics may underestimate discharge by 4.5%. Discharge measurements are reproducible to ± 0.9 cm/s at a velocity of 20 cm/s and to ± 0.7 cm/s at 41 cm/s using the General Oceanics flow meter.

Table 2. Current meter calibration.

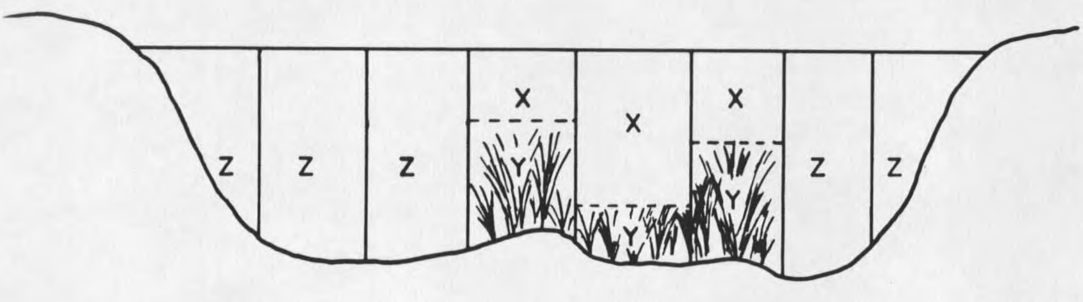
General Oceanic (cm/s)	Electronic (cm/s)	Difference (%)
10.1	10.4	-3.0
19.8	21.7	-9.6
19.8	21.7	-9.6
20.0	18.9	5.5
20.1	22.0	-9.5
22.0	24.4	-10.9
39.7	40.3	-1.5
40.9	41.5	-1.5
41.0	41.5	-1.2

Sections of the stream chosen for measurements were relatively straight and free of obstruction at the start of the field season. The current meter was oriented parallel to the mean flow direction and adjusted to the 0.6 depth below the water surface. The current was measured for 30 seconds at each station. Where the water was too shallow or the current too slow, the floating stick method was used. The surface velocity from the floating stick method was corrected to the 0.6 velocity depth by multiplying the surface velocity measurement by a factor of 0.8 (United States Department of the Interior, 1982). The current was measured at regular intervals at each measurement station where the stream channels were continuous. The measurement interval used depended upon the width of the stream at the particular station. Measurement intervals ranged from 0.5 m to 0.15 m depending upon the stream width at the particular station.

In-stream vegetation proved to be a major problem in measuring discharge. Where the vegetation established on the stream floor was well-developed and tall, flow was often channelized. In such instances, the velocity was not measured at regular intervals across the channel. Instead, velocity measurements were taken at every channelized position regardless of their spacing. Without measuring each channelized position, the flow would have been greatly underestimated.

Where stations were in areas of dense, tall vegetation, the total stream depth and the depth between the top of the vegetation and water surface were measured (Figure 5). The velocity was then measured at 0.6

Figure 5. Discharge measurement methodology for a vegetated stream channel. x = unvegetated partition, y = vegetated partition, z = unvegetated partition.



the depth from water surface to vegetation top (Figure 5). The discharge in the vegetated portion of the partition was calculated using only 25 % of the velocity measured in the unvegetated portion of the same partition (Figure 5). This coefficient was selected because the Mannings "n" value indicates velocity in dense vegetation is one quarter of the velocity for the same channel with no vegetation (Linsley and others, 1982).

$$\text{Mannings Equation: } u = 1.49R^{2/3}\, S^{1/2}/n$$

u = water velocity

R = hydraulic radius

S = energy gradient = channel slope

n = Manning resistance coefficient

The "n" value for a clean straight channel is 0.029 (Linsley and others, 1982). At positions where the vegetation was dense the channel was very weedy, flow was sluggish. A channel of this description has a "n" value of 0.112 (Linsley and others, 1982). The calculated discharge values for each station in the measured section were then summed for the vegetated and the open portion above the vegetated portion of each partition (Figure 5).

The practice of velocity adjustment below the top of the vegetation is nonstandard. However, the procedure based on change in "n" for only the vegetated portion of the channel is equivalent to the usage of the "n" value assigned by Manning for streams choked to the surface with vegetation. Data presented above suggests that portions of streams choked with vegetation are on average capable of passing only 25 % of the discharge the same stream is capable of passing when void of vegetation. Therefore, this method was used to generate a first approximation of the discharges of the springs and it is considered to produce a better estimate of spring discharge than if the procedure had not been used.

The discharge-calculation procedures used in channel sections lacking vegetation followed the mid-section method (United States Department of the Interior, 1982). The method takes each measured station as an individual partition. The discharge of each partition is calculated by multiplying each partition area by the velocity measured in that partition. The discharge measurement for each partition is summed giving the total discharges for that stream section.

In order to ease the monitoring of stream discharges, staff gages were installed in order to develop rating curves for a few of the sites (Plate 3, stream measurement stations SA3, SA4, SA5, SA7). Rating curves equate stream stage with stream discharge, providing a quick approximation of stream discharge. Unfortunately, only the rating curve generated for SA4 is considered to be reliable, (R^2 = 0.97) and therefore usable (Figure 6). The equation for the SA4 rating curve ($y=2.63x-0.97$, where x=stage(m) and y= log discharge (m^3/s)) was computed using the Lotus spreadsheet program. The rating curve is reliable to within $\pm$ 0.03 log m^3/s. Rating curves can be greatly impacted by channel modification, variable backwater and vegetation (United States Department of the Interior, 1982). Channel modification and vegetation proved to be problems at sites SA7 and SA3 (Plate 3) and resulted in the loss of rating curves for those stream measurement stations.

Discharge was measured above and below the spring areas in an attempt to determine the contribution of individual spring areas by the difference method. However, the discharge of any particular spring other than springs A5 and N1 was less than the sensitivity of the discharge measurement method used and therefore the difference method proved to be an inadequate procedure to measure the discharge of individual channel spring.

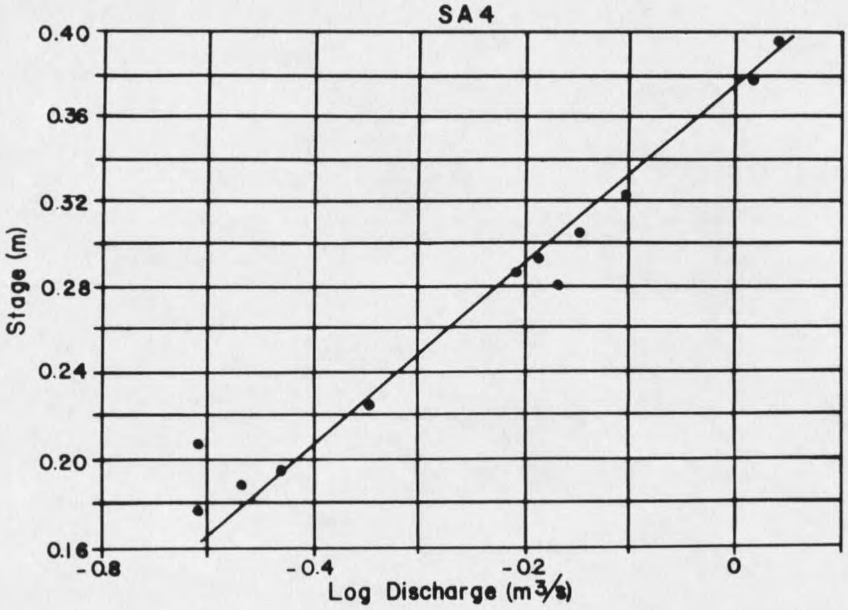


Figure 6. SA4 rating curve. See Plate 3 for location.

The total discharge of Armstrong Spring Creek was measured at station SA7 (Plate 3). The total discharge for all channel springs included in the study was measured at stations SA5 and SA6 (Plate 3). The discharge of spring A5 was not directly measured. Rather, the discharge of spring A5 was determined by calculating the difference in total spring creek discharge measured at station SA7 and the combined discharges measured at stations SA5 and SA6 (SA7 - (SA5 + SA6)). Discharge measured at site SN1 (Plate 3) measures contribution from spring N1 and some channel springs immediately below N1. Discharge measured at SN2 is the total discharge of Nelson Spring Creek.

Groundwater Measurements

Static groundwater levels in wells were measured between September 29 and November 13, 1987 (Figure 7). Depth to water in each well was measured using either a Lufkin Peerless Nubian 300 ft (91 m) steel tape, graduated in 0.01 ft (0 .003 m) increments, or with a Fisher M-Scope 300 ft (91 m) electrical tape. Reported depth to water is from the ground surface. The measuring point is corrected to ground surface which is reported in the well report data (Appendix A). Prior to measurement, the owners were questioned about recent use of the well to determine if the measured level could be assumed representative of true static conditions. If the wells had not been pumped for at least two hours, the water levels in the wells were measured. Recovery tests conducted in the field area indicate wells are fully recovered within two hours. All reported levels are believed to be true static measurements.

Ground-surface elevations were estimated from the Brisbin and Chimney Rock 7.5 minute topographic quadrangle maps. The contour interval of the maps is 40 $\pm$ 20 ft (12 $\pm$ 6.6 m). The static water-table elevation at each well point was calculated by subtracting the depth to water from ground surface from the estimated ground-surface elevation at that position. The method of estimating ground-surface elevations from topographic maps limits the resolution at which subtle groundwater flow patterns can be interpreted. An instrument survey of the ground-surface elevation would have greatly enhanced interpretations of groundwater flow patterns. However, given the size of the study area, the limited time frame, and the project budget, an instrument survey was not practical. Furthermore, the concentration of measured wells along the steep valley margins suggests that the significance of the ground-surface elevation estimates will be minimized. For the purpose of this study, which was to gain an understanding of the basic pattern of groundwater flow on the

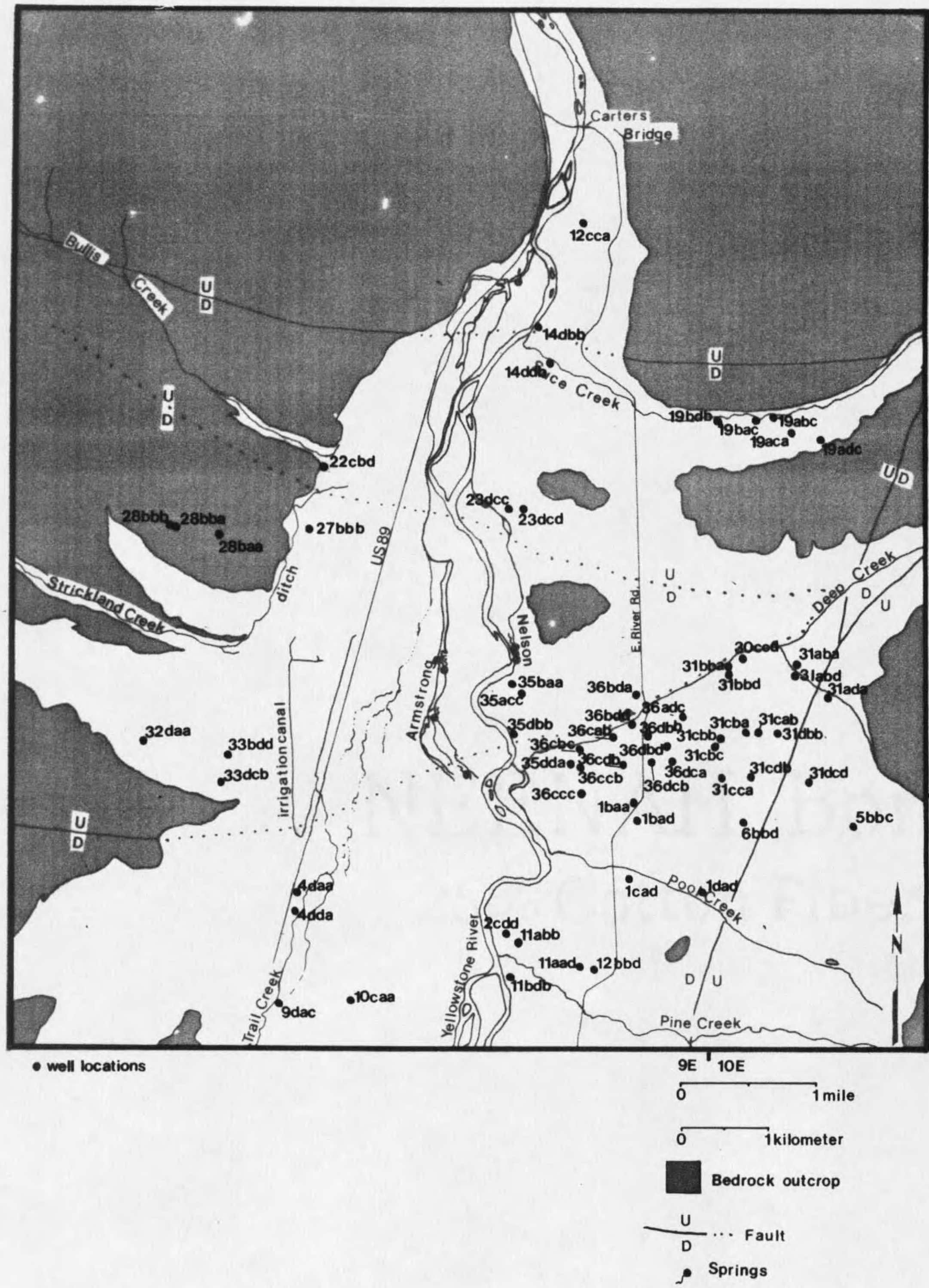


Figure 7. Location of wells used in this study. Numbers do not show township and range since this data is contained in the map margins.

scale of the entire study area, rather than an understanding of subtle groundwater flow patterns at specific locations within the study area, this procedure proved adequate.

Aquifer Test Field Methods

The aquifer tests consisted of a 20 minute pumping period and a 20 minute recovery period. The 40 minute test period was used because: (1) the test period was acceptable to well owners who needed to use their wells, (2) the 40 minute test proved successful in a groundwater study conducted near Bozeman (Moore, 1984), (3) the 40 minute test could be conducted for all wells tested, and (4) the 40 minute test length was time efficient. The short tests allowed more tests to be conducted; longer tests would have resulted in fewer tests.

Where the discharges were great enough to produce a measurable drawdown, the 40 minute test proved successful in this study. The cone of depression generally stabilized within the 20 minute pumping period, and the wells essentially recovered within the initial 20 minutes after the pump was stopped. Tests that proved unsuccessful failed because the pumps were not of large enough capacity to produce appreciable drawdown in the wells.

During the tests, the wells were pumped at maximum pump capacity. Two open spigots were required to keep the pumps running. The discharge of the wells was measured using a 5 gallon (19 l) bucket and a stop watch periodically throughout the test. A Fisher M-Scope electrical tape was used to measure drawdown because of the ability to take rapid measurements without removing the tape from the well.

Groundwater Temperatures

Groundwater temperatures were measured in wells and springs with a Taylor pocket thermometer calibrated in 1°C increments. Wells were operated at maximum capacity for a minimum of ten minutes prior to measurement of the groundwater temperature in order to ensure measurement of fresh formation water. The temperature measurement was collected at the spigot closest to the well by holding the thermometer in the discharge stream. Temperatures were repeatedly measured until the temperature stabilized and the readings could be replicated.

Groundwater temperatures were measured at springs by inserting the thermometer into the spring orifice. The thermometer was allowed to sit in the spring at least five minutes prior to taking a reading. The reading was taken immediately upon withdrawal of the thermometer from the spring. Measurements were taken until temperature readings were reproducible.

Chemical Sample Collection

Sample Collection Methods

Ten water samples were collected for chemical analysis: four samples from wells and six samples from springs (Figure 8). Collection of samples followed methods outlined by the United States Geological Survey (United States Department of the Interior, 1982).

Wells were pumped at least 20 minutes at full capacity prior to collection of the samples. Samples were collected directly from a spigot which was closest to the well and not connected to a water softening system.

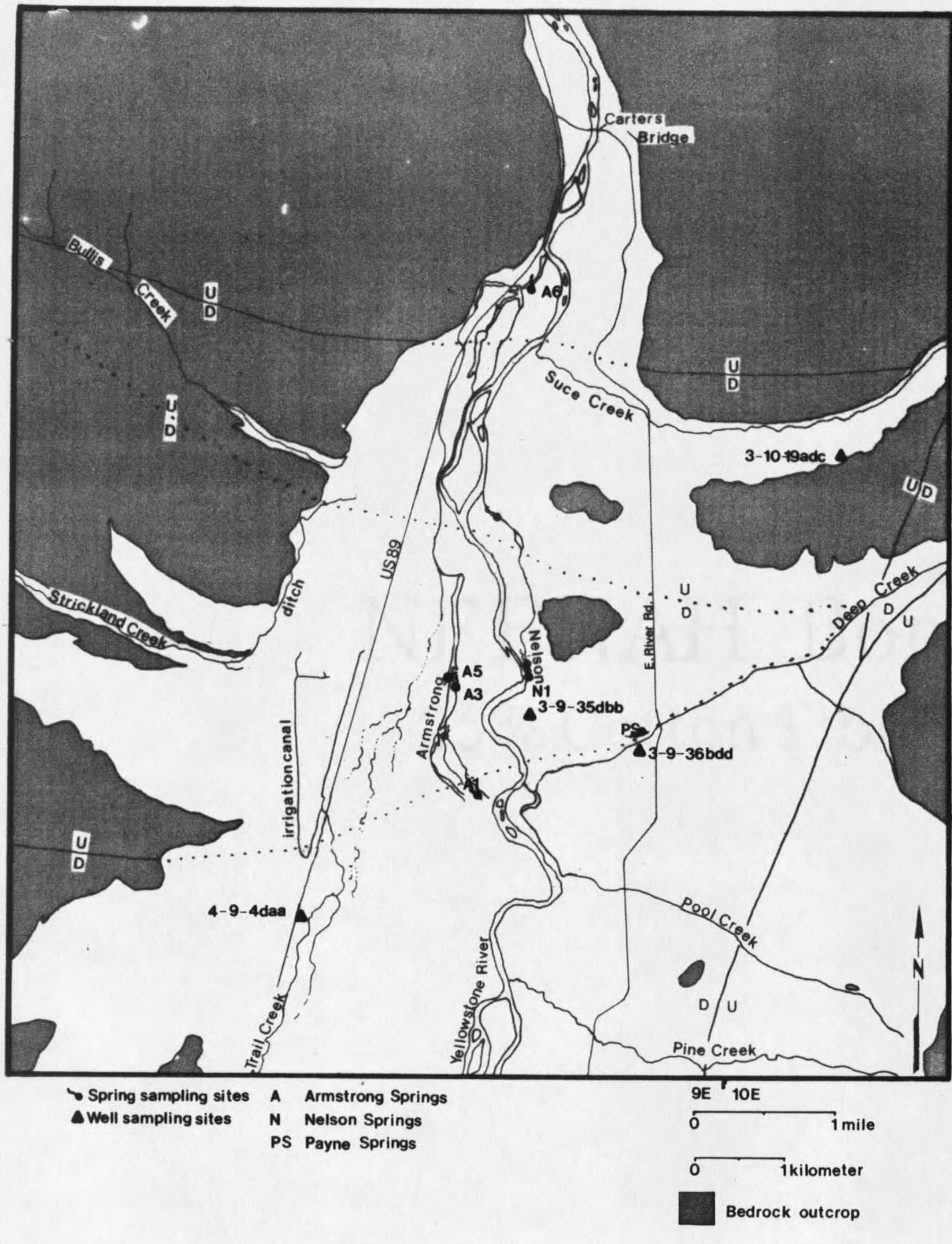


Figure 8. Well and spring chemical sampling sites.

Springs were sampled from depths of at least 0.45 meters beneath the surface following the methods of Lee and Cherry (1978). At the springs, groundwater is rising to the ground surface which suggests no surface-water contamination is likely. A mini-piezometer connected to a Master-Flex portable sampling pump was used to collect the samples. A 0.33 cm I.D. plastic line, perforated at the end and screened with fine mesh fiberglass window screen, was used for the piezometer. The springs were pumped approximately 15 minutes prior to sampling in order to ensure that sediment free water was collected.

Sample Handling Methods

Methods of handling the samples were prescribed by the MBMG analytical laboratory. Three samples were collected at each well and spring site. One (500 ml) sample was unfiltered and unpreserved, the second (500 ml) sample was filtered (0.45 um) and unpreserved, and the third (250 ml) sample was filtered and preserved with an ampule of 1% HNO_3 provided by the MBMG laboratory. In all cases the new sterile sample bottles were rinsed at least three times with distilled water and then rinsed at least three times with sample water prior to collecting the final sample. Sample bottles were filled to the top and all air was extruded prior to sealing the bottle. Replicate samples were not collected due to limited funds for chemical analyses. Samples were placed immediately on ice upon collection and cooled to approximately 4°C. Samples were hand delivered to the MBMG laboratory where the samples were kept refrigerated until analyzed.

Field Measurements

Bicarbonate content, pH, specific conductance, and temperature were determined in the field on unfiltered waters. Bicarbonate content was determined immediately upon sampling through potentiometric titration of unfiltered, unpreserved samples with 0.0169 N H_2SO_4 to a pH of 4.5 (United States Department of the Interior, 1982). A Sargent - Welch Model RB temperature compensating portable pH meter was used to measure the pH. The pH probe was rinsed a minimum of three times with the sample before taking a reading. Buffers of pH 5 and 8 were used to calibrate the pH meter at the appropriate temperature.

Specific conductance is an indirect measure of total dissolved solids (TDS) in the water (Freeze and Cherry, 1979). Specific conductance was measured using temperature compensating Model MC-1 Mark IV Lab-Line Lectro Mho and Cole Parmer Model 1040 electrical conductivity meters. The probes to the conductivity meters were rinsed a minimum of three times with the sample water prior to taking a reading. Specific conductance readings were taken until values could be replicated. The values for specific conductance were compensated to 25°C and are reported in micromhos per centimeter (uMhos/cm).

The calibration and performance of all equipment was checked against standards prior to field use following the methods outlined by the United States Geological Survey (United States Department of the Interior, 1982).

Aquifer Test Analysis

Time-drawdown semi-log plots were used to estimate the transmissivity of the aquifer. The equation used for calculation of transmissivity (T) is T = 264 Q/^s where Q is discharge (gpm) and ^s (ft) is the drawdown over one log cycle of time (minutes) (Driscoll, 1986). Values were subsequently converted to metric terms for transmissivity (m^2/s). The equation is a modification of Theis nonequilibrium equation (T = 114.6 Q W(u)/^s) developed by Cooper and Jacob (1946) (Freeze and Cherry, 1979; Driscoll, 1986). The well function term W(u) describes the head distribution about a pumped well.

The assumptions of the Cooper-Jacob (1946) test method are that u is small, the aquifer is confined, the well is fully penetrating and 100% efficient, the aquifer is isotropic, homogeneous, of infinite extent and not recharged, all groundwater is derived instantaneously from aquifer storage, and the water table has no slope (Driscoll, 1986). The W(u) term can be truncated if the distance from the pumped well to where the drawdown is measured is small (Freeze and Cherry, 1979; Driscoll, 1986). The use of single well tests indicates truncation of the well function term is appropriate and therefore the first assumption that u is small is satisfied.

The aquifers are believed to be unconfined. The Cooper-Jacob test method was designed for confined systems, but the method can be used for unconfined systems provided vertical gradients are insignificant, and drawdown is small compared to the saturated thickness of the aquifer (Freeze and Cherry, 1979). The thickness of aquifer, which was taken to be the thickness of saturated material the wells penetrate, ranged from

3.5 m to 29.7 m. The thicknesses probably are minimum aquifer thicknesses. The slope of the water table varied between 0.3 and 10%. With respect to the thickness of the aquifer, the slope of the water table is not considered significant. However, the amount of drawdown in wells probably cannot be considered small for all the tested wells. Drawdowns in the tested wells were between 17 and 85% of the saturated aquifer thickness the wells penetrate. Thus, some corrections must be made for partial dewatering.

Dewatering must be taken into account when the drawdown exceeds 20% of the saturated aquifer thickness. Therefore, the equation $S_t = S_a - (S_a^2/2b)$ where S_a = actual measured (m), S_t = corrected drawdown (m), and b = saturated aquifer thickness (m) (Driscoll, 1986) was used to correct the test data from wells 4-10-6bbd and 3-9-23dcc for the effects of partial dewatering. The tests, following corrections for partial dewatering, are considered to reasonably satisfy the assumptions of insignificant vertical gradients and a small drawdown to saturated thickness ratio.

The degree to which the third assumption of full penetration is satisfied cannot be determined from the available well reports. The wells probably do not fully penetrate the aquifer and therefore low estimates of transmissivity may be reported (Chow, 1964).

The assumptions that the aquifer is isotropic, homogeneous, of infinite aerial extent, and not recharged while testing are probably valid. The assumption that the well is 100% efficient is offset by the analysis of recovery data. Flow to the well during a recovery test should be laminar and not turbulent as may be the case during pumping. As turbulence increases, well efficiency decreases; therefore, the well can be considered to be 100% efficient during a recovery test. The assumption that groundwater is instantaneously released from storage is violated. Casing storage was a major problem due to the short test period and small discharge. Tests in wells 3-10-31aba, 3-9-32dac, 4-9-1baa, and 3-10-31cba were invalidated because the calculated period at which casing storage effects are no longer a factor in the tested wells exceeded the test period. Tests in wells 4-9-1bad and 3-9-23dcc, however, were corrected for casing storage effects using the equation:

$$t_c = 0.1\ (d_c^2 - d_p^2)/(Q/s) \text{ where}$$

t_c = time in minutes when casing storage effect becomes negligible,

d_c = inside diameter of well casing (mm),

d_p = outside diameter of pump column (mm), and

Q/s = specific capacity of the well in m^2/day of drawdown at time t_c (Driscoll, 1986). With all assumptions either valid or corrected for, the Cooper-Jacob test (1946) is considered a suitable method of analysis of the test data collected during this study to calculate aquifer transmissivities and thus aquifer hydraulic conductivities. The hydraulic conductivity (K) of the aquifers was estimated by dividing estimated transmissivity (T) by the saturated thickness of the aquifer penetrated by the well (b).

Aquifer transmissivity and hydraulic conductivity were also estimated from specific capacity tests conducted in the field and from specific capacity tests reported in well reports. Specific capacity of a well is the rate of pumping (m^3/s) divided by the measured drawdown (m) at that pumping rate. To estimate transmissivity from measured and reported specific-capacity tests, the specific capacities of the Yellowstone and alluvial fan aquifers were multiplied by 1,500 (Driscoll, 1986). The empirical value of 1,500 was used because it is believed unconfined conditions exist in the Yellowstone and alluvial fan aquifers.

RESULTS

Aquifer Characteristics

For the purpose of this study it is important to understand the water-bearing properties of each of the three identified aquifers within the study area. The information which will be presented in this section was obtained through the review of the lithologic logs contained within the drillers' well reports for the inventoried wells and from conversations with many of the well owners concerning the reliability of their particular well as a groundwater source. Other information which pertains to tested hydrogeologic characteristics of the various aquifers will be presented in a later section. The lithologic reviews and conversations with the various well owners were conducted with the intent to identify particular locales within the study area which do not readily yield groundwater and therefore may be less likely to influence the springs.

Madison Group Aquifer

Within the study area only one well, 3-10-19adc is completed within the Madison Group aquifer. This well is located within the Suce Creek drainage near the northeast corner of the field area (Figure 7). The well, the deepest in the study area, is completed to a depth of 116 m and penetrated hard fractured limestone from 104 m to total depth (Appendix

B). From 103 m to total depth, cavernous limestone was penetrated. The well is screened across the cavernous zone.

The well owner stated the well has delivered an extremely reliable supply of groundwater. The well reportedly had not gone dry since the well was completed in 1985. In fact, the well is also used to at least partially supply the adjacent owners of well 3-10-19aca with water when well 3-10-19aca goes dry during the summer months.

Alluvial Fan Aquifer

Forty-six wells which penetrate the alluvial fan aquifer were inventoried; however, only 25 of the 46 wells have drillers' reports with lithologic descriptions (Appendix B). The lithologic descriptions are general, but nevertheless, the logs provide some valuable information concerning the alluvial fan aquifer. From the well logs, the wells on average penetrate 93.5 % gravels and sands and 6.5 % silts and clays with a standard deviation of 9.1 %.

From conversations with well owners, the aquifer generally supplies a reliable source of groundwater except at well 3-10-31dbb which is between the Deep Creek and Pool Creek drainages and wells 3-10-19bdb and 3-10-19aca which are within the Suce Creek drainage (Figure 7). Well 3-10-31dbb is completed to a subsurface depth of 30 m and penetrates sand, gravel, and rock. No fine grained materials are mentioned in the well log (Appendix B). As was witnessed in the field, the well was pumped dry during an aquifer test after approximately 13 minutes of continuous discharge at 6.4×10^{-4} m^3/s. Owners of other wells in the immediate vicinity of well 3-10-31dbb report no yield problems.

In the Suce Creek drainage, well 3-10-19aca is completed to a subsurface depth of 92 m and penetrates approximately 56 % gravels and sands and 44 % silts and clays. The well fails to yield enough water for domestic purposes approximately one-half of the year. Well 3-10-19bdb also fails to adequately provide a domestic source of groundwater. Well 3-10-19bdb is completed to 24 m and penetrates sand, gravel, and clay, but the percentage of each cannot be determined due to the very general nature of the well log (Appendix B).

Yellowstone Aquifer

Eleven of the 16 inventoried wells which penetrate the Yellowstone aquifer have well logs with lithologic descriptions. The wells on average penetrate 95.4 % gravel and sand and 4.6 % silt and clay with a standard deviation of 8 %. According to the local well owners, all of the wells produce a reliable source of domestic water.

Laramide Structures

During the course of this study, indirect evidence was compiled which suggests a third reverse fault may exist between the Suce Creek fault and the Center Hill fault (Plate 1). For the purpose of this study, the fault will be referred to as the Spring Creek Hills fault. Cross-sections by Skeels (1939), Robbins and Erslev (1986), and Miller (1987) show a fault at the same approximate position, but give no evidence for its existence. Their positioning of the fault was constrained only by cross-section interpretation. Several lines of indirect evidence have been found during this study which support the existence of the fault and constrain the position of the fault.

First, two small outcrops of repeated Mississippian and Pennsylvanian section protrude through the alluvial fan aquifer immediately to the east of the Yellowstone River in the vicinity of Nelson Spring Creek (Plate 1). The two outcrops have been assumed to be slide blocks by a previous worker (Personius, 1982), but the two blocks are coherent. Dips and strikes collected from the two outcrops during this study are similar in attitude to more extensive outcrops to the east (Plate 1). Coherency and similarity in attitude suggests the outcrops are not slide blocks. Neither Roberts (1964a) nor Richards (1957) recognized the presence of the Amsden Formation and possibly the Quadrant Formation in the southern outcrop, but rather mapped the outcrop as Madison Group. The mapped relationships require Mississippian Madison Group in the northern block to be atop Pennsylvanian Quadrant Formation in the southern block (Plate 1). This relationship requires the presence of a reverse fault between the two outcrops.

Second, at Spring Creek Hills (Plate 1), a subdivision west of the Yellowstone River and south of the Suce Creek thrust fault trace, Cretaceous shales are folded and their outcrop rapidly thins to the west. The flexure in the Cretaceous shales may overlay a blind thrust fault in which displacement transfer along the fault is taken up by folding of the overlaying strata.

Third, the geologic map of Reid and others (1975, Plate 1) depicts two parallel Precambrian faults (Davis Creek faults) at the head of Deep Creek. Common association of Proterozoic faults and Laramide foreland structures suggests the Spring Creek Hills fault may be a reactivation of the preexisting Davis Creek faults. Although Spring Creek Hills fault is nowhere present at the surface, the inferred trace of the fault is constrained by outcrop positions in the valley floor, position of the flexure in the Cretaceous units, and the presence of the Proterozoic Davis Creek fault zone. Based upon the dips of the outcrops in the center of the valley, this fault probably dips to the northeast parallel to the other Laramide faults (Plate 2).

Armstrong and Nelson Spring Creek Description

Armstrong Spring Creek parallels the Yellowstone River on the west for approximately 7.24 water km prior to entering the Yellowstone River (Plate 3). Nelson Spring Creek is east of the Yellowstone River (Plate 3) and flows for approximately 2.4 km before entering the river. The springs originate relatively close to the river. Armstrong Springs are between 30 and 300 m from the river. Nelson Springs vary from 15 to 335 m from the river. The elevations of the springs vary between approximately 1.5 and 4.5 m above the Yellowstone River.

The spring creeks have been altered from their "natural" conditions by the local landowners. Nelson Spring Creek has one artificial dam and two spillways which check stream flow to create ponds that supply water for flood irrigation (Plate 3). Diversion control gates are used to regulate flow and water levels. Stream flow diverted for irrigation is a net water loss to Nelson Spring Creek because the diverted water does not reenter the stream directly. Flow in Nelson Spring Creek is also diverted to trout hatchery channels, but this is not a net loss to the creek because the diverted water reenters the stream below the hatcheries.

Armstrong Spring Creek is much less artificially altered. However, flow in Armstrong Spring Creek is also partially diverted to flow through trout hatcheries. The diversion reenters the spring creek below the hatchery and therefore probably does not represent a net loss to stream flow. Artificial riffles have been emplaced for trout habitat management purposes and one dam has been built to reroute the stream channel (Plate 3). The flow within Armstrong Spring Creek is not regulated. Therefore, flow within the creek can be assumed to reflect natural spring discharge. Because Armstrong Spring Creek is less modified by engineered structures than Nelson Spring Creek and can be assumed to reflect natural spring discharge, stream-discharge monitoring was primarily focused on Armstrong Spring Creek.

Spring Systems

Armstrong and Nelson Spring Creeks are perennial, and receive water from three types of springs. Both systems are composed of: (1) many small springs located in the floor of the streams (channel springs), (2) a conspicuous spring located topographically above the stream (A5 and N1) and (3) marshes or high water-table areas which border the spring creeks (Plate 3).

The channel springs are most easily recognized where sediment is lifted as groundwater discharges from the alluvium in the floor of the clear streams. The lack of cohesion in the sandy sediment causes continual collapse of the orifice so that lifted sediment falls back to the stream floor which imparts a "boiling" appearance to the stream bed. Individual channel springs occur in groups which are mapped as spring areas. Very rarely are there less than three channel-spring orifices in a single area.

Channel springs which issue from fine sediment at the head of Armstrong Spring Creek are less noticeable (Plate 3, A1). The cohesiveness of the fine sediment allows the orifices to remain open. In this area, the "boiling" appearance is not present and the orifices are not clustered. The largest recognized orifice was 20.5 cm in diameter; however, many smaller springs probably exist that were not observed. Discussions with landowners indicate some channel springs are submerged beneath ponds.

The springs topographically above the creeks (A5 and N1, Plate 3) are much more obvious. The topographic relief creates waterfalls as the water cascades to the streams. Armstrong Spring A5 (Plate 3) discharges from a depression in the top of the terrace bordering the streams approximately three meters above the stream floor. The spring is positioned approximately midway along Armstrong Spring Creek. Nelson Spring N1 discharges at the head of Nelson Spring Creek beneath a ravine incising the alluvial fan aquifer (Plate 3). Many channel springs are present in the floor of the stream directly beneath spring N1.

Seepage areas and areas of standing water or marshes are prevalent throughout the spring creek areas (Plate 3). The greatest concentration of such areas is in the immediate vicinity of the spring creeks (Plate 3). A seepage area is present to the immediate southwest of spring A5 (Plate 3). This high water-table area is atop the terrace from which spring A5 discharges. The seepage areas remained wet irrespective of season during the study, increasing in area through summer and declining in area through the fall. Fluctuations in areas of seepage were not measured.

Spring Discharge

The discharges of Armstrong and Nelson Spring Creeks were measured at several stations at various times throughout the study period in an attempt to determine the discharge of individual springs by calculating the difference in discharge measured upstream and downstream of individual springs (Table 3). The method proved to be ineffective for springs other

Table 3. Spring creek discharges (m^3/s). Asterisk (*) denotes estimation from rating curve. See Plate 3 for station locations. The discharge for spring A5 is calculated SA7-(SA5+SA6).

DATE	SA7	A5	SA5	SA6	SA4	SA3	SA2	SA1	SN2	SN1
3-14-87	2.12	1.67	0.31	0.14	----	----	----	----	----	----
4-4-87	2.24	----	0.31	----	----	0.20	----	----	1.25	----
4-17-87	2.38	1.93	0.31	0.17	0.31	0.45	----	----	1.25	----
5-1-87	2.44	1.84	0.40	0.23	0.34	----	----	----	1.10	----
5-8-87	2.66	1.98	0.40	0.28	0.37	----	----	----	----	----
5-22-87	2.86	2.07	0.57	0.23	0.45	0.42	----	----	0.96	----
6-30-87	3.77	2.86	0.74	0.17	0.68	0.62	----	----	0.93	----
7-2-87	3.77	2.97	0.62	0.17	0.62	0.71	----	----	0.93	----
7-6-87	3.77	2.89	0.74	0.14	0.65	0.68	0.62	----	----	----
7-8-87	3.03	2.21	0.65	0.17	0.65	0.62	----	----	----	----
7-14-87	3.03	2.18	0.71	0.14	0.71	0.65	0.59	----	----	----
8-4-87	3.12	2.07	0.91	0.14	*0.90	0.93	0.59	0.23	----	----
8-24-87	----	----	0.68	0.06	*1.10	0.71	0.68	----	----	----
9-23-87	2.61	1.50	0.96	0.17	1.10	1.22	----	0.14	----	----
10-31-87	----	----	0.88	0.17	1.04	1.19	----	0.11	----	0.25
11-30-87	----	----	0.76	0.11	0.79	0.82	----	0.11	----	0.23
2-29-88	2.58	2.12	0.34	0.08	0.31	0.31	0.23	0.03	1.70	0.28

than Armstrong Spring A5 and Nelson Spring N1 (NS1, Table 3). The discharge of other springs was less than the sensitivity of the discharge meter and difference method utilized to determine spring discharge.

During the study, discharges of Armstrong Spring Creek measured at station SA7 (Plate 3) varied from 2.12 m^3/s to 3.77 m^3/s. The average measured discharge of Armstrong Spring Creek was 2.88 m^3/s (Table 3). Measured discharges of Spring A5 (SA7 - (SA6 + SA5)) (Table 3) ranged from 1.50 to 2.97 m^3/s and averaged 2.18 m^3/s. Discharge from spring A5 contributed an average of 76 % of the total discharge of Armstrong Spring Creek measured at station SA7 near the bridge. The measured total discharge of Armstrong Channel Springs (SA6 + SA5) (Table 3) ranged from 0.42 to 1.13 m^3/s and averaged 0.78 m^3/s. Measured Nelson Spring Creek discharges varied from 0.93 to 1.25 m^3/s (Table 3), with only 17 % contributed by N1.

The number of individual channel springs and the size of the discharge area of the channel springs fluctuated throughout the study period. The size of spring area A3 (Plate 3) increased from approximately 0.8 m to 1.4 m by June 30, 1987 as the number of individual springs increased. A new spring area approximately 9 m downstream from spring A3 was observed June 30. The exact date of the first appearance of the spring is not known, but it was between May 22, 1987 and June 30, 1987. The size of spring area A3 and the new spring (A4) decreased in area from July 2, 1987 to sometime prior to August 24, 1987. The new spring area was not present August 24, 1987 and had not reappeared on February 29, 1988. A similar phenomenon was observed in Nelson Spring Creek. Channel spring area N2 (Plate 3) was present from June through late September 1987. The spring was not present when an attempt was made to collect a chemical sample from it on October 23, 1987. The spring was also not present February 29, 1988.

Spring Temperatures

Temperatures of the springs were measured during the study in order to determine the temperature range, temperature consistency, and seasonal variability of individual spring areas (Table 4). Temperatures of spring A5 ranged between 11.5°C and 12.0°C (Table 4). Channel spring A3 exhibited temperatures from 8.0°C to 11.0°C. Nelson springs varied from 9.5°C to 12.0°C.

Table 4. Spring temperatures (°C).

DATE	A5	A3	A1	N1	N2
4-4-87	12.0	8.0	----	----	----
4-17-87	12.0	8.0	----	----	----
5-1-87	12.0	9.0	----	----	----
5-8-87	12.0	10.0	----	----	---
5-22-87	11.5	10.0	10.0	----	----
6-30-87	12.0	9.0	----	----	12.0
7-2-87	12.0	9.5	----	----	----
7-6-87	11.5	10.0	----	----	----
7-8-87	12.0	9.0	----	----	----
7-14-87	12.0	9.0	----	----	----
8-4-87	11.8	9.5	----	----	----
8-24-87	12.0	10.5	----	----	----
9-23-87	12.0	11.0	10.0	10.0	12.0
10-31-87	11.8	10.5	8.9	10.0	----
11-30-87	11.5	9.8	8.5	9.5	----
2-29-87	11.8	8.0	8.5	9.8	----

Local Groundwater

Static groundwater levels were measured in 67 domestic wells (Figure 7) (Table 5) between September 29, 1987 and November 13, 1987 in order to define the water-table configuration and thus, groundwater flow patterns. Depths to groundwater in the study area varied from 0 m at springs to 67 m in well 3-10-19aca. Depths to groundwater in the alluvial-fan aquifer ranged from 0 m at well 3-9-35baa to 46 m at well 4-10-5bbc (Table 5). In the Yellowstone aquifer, groundwater occurs at depths which range from 0 m at well 3-9-23dcc to 19 m at well 4-9-4dac (Table 5) (Figure 7). Groundwater in the Madison Group aquifer occurred at 56 m below ground surface in well 3-10-19adc.

Table 5. Static groundwater levels measured in wells. Groundwater levels are corrected to ground surface as reported in Appendix A. See Figure 7 for well locations.

Date	Well	Static Groundwater Level (m)	Static Groundwater Elevation (m)	Aquifer
11-13-87	3-9-14dbb	8	1385	Yellowstone
11-13-87	3-9-14ddb	11	1394	Yellowstone
10-12-87	3-9-22cbd	20	1405	Yellowstone
10-29-87	3-9-23dcc	2	1401	Yellowstone
10-29-87	3-9-23dcd	9	1404	Yellowstone
9-29-87	3-9-27bbb	10	1410	Yellowstone
9-28-87	3-9-28baa	53	1481	Mowry/Thermopolis
9-29-87	3-9-28bba	10	1525	Mowry/Thermopolis
9-29-87	3-9-28bbb	11	1524	Mowry/Thermopolis
11-10-87	3-9-32daa	19	1455	Amsden
10-05-87	3-9-33bdd	8	1421	Yellowstone
10-05-87	3-9-33dcb	13	1420	Yellowstone
10-10-87	3-9-35acc	5	1415	Alluvial fan

Table 5. Static groundwater levels measured in wells. Groundwater levels are corrected to ground surface as reported in Appendix A. See Figure 7 for well locations-Continued.

Date	Well	Static Groundwater Level (m)	Elevation (m)	Aquifer
10/12/87	3-9-35baa	0	1414	Alluvial fan
10/28/87	3-9-35dbb	13	1410	Alluvial fan
9/29/87	3-9-35dda	20	1428	Alluvial fan
9/28/87	3-9-36adc	22	1480	Alluvial fan
10/03/87	3-9-36bdd	23	1447	Alluvial fan
9/30/87	3-9-36cab	16	1450	Alluvial fan
9/29/87	3-9-36cbc	20	1432	Alluvial fan
9/29/87	3-9-36ccb	18	1433	Alluvial fan
10/01/87	3-9-36ccc	19	1431	Alluvial fan
10/01/87	3-9-36cdb	20	1456	Alluvial fan
9/29/87	3-9-36dbb	9	1472	Alluvial fan
9/30/87	3-9-36dbd	17	1476	Alluvial fan
9/30/87	3-9-36dca	18	1479	Alluvial fan
9/28/87	3-9-36dcb	15	1468	Alluvial fan
10/05/87	3-10-19abc	52	1462	Alluvial fan
10/05/87	3-10-19aca	67	1454	Alluvial fan
10/12/87	3-10-19adc	56	1502	Madison Group
10/05/87	3-10-19bac	25	1476	Alluvial fan
10/05/87	3-10-19bdb	23	1461	Alluvial fan
9/28/87	3-10-30ccd	14	1504	Alluvial fan
10/01/87	3-10-31aba	20	1551	Alluvial fan
10/01/87	3-10-31abd	9	1564	Alluvial fan
10/12/87	3-10-31ada	12	1599	Alluvial fan
9/28/87	3-10-31bba	13	1501	Alluvial fan
9/28/87	3-10-31bbd	23	1503	Alluvial fan
10/01/87	3-10-31cab	20	1534	Alluvial fan

Table 5. Static groundwater levels measured in wells. Groundwater levels are corrected to ground surface as reported in Appendix A. See Figure 7 for well locations-Continued.

Date	Well	Static Groundwater Level (m)	Elevation (m)	Aquifer
10/12/87	3-10-31cba	20	1529	Alluvial fan
9/30/87	3-10-31cbb	15	1519	Alluvial fan
9/28/87	3-10-31cbc	15	1515	Alluvial fan
10/03/87	3-10-31cca	17	1526	Alluvial fan
9/28/87	3-10-31cdb	20	1538	Alluvial fan
10/12/87	3-10-31dbb	23	1563	Alluvial fan
10/05/87	3-10-31dcd	13	1597	Alluvial fan
9/30/87	4-9-1baa	13	1454	Alluvial fan
10/03/87	4-9-1bad	12	1460	Alluvial fan
9/30/87	4-9-1cad	21	1459	Alluvial fan
9/30/87	4-9-1dad	13	1493	Alluvial fan
10/08/87	4-9-2cdd	5	1423	Alluvial fan
10/13/87	4-9-4daa	14	1420	Yellowstone
10/05/87	4-9-4dda	15	1421	Yellowstone
10/05/87	4-9-9dac	19	1422	Yellowstone
9/29/87	4-9-10caa	13	1424	Yellowstone
9/29/87	4-9-11aad	13	1454	Alluvial fan
10/08/87	4-9-11abb	7	1432	Alluvial fan
9/29/87	4-9-11bdb	4	1420	Yellowstone
9/29/87	4-9-12bbd	18	1455	Alluvial fan
10/03/87	4-10-5bbc	46	1638	Alluvial fan
10/05/87	4-10-6bbd	11	1529	Alluvial fan

Transmissivity and Hydraulic Conductivity

Nineteen single well aquifer tests were conducted during the study in order to estimate alluvial fan and Yellowstone aquifer hydraulic conductivities and transmissivities (Figure 8) (Table 6). An aquifer test could not be conducted for Madison Group aquifer well 3-10-19adc because drawdown quickly fell beyond the length of the Fisher M-Scope electrical tape used to measure drawdown during the tests. Where aquifer tests were not conducted or where the tests yielded inconclusive results, hydraulic conductivity and transmissivities were also estimated from specific capacity tests conducted during this study and from any tests reported in drillers' well reports (Appendix C).

Of the 35 wells used to estimate aquifer hydraulic conductivity and transmissivity in this study, 26 tap the alluvial fan aquifer, 8 tap the Yellowstone aquifer, and 1 taps the Madison Group aquifer (Table 6). The determination of which aquifers the wells tap is based upon geographic position with respect to geologic outcrops, inferred subcrops, and drillers' well reports. Drillers' well reports were of primary use in distinguishing wells which withdraw groundwater from a bedrock aquifer from wells which withdraw groundwater from either the alluvial fan or Yellowstone aquifers.

Specific capacities for the alluvial fan aquifer range from 4.1×10^{-5} m^2/s to 1.4×10^{-3} m^2/s. Transmissivities for the alluvial fan aquifer range from 4.6×10^{-5} m^2/s to 1.6×10^{-3} m^2/s (Table 6). The hydraulic conductivity of the alluvial fan aquifer ranges from 3.0×10^{-6} m/s to 4.1×10^{-3} m/s. Specific capacities for the Yellowstone aquifer range from 2.5×10^{-4} m^2/s to 4.1×10^{-3} m^2/s. Transmissivities for the Yellowstone aquifer range from 2.7×10^{-4} m^2/s to 4.3×10^{-3} m^2/s. The hydraulic conductivity of the Yellowstone aquifer ranges from 1.5×10^{-4} m^3/s to 1.2×10^{-2} m^3/s. The single test for the Madison Group indicates

Table 6. Aquifer parameters estimated from test information provided in drillers' well reports (Appendix C). SC = specific capacity, T = transmissivity, K = hydraulic conductivity. Asterisk (*) denotes values calculated from well tests conducted for this study.

Well	SC (m^2/s)	T (m^2/s)	K (m^3/s)	Well	SC (m^2/s)	T (m^2/s)	K (m^3/s)
Alluvial Fan Aquifer (x 10^{-4})				Madison Group (x 10^{-3})			
3-9-35acc	5.20	6.30	0.34	3-10-19adc	41.00	55.00	4.50
3-9-35baa	12.00	12.00	2.70	Yellowstone Aquifer (x 10^{-4})			
3-9-35dbb	3.30	4.00	2.60	3-9-22cbd*	2.70	2.70	7.70
3-9-35dda	1.60	1.90	5.40	3-9-23dcc*	1.60	16.00	1.50
3-9-36bda	14.00	14.00	41.00	3-9-27bbb	20.00	21.00	7.00
3-9-36bdd	10.00	13.00	0.08	3-9-33bdd	41.00	43.00	120.00
3-9-36ccc	1.00	12.00	40.00	4-9-2cdd	2.50	2.70	7.60
3-10-19abc	0.85	1.10	3.10	4-9-4daa	4.10	4.30	12.00
3-10-19aca	1.80	1.90	----	4-9-10caa	2.60	2.70	7.70
3-10-19bac	1.80	0.10	0.17	4-9-11bdb	21.00	22.00	62.00
3-10-19bdb	1.30	1.60	4.50				
3-10-31aba	0.41	0.46	0.03				
3-10-31ada	1.60	1.90	5.40				
3-10-31cab	0.75	0.91	2.60				
3-10-31cba	3.10	4.40	0.58				
3-10-31cbb	3.30	4.00	11.00				
3-10-31cdb	1.10	1.40	0.30				
3-10-31dbb	2.10	2.50	7.20				
3-10-31dcd	0.81	0.98	2.80				
4-9-1baa	1.20	1.00	0.05				
4-9-1bad*	2.50	8.00	0.30				
4-9-1cad	2.10	2.50	7.20				
4-9-11aad	0.79	0.96	2.70				
4-9-11abb	7.80	16.00	1.60				
4-9-12bbd	1.20	1.90	5.40				
4-10-6bbd*	2.50	1.40	0.06				

a specific capacity of 4.1×10^{-3} m^2/s, a transmissivity of 5.5×10^{-3} m^2/s, and a hydraulic conductivity of 4.5×10^{-4} m/s (Table 6).

Groundwater Temperatures

Groundwater temperatures were measured in 41 water wells between September 28, 1987 and November 13, 1987 (Table 7). Groundwater temperatures represent only one point in time and the period over which the point measurements were collected is large. The purpose of the temperature survey, however, was to collect temperature data in order to compare the temperature of groundwater in the three aquifers with each other and with the temperatures of the groundwater discharged at the springs. The purpose of the survey was not to determine the distribution of groundwater temperatures in the valley at an instantaneous point in time. Therefore, the period of time over which the samples were collected should not invalidate interpretations based upon this data.

Temperatures from all wells, regardless of depth of intake, were used in the study for several reasons: (1) wells in the same vicinity, regardless of depth, generally produce similar temperatures (Table 7), (2) only ten wells greater than 30 m in depth were available for survey, (3) only two of the ten wells which exceeded 30 m in depth (3-10-31aba, 3-31-31cdb) produce temperatures which are in contrast to temperatures yielded from shallow wells in the vicinity, (4) the warmest and coolest well temperatures are present in shallow wells (Table 7) which indicates surface water infiltration does not significantly affect resident groundwater temperatures, and (5) regression between well depth and temperatures for all wells and the ten deep wells presented in Table 5

Table 7. Temperatures of well waters. Depth is the intake level. Dashes indicate that the intake level is not reported in the well report.

Well	Date	Temp (°C)	Depth (m)
Alluvial Fan Aquifer			
3-9-35dbb	11-3-87	9.8	20
3-9-36adc	9-28-87	8.0	73
3-9-36bda	11-13-87	8.2	12
3-9-36bdd	10-3-87	8.5	36
3-9-36cbc	9-29-87	8.8	28
3-9-36ccc	10-3-87	9.0	37
3-9-36dbb	11-13-87	9.0	--
3-9-36dbd	9-30-87	7.6	--
3-10-19abc	10-5-87	7.9	28
3-10-19bdb	10-5-87	7.5	24
3-10-30ccd	9-30-87	7.5	--
3-10-31aba	10-1-87	9.8	38
3-10-31ada	10-12-87	5.5	18
3-10-31bbd	9-28-87	7.5	--
3-10-31cab	10-1-87	7.6	29
3-10-31cba	10-12-87	7.9	27
3-10-31cbb	9-28-87	7.9	--
3-10-31cbc	9-30-87	7.9	27
3-10-31cdb	9-28-87	9.0	43
3-10-31dbb	10-12-87	7.6	30
3-10-31dcd	10-5-87	7.8	26
4-9-1bad	10-3-87	8.9	37
4-9-11aad	9-29-87	10.0	31
4-9-11abb	10-5-87	9.8	18
4-9-12bbd	9-29-87	10.0	43
4-10-6bbd	10-5-87	7.5	33

Well	Date	Temp (°C)	Depth (m)
Madison Group			
3-10-19adc	10-12-87	8.7	116
Yellowstone Aquifer			
3-9-12cca	11-3-87	8.9	--
3-9-14dbb	11-3-87	10.9	--
3-9-14ddb	11-3-87	9.2	--
3-9-22cbd	10-12-87	11.9	23
3-9-23dcd	11-3-87	10.8	--
3-9-23dcc	11-3-87	12.0	12
3-9-27bbb	10-12-87	11.9	23
3-9-32daa	10-5-87	10.8	43
3-9-33dcb	10-5-87	12.0	--
3-9-33bdd	10-5-87	11.2	11
4-9-4daa	10-5-87	10.5	21
4-9-4dda	10-5-87	10.5	--
4-9-9dac	10-5-87	10.5	22
4-9-10caa	9-29-87	12.1	20

yielded coefficients of 0.012 and 0.036 respectively. This low correlation suggests little relationship between well depth and temperature. For the purpose of this study, all well-water temperatures will be considered to be representative of local groundwater temperatures.

Groundwater temperatures in the alluvial fan aquifer varied from 5.5°C at well 3-10-31ada to 12.5°C in well 3-9-23dcd (Table 7). Groundwater temperatures in the Yellowstone aquifer ranged from 10.5°C in wells 4-9-9daa, 4-9-4dac, and 4-9-4dda to 12°C in well 3-9-33dcb (Table 7). Madison Group groundwater was 8.7°C in well 3-10-19adc (Table 7).

Groundwater Chemistry

Ten groundwater samples were collected for chemical analysis between October 23 and October 28, 1987 (Figure 8) (Table 8). The samples were analyzed by the MBMG Analytical Laboratory.

Cation - anion balance (Freeze and Cherry, 1979, p. 97) of the samples yielded values ranging from - 0.9 % to 0.02 % error utilizing the field measured HCO_3 content. The cation - anion balance errors are significantly less than the critical 5.0 % error maximum for acceptable results (Freeze and Cherry, 1979). Therefore, the analyses are considered valid.

Field bicarbonate is used for chemical interpretation because immediate field determination is more representative of actual concentrations than later laboratory analysis (United States Department of the Interior, 1982). The mean percent error between respective field and laboratory bicarbonate concentration (meq/l) is + 7.6 %. This error is as expected because bicarbonate concentration increases up to a pH of

Table 8. Chemical analyses for groundwater samples collected from wells and springs in the study area. All results reported as milliequivalents/liter (meq/l) except as indicated. Ca = calcium, Mg = magnesium, Na = sodium, HCO_3 = bicarbonate, Cl = chloride, SO_4 = sulfate, SiO_2 = silica, Sr = strontium (ug/l), SI = calcite saturation index, TDS = total dissolved solids (mg/l), SC = specific conductance (uMhos/cm). Carbonate was not present and is therefore not reported. Qya = Yellowstone aquifer, Mmc = Madison Group, Qaf = alluvial fan aquifer. See Figure 8 for sampling locations.

Location	A1	A3	A5	N1	A6
Ca	1.36	1.79	1.70	1.63	2.20
Mg	0.68	0.80	0.87	0.77	1.25
Na	0.45	0.43	0.61	0.52	0.65
K	0.09	0.10	0.10	0.06	0.11
HCO_3 FIELD	2.19	2.56	2.91	2.82	3.94
HCO_3 LAB	2.02	2.54	2.52	2.33	3.45
Cl	0.13	0.15	0.13	0.11	0.15
SO4	0.38	0.39	0.60	0.50	0.64
SiO_2	0.37	0.46	0.55	0.42	0.46
Sr	120.0	150.0	180.0	160.0	270.0
SI_c	-0.84	-0.82	-0.46	-0.07	-0.13
pH Field	6.90	6.62	7.71	7.85	7.36
pH Lab	7.21	6.97	7.40	7.85	7.48
TDS (mg/l)	156.17	188.20	205.30	180.80	246.50
SC Field	256.0	300.0	312.0	295.0	398.0
SC Lab	267.9	317.1	336.1	308.2	424.4

Location	4-9-4daa	3-9-10adc	3-9-36bdd	3-9-35dbb	PS
Ca	1.46	1.78	1.78	1.53	1.38
Mg	0.77	1.67	0.60	0.49	0.49
Na	0.47	0.15	0.32	0.23	0.18
K	0.09	0.04	0.08	0.06	0.06
HCO_3 Field	2.50	3.35	2.76	2.23	1.85
HCO_3 Lab	2.21	3.36	2.49	2.09	1.93
Cl	0.17	0.04	0.03	0.02	0.02
SO_4	0.37	0.19	0.21	0.17	0.14
SiO_2	0.51	0.21	0.20	0.16	0.18
Sr	150.0	75.0	43.0	36.0	32.0
SI_c	-0.94	0.14	-0.40	-0.77	-1.07
pH Field	6.90	8.29	7.20	7.06	7.25
pH Lab	7.05	7.87	7.45	7.22	6.95

Table 8. Chemical analyses for groundwater samples collected from wells and springs in the study area. All results reported as milliequivalents/liter (meq/l) except as indicated. Ca = calcium, Mg = magnesium, Na = sodium, HCO_3 = bicarbonate, Cl = chloride, SO_4 = sulfate, SiO_2 = silica, Sr = strontium (ug/l), SI = calcite saturation index, TDS = total dissolved solids, (mg/l), SC = specific conductance (uMhos/cm). Carbonate was not present and is therefore not reported. Qya = Yellowstone aquifer, Mmc = Madison Group, Qaf = alluvial fan aquifer. See Figure 8 for sampling locations-Continued.

Location	4-9-4daa	3-9-10dac	3-9-36bdd	3-9-35dbb	PS
TDS	174.4	185.9	152.2	126.4	116.8
SC Field	247.0	335.0	266.0	231.0	202.0
SC Lab	288.3	352.3	286.5	242.1	221.7
Geologic Source	Qya	Mmc	Qaf	Qaf	Qaf

7.5 and then stabilizes up to a pH of 9 (Freeze and Cherry, 1979). The mean percent change between field pH and laboratory pH is - 0.5 %. The pH of the samples increased from an average field value of 7.32 to an average laboratory value of 7.35. The change is most likely due to loss of CO_2 with time which results in an increase in pH (United States Department of the Interior, 1982). The pH increase is small, but it suggests the bicarbonate content of the samples should have decreased after sample collection in the field.

The mean percent error between respective field and laboratory specific conductivity is - 7.3 %. The field measurements are probably better estimates of the total dissolved solids content of the groundwater because of precipitation and chemical reactions which can occur as the sample sits in the sample bottle (United States Department of the Interior, 1982). Field values of specific conductivity are used for discussion.

INTERPRETATIONS

Basin Hydrogeology

Before interpretations can be made which concern the hydrogeology of Armstrong and Nelson Springs, the conditions under which local groundwater occurs in the Paradise Valley must first be established. Therefore, consideration is given to the hydrogeologic setting in which the springs occur.

Aquifers and Groundwater Development

The Yellowstone and alluvial fan aquifers are the principal shallow aquifers in the study area. The two aquifers are the most extensive on the valley floor and are the groundwater source for most of the wells in the study area (Plate 1, Figure 7). With minor exceptions, the two aquifers yield a reliable source of groundwater to the domestic users.

The alluvial fan and Yellowstone aquifers are composed of a heterogeneous mixture of unconsolidated sands and gravels with minor amounts of clay in some locations (Appendix B). In short, well logs show no low-permeability layer which can act as a pronounced aquitard. On average, gravel and sand compose greater than 93 % of the aquifer material. Therefore, the permeability of the aquifer should be moderate to high throughout and should not be conducive to the development of a confined aquifer system. In summary, the lithology of the aquifers suggests the aquifers are unconfined.

It should be pointed out, however, that the alluvial fan deposits within the Suce Creek drainage contain more fine-grained deposits than those to the south of the Suce Creek drainage as is indicated in the lithologic log of well 3-10-19aca (Appendix B). Also, two of the three wells completed within the alluvial fan aquifer which produce an unreliable source of groundwater (wells 3-10-19bdb, 3-10-19aca, and 3-10-31dbb, Figure 7) are located within the Suce Creek drainage. The unreliable groundwater availability may be related to the presence of a greater amount of fine-grained deposits in the Suce Creek alluvial-fan deposits than in the Deep Creek-Pool Creek-Pine Creek alluvial fan. Suce Creek drains primarily a Paleozoic sedimentary section whereas Deep Creek, Pool Creek, and Pine Creek primarily drain a coarse-grained Precambrian metamorphic section (Roberts, 1964a; Reid and others, 1975). Thus, the drainage basin bedrock geology may account for the finer-grained nature of the Suce Creek alluvial fan. In contrast to the rest of the study area, the alluvial fan aquifer in the Suce Creek drainage probably should not be considered a promising source of domestic water.

In the Suce Creek drainage the Madison Group is probably the most promising source of groundwater. Some conditions under which groundwater occurs in the Madison Group aquifer can be established by the evaluation of well 3-10-19adc which taps the Madison Group within the Suce Creek drainage (Figure 7). Well 3-10-19adc is completed to 116 m below grade and is screened through cavernous limestone (Appendix B). The screening of the well across a cavernous zone suggests at least part of the groundwater in the Madison Group may be restricted to solution zones. The restriction of groundwater primarily to cavernous zones in the Madison

Group is noted in other areas of Montana (Groff, 1962b; Feltis, 1973; Wilke, 1983). The amount of groundwater transmission through the solution zones of the Madison Group must be adequate for domestic purposes, for by way of comparison, wells 3-10-19aca and 3-10-19bdb are screened within the alluvial fan of the Suce Creek drainage and produce very little water. In fact, the production problem with well 3-10-19aca may be remedied if the well were redrilled to greater depth in order to intercept the underlying Madison Group aquifer. The lithologic log for well 3-10-19aca describes red clay and rock from 80 to 92 m, the maximum depth of the well. The red clay may represent the lower portion of the Amsden Formation which immediately overlies the Madison Group. Groundwater development within the Suce Creek drainage appears to have the most promise within the Madison Group.

Groundwater Migration

Based upon lithology, the alluvial fan and Yellowstone aquifers are considered to be unconfined. The configuration of the water table within the aquifers suggest the same. In general, the contour of the water-table elevations in the alluvial fan and Yellowstone aquifers is a subdued expression of surface topography (Figure 9). The local topography slopes from the valley margins to the Yellowstone River and south to north in the valley floor.

Groundwater flow is perpendicular to the water-table contours in the water-table map (Figure 9). The water-table map is a composite of groundwater levels measured in the alluvial fan aquifer, Yellowstone aquifer, and the Madison Group. Thus, one component of groundwater flow is from the valley margins toward the Yellowstone River. A second

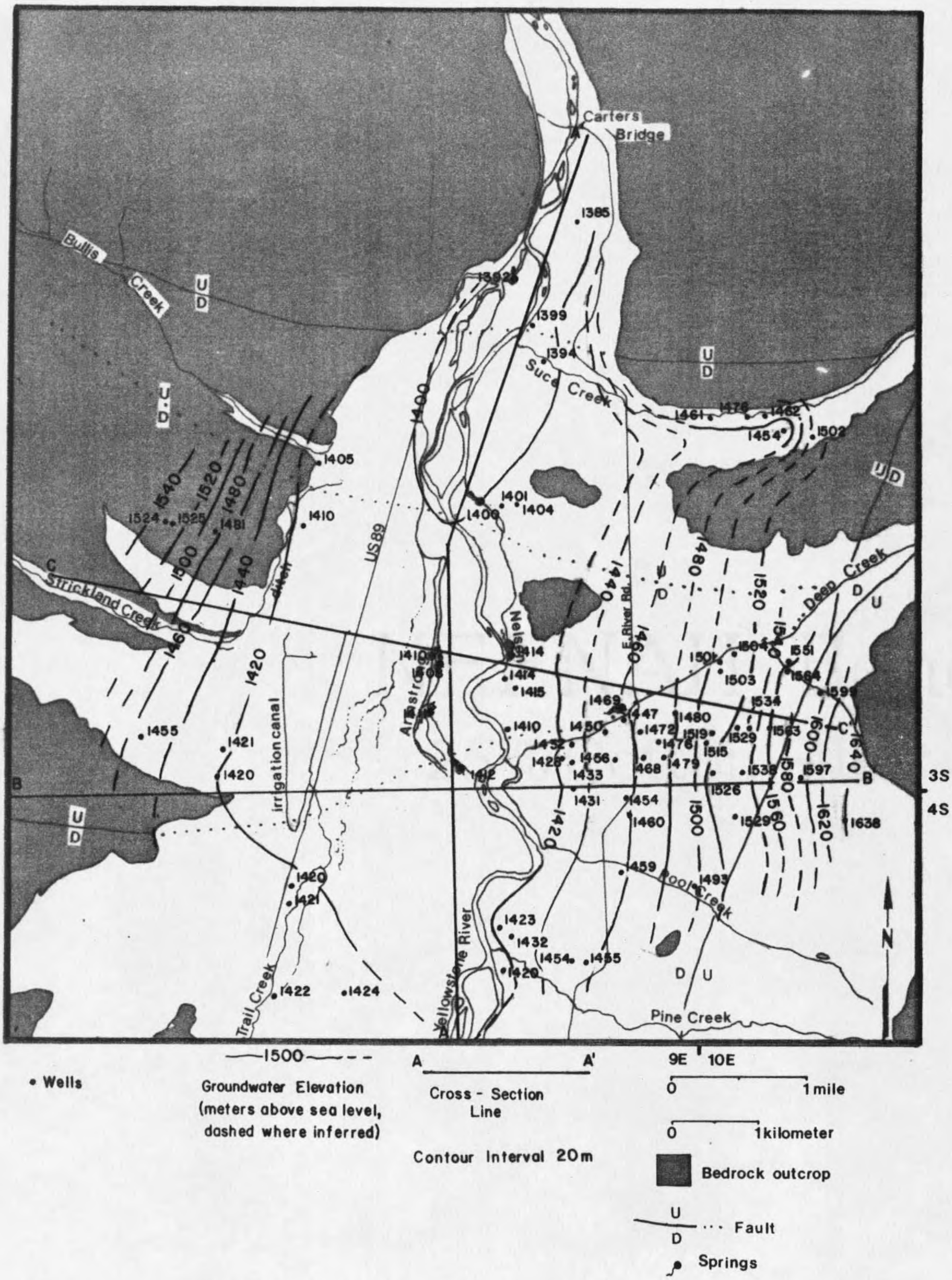


Figure 9. Groundwater gradient map. Groundwater elevations based on static water levels measured between September 29, 1987 and November 13, 1987. Elevations are measured from ground surface (m).

component of groundwater flow is northward in the valley floor where groundwater eventually exits the valley as underflow (Figures 9 and 10). The similarity between the surface topography and the configuration of the groundwater surface suggests the alluvial fan and Yellowstone aquifers are unconfined.

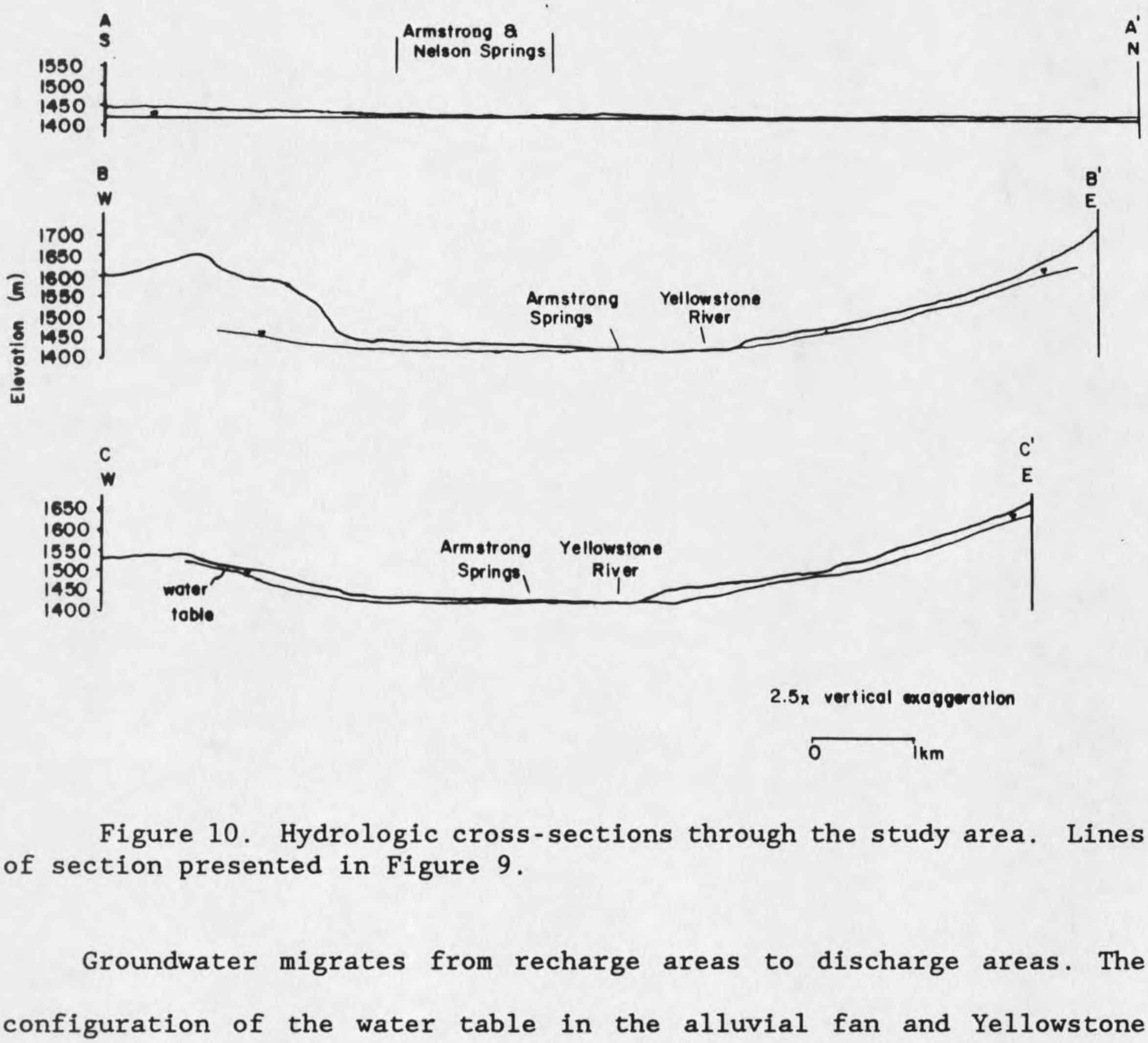


Figure 10. Hydrologic cross-sections through the study area. Lines of section presented in Figure 9.

Groundwater migrates from recharge areas to discharge areas. The configuration of the water table in the alluvial fan and Yellowstone aquifers suggests the aquifers are recharged primarily from the highland areas bordering the valley and along the trend of the Yellowstone River (Figures 9 and 10). Tertiary and Quaternary deposits may transmit groundwater into the area from the south, but this interpretation is speculative at this time due to the lack of hydrogeologic data south of the study area. However, the configuration of the water table and the occurrence of Armstrong and Nelson Springs in the valley floor clearly suggest the valley floor within the study area is at least a local groundwater discharge area for groundwater derived in surrounding highland areas and areas to the south.

Groundwater within the alluvial fan aquifer is interpreted to migrate toward the Yellowstone aquifer (Figures 9 and 10). The mean hydraulic conductivity calculated for the alluvial fan aquifer is 8.8×10^{-4} m^3/s and for the Yellowstone aquifer is 2.8×10^{-3} m^3/s and the geographic distribution of hydraulic conductivities in the two aquifers is random (Figure 11). Drillers' well reports for the wells tapping the two aquifers do not indicate any particular area of low permeability deposits (Table 6). Since the hydraulic conductivity of the Yellowstone aquifer is generally greater than that of the alluvial fan aquifer and there is no apparent zone of lower hydraulic conductivity, the Yellowstone aquifer should not impede any groundwater flow coming to it from the alluvial fan aquifer. In fact, this is supported by the flattening of hydraulic gradients from the alluvial fan aquifer into the Yellowstone aquifer (Figure 9). Therefore, the absence of a particular zone of lower hydraulic conductivity in the Yellowstone aquifer which would prevent or impede transmission of groundwater from the alluvial fan aquifer is a reasonable assumption. The Yellowstone aquifer can be interpreted to receive recharge from the alluvial fan aquifer.

The hydrogeologic setting of the study area suggests the valley floor may also be a discharge area for underlying bedrock. The presence

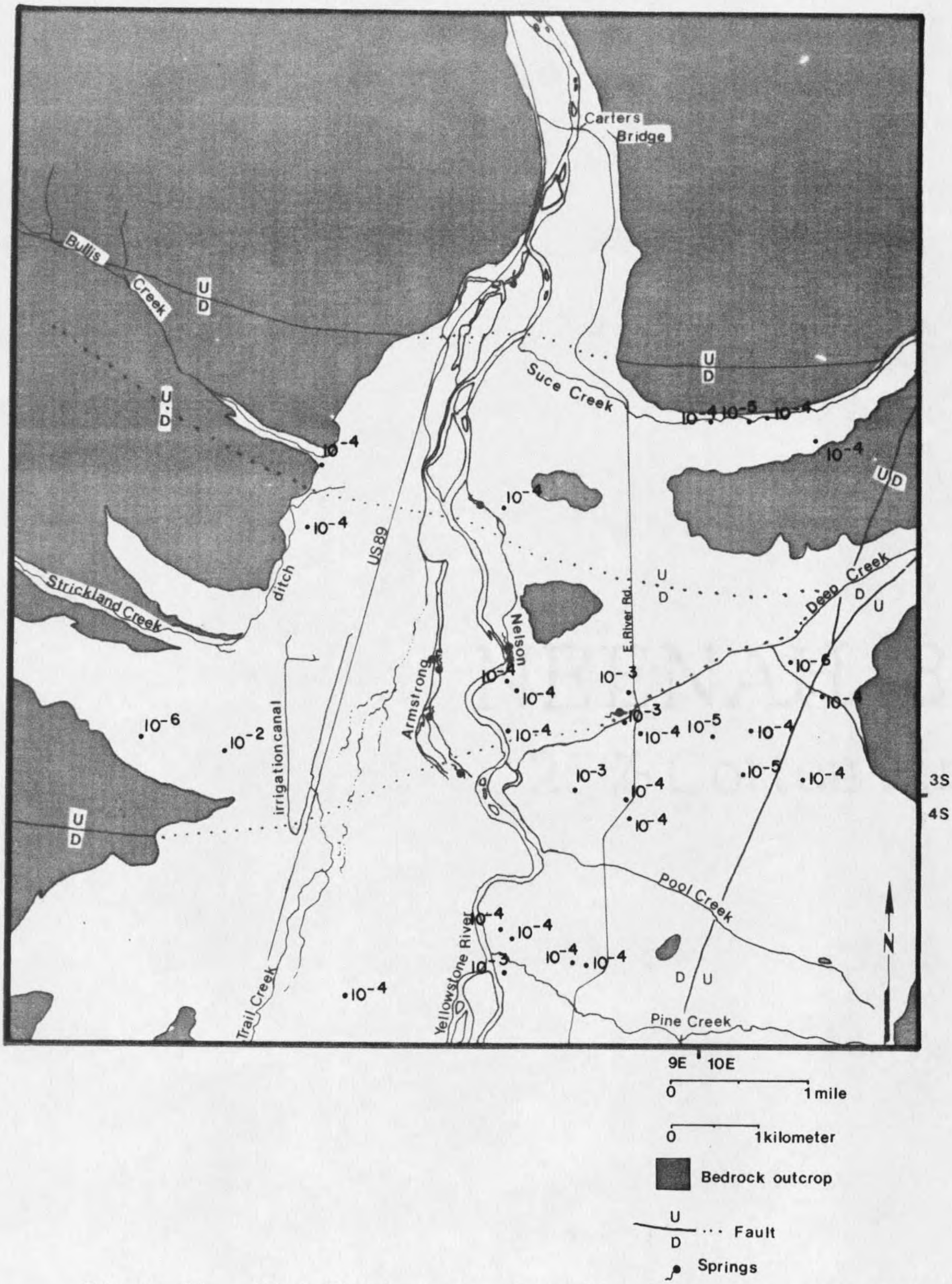


Figure 11. Distribution map of aquifer hydraulic conductivities (m^3/s). Order of magnitude estimates. Reference Table 6.

of Laramide thrust faults which cut the Madison Group suggests the Madison Group beneath the valley floor may recharge the overlaying alluvium. Faults which control groundwater migration are often identified by shifts in hydraulic gradients across the faults (Maxey, 1968; Huntoon, 1985a, 1985b).

If the Laramide faults do act as discharge pathways to the Yellowstone aquifer, a change in hydraulic gradient within the Yellowstone aquifer in the vicinity of the faults should be expected as groundwater from the underlaying Madison Group is introduced into the Yellowstone aquifer. The sparse water-table elevation data in the vicinity of the Laramide faults does not suggest the presence of any groundwater mounding (Figures 9 and 10). Recharge from the underlaying Madison Group is either not occurring, or the recharge is simply not recognized. Clearly, upwelling of groundwater from the faults may not be recognized due to the limited well control in the vicinity of the Laramide faults and the $\pm$ 20 foot (7 m) resolution of the topographic maps from which elevations are estimated. Also, the high hydraulic conductivity of the Yellowstone aquifer may readily compensate for the fault discharge and not allow for the development of a well developed groundwater mound which would be readily recognized. However, other data which will be discussed in subsequent sections also suggests the Madison Group is not recharging the overlaying Yellowstone aquifer. The Laramide structures are not considered to serve as groundwater migration zones for Madison Group groundwater; however, the structures have influenced Cenozoic depositional patterns which in turn influences groundwater migration within the Yellowstone aquifer.

Groundwater Migration--Structural and Depositional Influence

The gravity study conducted in the Paradise Valley by Bonini and others (1972) suggests the thicknesses of the Cenozoic deposits in the study area are influenced by the structural characteristics of the Paradise Valley. Based upon gravity data, Bonini and others (1972) infer that as much as 6,100 m of displacement along the Deep Creek fault has allowed the accumulation of approximately 1 km of Tertiary volcanic rock, 760 m of Tertiary sediments, and an estimated thickness of 60 m of Quaternary alluvium approximately 15 km south (upgradient) of the study area. Montana Power Company RP624-Hobbs test well located approximately 12 km to the south of the study area (Figure 1), penetrated approximately 243 m of alluvium blanketing 457 m of white tuff and 335 m of Eocene breccia (Montagne and Locke, 1989). Both the gravity data and well information indicate a large thickness of alluvium south of the study area. In contrast, Kirby (1940) determined that the depth to bedrock beneath the Yellowstone aquifer is 15.5 m at the northern terminus of Paradise Valley. Accordingly, Bonini and others (1972) indicate offset along the Deep Creek fault in the northern (downgradient) end of Paradise Valley is only 830 m to 1,800 m which no doubt accounts for the minimal accumulated thickness of Cenozoic deposits (the Yellowstone aquifer) in the northern end of the valley. Thus, the Yellowstone aquifer is a wedge of alluvium decreasing in thickness from 243 m approximately 12 km south of the study area to 15.5 m at the northern terminus of the study area (Figure 12).

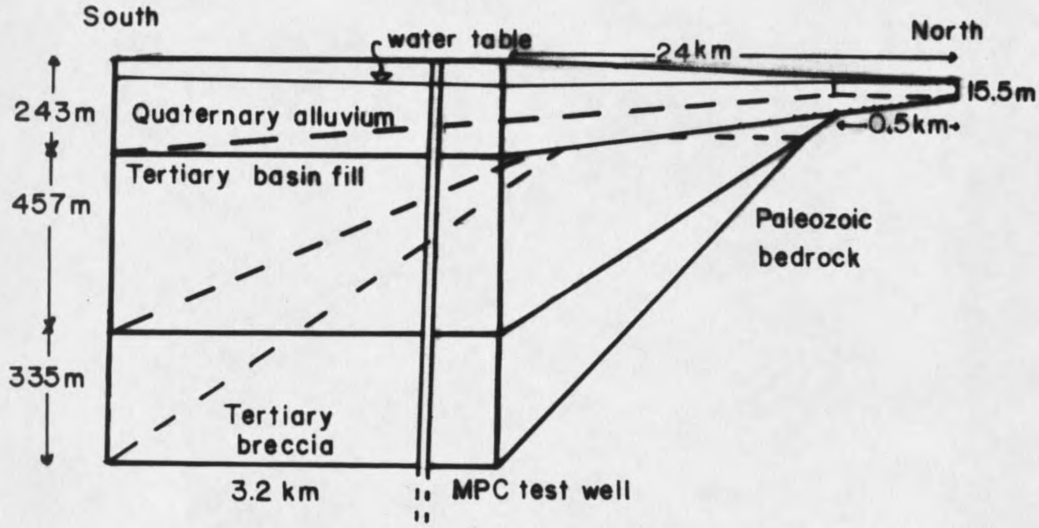


Figure 12. Diagrammatic section of the Paradise Valley extending from the MPC test well to the northern end of the study area illustrating the effect of the downgradient thinning of the Yellowstone aquifer (Quaternary alluvium) upon the hydraulic gradient.

The Laramide faults also are interpreted to be a factor in the downgradient thinning of the Yellowstone aquifer. The bedrock surface of the valley floor to the north of the Center Hill fault zone may be a topographic as well as a structural high with respect to the bedrock surface of the valley floor immediately south of the fault zone. The support for this interpretation is the presence of the Paleozoic outcrops to the immediate east of the Yellowstone River and to the north of the trace of the Center Hill fault (Plate 1). These outcrops indicate the bedrock surface is at shallow depth immediately north of the Center Hill fault. To the south of the Center Hill fault and near the trace of the Deep Creek fault, the Cambrian Flathead Quartzite outcrops (Plate 1). This also suggests that the bedrock is at a shallow depth to the south of the Center Hill fault. However, the presence of Armstrong and Nelson Springs to the north of the Center Hill fault indicates that north of the Center Hill fault the Yellowstone aquifer is thinned such that the aquifer cannot transmit the volume of groundwater present south of the Center Hill fault.

The Yellowstone aquifer is also influenced by the narrow canyon at the northern terminus of the Paradise Valley. The narrow canyon cut through Canyon Mountain Anticline by the Yellowstone River constricts the Paradise Valley from 8 km to less than 0.5 km in width at the northern terminus of the valley (Plate 1). Likewise, the width of the Yellowstone aquifer itself is narrowed from 3.2 km within the valley to 0.5 km within the canyon (Plate 1). In short, the Yellowstone aquifer is a wedge-shaped alluvial body which thins vertically and horizontally from south to north in the valley.

The hydraulically downgradient (south to north) reduction in the cross-sectional area of the Yellowstone aquifer is important to groundwater flow in the valley. According to Darcy's equation, the discharge of an aquifer is a function of the hydraulic conductivity of the aquifer, the hydraulic gradient within the aquifer, and the cross-sectional area of the aquifer (Freeze and Cherry, 1979). From south to north in Paradise Valley, the volume of groundwater in the aquifer probably increases as additional recharge is received from creeks, irrigation and precipitation. The hydraulic conductivity of the aquifer may be essentially constant. The critical factor in the Paradise Valley is the downgradient reduction in aquifer cross-sectional area of the aquifer due to aquifer thinning and constriction. A first approximation of the degree of reduction in aquifer cross-sectional area between the location of the MPC test well and the northern terminus of the study area (Figures 1 and 12) suggests the aquifer cross-sectional area is reduced by as much as 99%. The cross-sectional area of the Yellowstone aquifer as modeled (Figure 12) is reduced from 0.77 km^2 at the MPC test well to 0.008 km^2 at the northern terminus of the study area.

As a result of aquifer thinning, the hydraulic gradient within the aquifer at the northern end of the valley must increase in order to transmit the same or an increased volume of groundwater through the aquifer (Figure 12). As hydraulic gradients increase in a downgradient direction, groundwater is forced toward the valley floor at the northern end of Paradise Valley. Due to the lack of groundwater gradient data south of the study area, this interpretation is somewhat speculative. However, this interpretation is analogous to the hydrogeologic model proposed for the Big Hole basin (Levings, 1986).

The hydrogeology of the Big Hole basin (Levings, 1986) is here proposed as an analogue for the hydrogeology of Paradise Valley. Tertiary and Quaternary materials serve as aquifers in the Big Hole basin. The Big Hole basin is a Tertiary basin with as much as 4,900 m of Tertiary volcanic and basin-fill rocks and Quaternary deposits at the southern (upstream) end of the basin. In contrast, the presence of pre-Tertiary outcrops in the basin floor coupled with a constriction of basin width by 80 % suggests only a thin, narrow veneer of Tertiary rocks and Quaternary deposits are present in the northern (downgradient) end of the basin. As a result, Levings (1986) concluded groundwater which is underflow in the middle and upper reaches of the basin is forced toward the ground surface (the Big Hole River) at the downgradient end of the valley.

In summary, groundwater is interpreted to discharge in the valley floor at the northern terminus of the Paradise Valley due to a thinning of the Cenozoic aquifer system. Groundwater temperature patterns also suggest that the Cenozoic aquifer system discharges in the valley floor.

Groundwater Temperatures

Recharge waters generally change in temperature during flow through the aquifer in order to equilibrate with the temperature of the aquifer materials. The temperature of the aquifer is a function of the local geothermal gradient, mean annual temperature, thermal conductivity of the rock, and depth within the aquifer (Cartwright, 1970). Therefore, profiling groundwater temperatures is a tool to determine the direction of groundwater movement (Cartwright, 1970, 1979). In the Paradise Valley, groundwater temperature patterns (Figure 13) suggest several things about the local basin hydrogeology. First, groundwater temperature patterns are consistent with groundwater gradients and spring positions in suggesting that the valley walls are recharge areas for the alluvial fan and Yellowstone aquifers, and that the valley floor serves as a groundwater discharge area for the two aquifers. The incidence of cooler temperatures near recharge areas and warmer temperatures in the discharge area indicates groundwater gradually warms as it flows from the recharge area to the discharge area.

Second, groundwater temperatures and temperature patterns suggest the alluvial fan and Yellowstone aquifers are hydraulically continuous. Groundwater within the alluvial fan aquifer progressively warms in a downgradient direction and the warming trend continues into the Yellowstone aquifer. The groundwater temperature trend is not disrupted at the aquifer contact as may occur if the two aquifers were not hydraulically continuous. Therefore, a transition of groundwater from

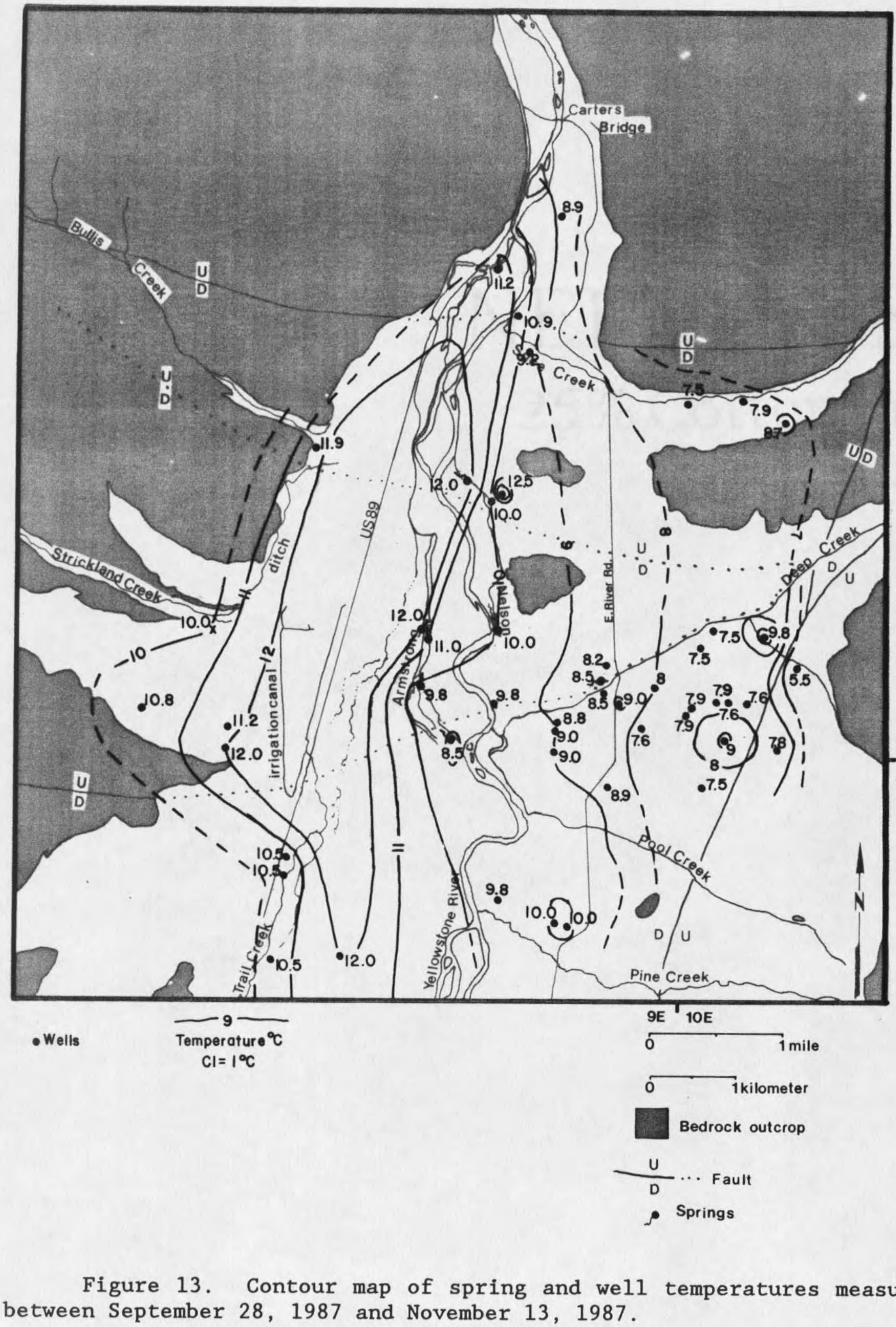


Figure 13. Contour map of spring and well temperatures measured between September 28, 1987 and November 13, 1987.

the alluvial fan aquifer into the Yellowstone aquifer occurs and thus the Yellowstone aquifer and the alluvial fan aquifer may constitute a continuous groundwater flow system.

Third, groundwater temperatures in the floor of the valley and within the two Cenozoic aquifers further suggest the valley floor represents a natural groundwater discharge area for a local or shallow groundwater flow system. The two Cenozoic aquifers are shallow surficial aquifer systems in which groundwater is present at depths ranging from 0 to 43 m and averaging 27 m. The temperature of groundwater at depths of 9 to 18 m normally exceeds the mean annual air temperature by 1°C to 1.5°C and in exceptional localities the excess may be 2.8°C to 3.3°C (Walton, 1970; Collins, 1925). Below 18 m in depth, groundwater temperatures generally increase by 0.5°C for each additional 21 m in depth (Collins, 1925). Therefore, groundwater discharging from great depths should be appreciably warmer than groundwater discharging from shallow depths. Measured groundwater temperatures in the study area, which range from 5.5°C to 12.5°C and average 9.0°C, are slightly warmer than the local mean annual temperature of 7.1°C (Ruffner, 1985), but no more than should be expected according to Walton (1970) and Collins (1925). Therefore, the shallow depths at which groundwater occurs within the Cenozoic aquifers and the close approximation of groundwater temperatures which would be predicted by Walton (1970) and Collins (1925) suggest a shallow groundwater flow system discharges in the valley floor.

Fourth, groundwater temperatures suggest a regional groundwater flow system may be developed in the glacial outwash alluvium of the Yellowstone aquifer (Plate 1). The groundwater temperature profile is asymmetrical with respect to the Yellowstone River and associated floodplain (Figure 13). The warmest groundwater is typically present to the west of the Yellowstone River and coincides almost exclusively with the outcrop pattern of the Yellowstone aquifer glacial outwash alluvium (Figure 13 and Plate 1). Temperatures of wells and Armstrong Springs located in the Yellowstone floodplain alluvium indicate groundwater is typically cooler within the Yellowstone floodplain than within the Yellowstone aquifer glacial outwash deposits. Hence, the temperature profile suggests that groundwater within the glacial outwash deposits may constitute a separate flow system from that of the Yellowstone floodplain alluvium.

The relatively warm temperature of groundwater within the glacial outwash deposits of the Yellowstone aquifer may be explained in several ways. One explanation is that the warm temperatures represent a discharge area for a regional groundwater flow system developed within the glacial outwash deposits of the Yellowstone aquifer. The regional presence (Paradise Valley scale) of glacial outwash alluvium and the correlation of the distribution of the relatively warm groundwater temperatures with the outcrop pattern of glacial outwash deposits, is consistent with this interpretation.

Another possible explanation for the relatively warm temperatures of the Yellowstone aquifer glacial outwash deposits is that the glacial outwash deposits may receive recharge from an underlaying groundwater source such as the Madison Group aquifer. The presence of the Madison Group beneath the glacial outwash deposits in the floor of the valley suggests groundwater may be leaking to the glacial outwash deposits of the Yellowstone aquifer from the underlaying Madison Group. Groundwater leaking to the surface from depth may be distinguished by anomalous groundwater temperature patterns near the points of leakage such as the asymmetrical temperature profile noted here (Maxey, 1968; Cartwright, 1970). The actual temperature at which groundwater occurs in the Madison Group below the Yellowstone aquifer is unknown. No wells tap the Madison Group in the valley floor. Hence, any interpretations concerning Madison Group recharge based solely upon groundwater temperature within the study area are probably equivocal at best. Similarly, no unequivocal evidence for vertical recharge along the Laramide faults which cut the Madison Group is apparent in the groundwater gradient data (Figure 9). However, the recharge to the Yellowstone aquifer from the Madison Group should be identifiable from the chemistry of the aquifer waters.

Aquifer Chemistries

In order to delineate which aquifers may be the source for Armstrong and Nelson Springs, samples of the Madison Group groundwater, Yellowstone groundwater, and alluvial fan groundwater were collected for field and laboratory analysis and to compare the aquifer waters and the spring waters. The water chemistry of the springs should be virtually identical to the water quality of the aquifer from which it issues unless significant groundwater mixing occurs (Wilke, 1983; Levings, 1986). The assumption is made that the groundwater samples are representative of the average chemical composition of the groundwater yielded from the three aquifers. There was insufficient funding for extensive chemical analyses; however, the source of the water can be checked for the Madison Group and alluvial fan aquifer samples.

To test the validity of the assumption that the Madison Group sample is a valid representation of the quality of Madison Group groundwater in the study area, the chemical composition of the sample collected from well 3-10-19adc was compared with the chemical composition of five shallow (<116 m depth) Madison Group samples taken from the literature (Groff, 1965; and Zimmerman, 1966) (Table 9, Figure 14). The sample collected from well 3-10-19adc is similar to the composition of the samples taken from the literature. Therefore, the Madison Group sample collected during this study is considered to be representative of the quality of shallow Madison Group groundwater in the study area. The Madison Group samples in the diamond shaped field of the Piper diagram (Figure 14) denotes the range of variability that may be expected of Madison Group groundwater at depths less than 116 m.

Table 9. Representative groundwater quality of the Madison Group aquifer at depths less than 116 m. Source 1 - Groff, 1965; Source 2 - Zimmerman, 1966. Reported as (meq/l). TDS reported as (mg/l).

Source	Ca	Mg	Na+K	HCO_3	SO_4	Cl	TDS
1	1.85	1.75	0.00	4.70	0.39	0.20	150.00
1	2.55	2.25	0.00	5.52	0.94	0.17	220.00
1	3.30	1.08	0.08	8.45	0.17	0.06	231.00
2	1.85	1.83	0.00	6.29	0.21	0.11	182.00
2	4.60	2.00	0.13	9.29	0.38	0.74	330.00
3-9-10adc	1.46	1.67	0.19	3.35	0.19	0.04	185.90

Three groundwater samples were collected from different geographic locations within the alluvial fan aquifer (Figure 8). Two of the samples were collected from wells and the third sample was collected from a spring

(3-9-36bdd, 3-9-35dbb, and PS of Table 8). The percent cation and anion (meq/l) content was calculated for the samples to allow for comparison of

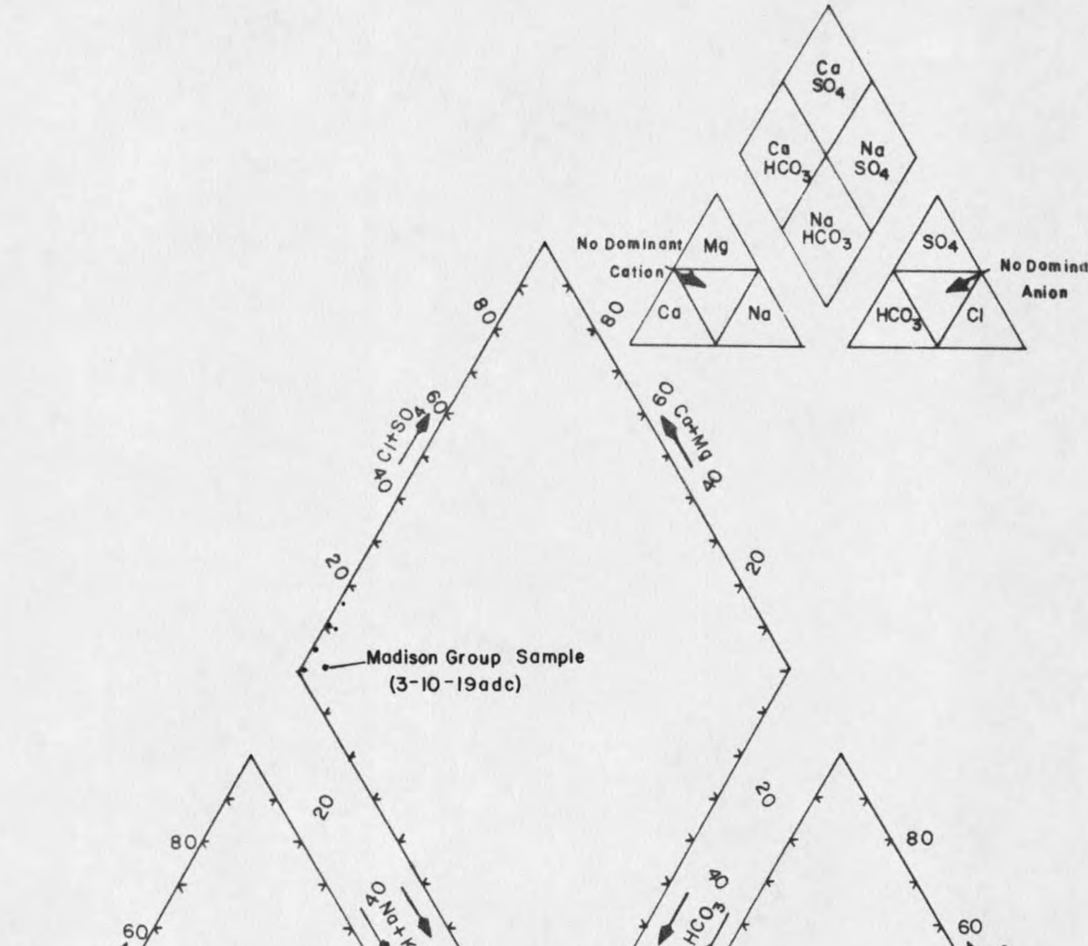


Figure 14. Piper diagram of representative Madison Group groundwater samples collected from the literature (Groff, 1965; Zimmerman, 1966) in comparison with the Madison Group sample collected from well 3-10-19adc.

samples on a Piper diagram (Figure 15). The three, widely spaced, samples are virtually identical. Therefore, the samples should be considered

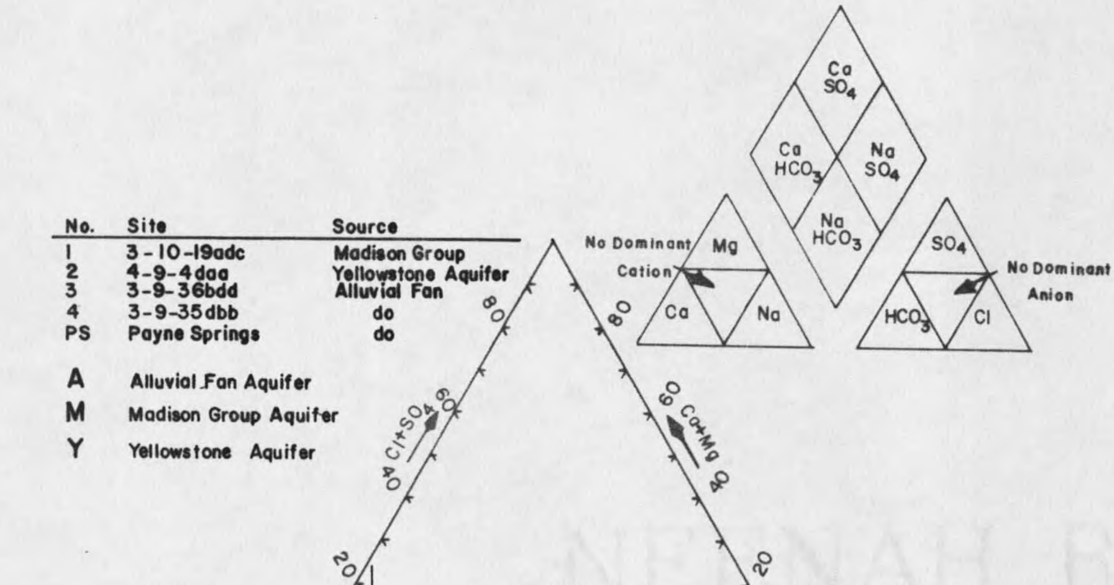


Figure 15. Piper diagram of samples from study area wells and from Payne Springs. For sample locations see Figure 8.

representative of the average quality of groundwater discharged from the alluvial fan aquifer.

Only one sample was collected from the Yellowstone aquifer (Table 8, Figure 15). Therefore, this sample is assumed to be representative of the average chemical composition of Yellowstone aquifer groundwater, but aquifer variability is unknown.

The chemical composition of groundwater from the three aquifers may be chemically distinct from one another (Figure 15). Arguments can be made that the chemical compositions of the three aquifer waters are not statistically distinct from one another, but the small number of samples is not conducive to statistical comparison. A larger sample population would be necessary to statistically test this interpretation. Samples from the three aquifers do, however, plot separately in each field of the Piper diagram with the exception of the Madison Group and the alluvial fan samples in the anion field (Figure 15). For the purpose of later discussion, the three aquifer waters will be interpreted to be chemically distinct. Separate fields of Madison Group, Yellowstone aquifer, and alluvial fan aquifer waters are depicted to clarify the interpreted distinction in water chemistry (Figure 15).

Armstrong and Nelson Springs

Now that the basin hydrogeology of the study area is established, the hydrogeology of Armstrong and Nelson Springs can be interpreted.

Spring Formation

A possible control on spring formation is the stratigraphy. Armstrong and Nelson Springs parallel the contact of the Yellowstone and the alluvial fan aquifers (Plate 1). The broad parallelism particularly of Nelson Springs to the contact of the two aquifers suggests the possibility that the springs may be stratigraphically controlled. Stratigraphic springs or contact springs form due to a significant change in formation hydraulic conductivity which precludes transmission of all groundwater which moves through the upgradient formation into the downgradient formation (Fetter, 1980). Contact springs generally issue in a line along the contact of two or more formations (Fetter, 1980).

Armstrong and Nelson Springs do roughly parallel the contact of the Yellowstone alluvium with the alluvial fan and glacial outwash (Plate 1); however, the hydraulic conductivities of the aquifers are similar based both on lithologic logs and on measured hydraulic properties (Table 6). Groundwater gradients and groundwater temperature patterns both suggest the aquifers are hydraulically continuous. Hence, the hydrogeology of the aquifers is not consistent with a contact spring origin.

The presence of major Laramide faults in the vicinity of Armstrong and Nelson Springs suggests the springs may be fault controlled. Additionally, in the nearby Centennial Valley of southwestern Montana, geophysical evidence suggests normal faults are buried beneath the alluvium covering the valley floor (Sondregger and others, 1982). The buried normal faults are interpreted to influence the formation of some springs in the Centennial Valley (Sondregger and others, 1982). Similar buried normal faults could be present beneath the Yellowstone aquifer in the Paradise Valley and control spring formation, but this idea was not tested during this study. The Laramide faults or buried normal faults could directly control spring formation by serving as a conduit for discharge water from the underlying faulted Madison Group aquifer. If fault zones serve as hydraulic conduits, such structural control of groundwater flow and spring formation should be reflected in groundwater gradients and there should be a correlation between the trace of faults and locations of springs (Maxey, 1968; Williams, 1970; Freeze and Cherry, 1979; Huntoon, 1976c, 1985a, 1985b). The Laramide faults trend east-west, normal to the orientation of Paradise Valley (Plate 1). In contrast, the springs issue alongside the Yellowstone River in a trend which roughly parallels the axis of the valley (Figure 2, Plate 3) which suggests the springs do not issue from the Laramide faults. With respect to possibility of buried normal faults that may control spring formation, no inference can be made except that if the springs were controlled by normal faults, the Madison Group aquifer would most likely discharge along the faults to the springs. As will be discussed later, chemical evidence suggests otherwise.

Faults are interpreted to be indirectly control the formation of Armstrong and Nelson Springs. The Yellowstone aquifer is interpreted to be thinned across the Laramide faults such that groundwater discharges at Armstrong and Nelson Springs. Paleozoic bedrock outcrops to the immediate north of the trace of the Center Hill fault which suggests a thin Cenozoic cover (Plate 1). Furthermore, Kirby (1940) determined that the thickness of the alluvium at the north end of the study area is only 15.5 m. As discussed earlier, the thinning of the aquifer across the faults limits the volume of groundwater that can be transmitted within the aquifer. Thus, groundwater is discharged at Armstrong and Nelson Springs. Buried normal faults could also influence the thickness of the Yellowstone aquifer and thus spring formation, but this is only speculation at this time.

The hydrogeology of the area suggests Armstrong and Nelson Springs are depression springs, localized along the axis of the valley in response to aquifer thinning. Depression springs form in topographically low areas where the water table intersects the ground surface (Fetter, 1980). The position of Armstrong and Nelson Springs on or directly bordering the Yellowstone River floodplain, the lowest topographic position in the valley, is consistent with depression spring formation (Figures 9 and 10). Hydrologic sections constructed through the Armstrong and Nelson Springs area clearly show the depressed topography at the positions of the Yellowstone River and the Armstrong and Nelson Spring areas (Figure 10). The hydrologic sections also illustrate that the depth to the water table decreases towards the floor of the valley and intersects the ground surface in topographically depressed areas. The topographically depressed areas correlate to where Armstrong and Nelson Springs issue.

The observed fluctuations in the size and number of individual Armstrong and Nelson Springs are consistent with depression spring formation. Groundwater in the alluvial fan and Yellowstone aquifers is interpreted to be unconfined. Therefore, the water table should fluctuate in response to seasonal changes in recharge (Freeze and Cherry, 1979; Fetter, 1980). As the water table fluctuates in response to recharge, the size of groundwater discharge areas should also fluctuate (Driscoll, 1986). For example, the size of the groundwater discharge area should increase when the water table rises in the discharge area because more of the water table should intersect the ground surface. The converse should be true for declines in the water table in the discharge area.

The pattern of increase and decrease in the size of the groundwater discharge area in response to recharge fluctuation may be represented by the fluctuation in the size and number of individual Armstrong and Nelson Springs observed during this study. During probable rising water-table conditions in spring and early summer, the size of the discharge area increased via an increase in the size and number of individual springs. Conversely, during the period of probable water-table decline from late summer into winter, the size and number of individual springs decreased which represents a decrease in the size of the groundwater discharge area (Table 3).

Comparison of Armstrong Spring Creek discharges measured at SA5, SA4, and SA3 on the same day suggests Armstrong Spring Creek fluctuates between gaining and losing conditions (Plate 3, Table 3). Gaining and losing conditions are commonly dictated by water-table position (Freeze and Cherry, 1979). Between April 17, 1987 and September 23, 1987, Armstrong Spring Creek generally gained flow between stations SA3 and SA5 (Table 3). For example, on May 22, 1987 discharges of 0.42 m^3/s at SA3, 0.45 m^3/s at SA4, and 0.57 m^3/s at SA5 (Table 3) clearly show the spring creek gains flow downstream. After September 23, 1987, Armstrong Spring Creek generally lost flow between stations SA3 and SA5 (Table 3). Gaining conditions generally reflect high water-table conditions and losing conditions tend to reflect low water-table conditions (Freeze and Cherry, 1979). Therefore, spring discharge appears to be controlled by water-table fluctuations. The apparent relationship between spring discharge and water-table elevation suggests the springs are depression springs which form where the water table intersects ground surface in topographically low areas.

This interpretation suggests that springs should also be present along the banks of the Yellowstone River. Springs may be present, but may simply be obscured due to the coarseness (cobbles and boulders) of the Yellowstone River channel deposits. Within the floodplain proper, fine-grained sediment is deposited during waning Yellowstone River flood stage. The fine-grained deposits may force groundwater to concentrate within more permeable zones and thus groundwater is concentrated to form well developed spring orifices. The ability to visually detect individual channel spring orifices is also related to sediment size. The discharge of any single channel spring is too minimal to be visually detected without a disturbance in the sediment to mark the presence of an orifice. In very fine-grained sediment, orifices may be detected where they are of sufficient size to be readily apparent. In fact, the primary indicator of the presence of spring A5 (the largest spring in the system) is that the discharge cascades over a terrace scarp. Therefore, the coarse-grained deposits of the Yellowstone River channel probably allow groundwater to discharge in a less concentrated manner which does not accommodate the development of easily identifiable spring orifices.

Source Aquifer of Armstrong and Nelson Spring Creeks

In order to determine from which aquifer(s) the springs discharge, water samples were collected from various Armstrong and Nelson Springs for field and laboratory analysis to compare spring and specific aquifer water chemistries (Figure 8, Tables 8 and 10). If the springs represent a single source aquifer, the springs should be chemically similar to the source aquifer. The springs plot within the calcium bicarbonate field and cluster closely together (Figure 16). The clustering of the spring samples suggests the springs are of similar water chemistry. It should be noted that the springs do exhibit some variation in individual water chemistry. The majority of chemical variation is in relative cation compositions. This no doubt represents some variation in aquifer composition between individual spring localities and possible variation in residence times of the individual spring waters in the host aquifer.

Visual inspection of the Piper diagrams suggests the springs are most similar in chemical composition to the Yellowstone aquifer sample (2) (Figure 17, Table 8). The chemistries of the springs are least like the composition of the Madison Group sample (1) (Figure 17). Furthermore, the chemical compositions of the springs are dissimilar to the sample from Payne Springs (PS) which issue from the alluvial fan aquifer. The similarity of the qualities of Armstrong and Nelson Springs to each other and to the Yellowstone aquifer sample (2) (Figure 17) indicates each of the springs discharge Yellowstone aquifer water. This is consistent with the fact that the springs issue from Yellowstone aquifer deposits (Plate 1).

If the Madison Group recharges the springs, the water chemistries should make that apparent. Hydrogeologic studies in other parts of Montana (Sondregger and other, 1982; Wilke, 1983) interpreted the Madison Group as a source for springs via Piper diagrams where samples collected from the Madison Group and from the springs plot at essentially the same position. In fact, in the Wilke (1983) study a Madison Group signature was apparent even after groundwater which leaked from the Madison Group had mixed with overlying aquifer waters. Based upon the Piper diagrams

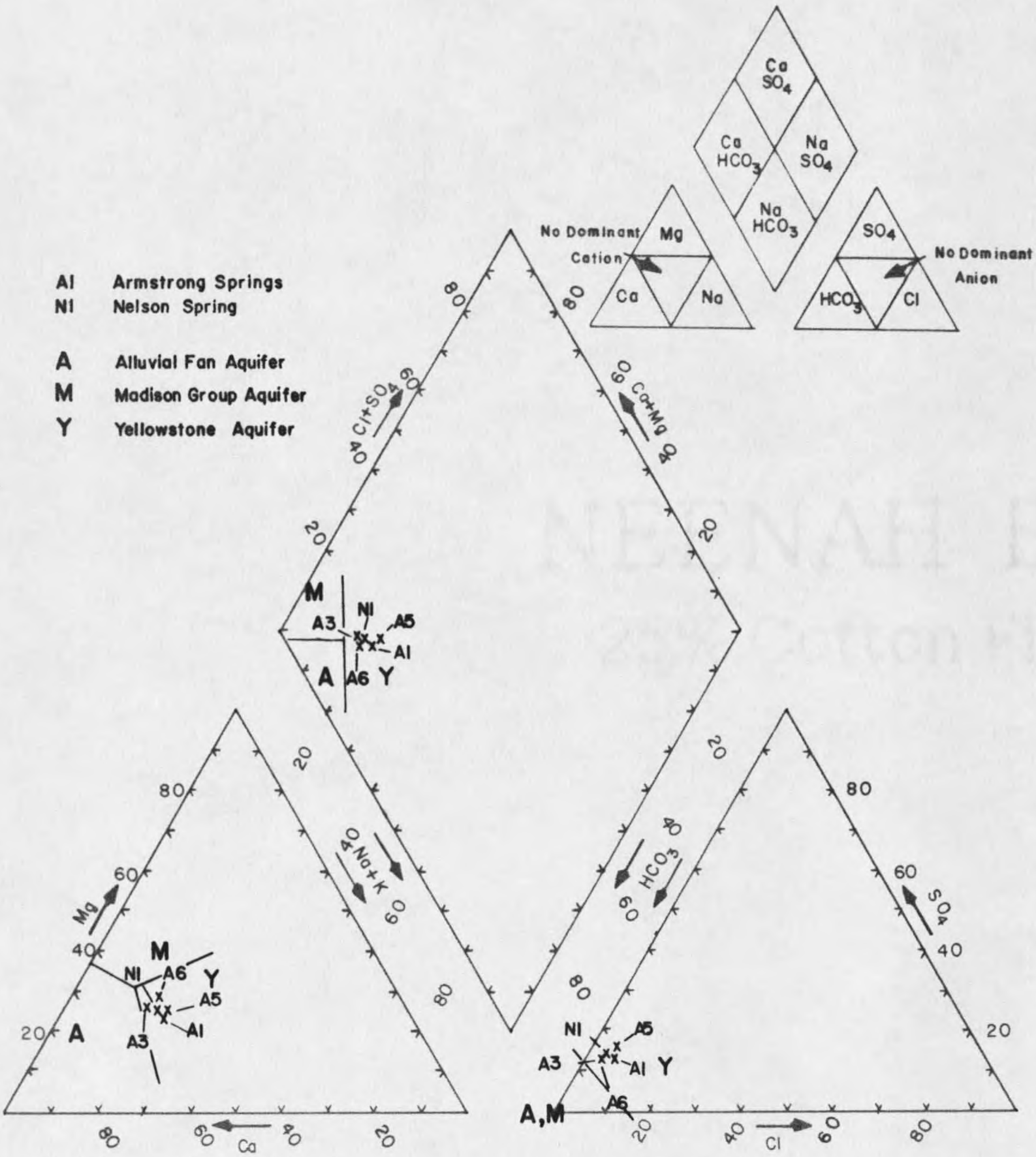


Figure 16. Piper diagram of Armstrong and Nelson Spring samples.

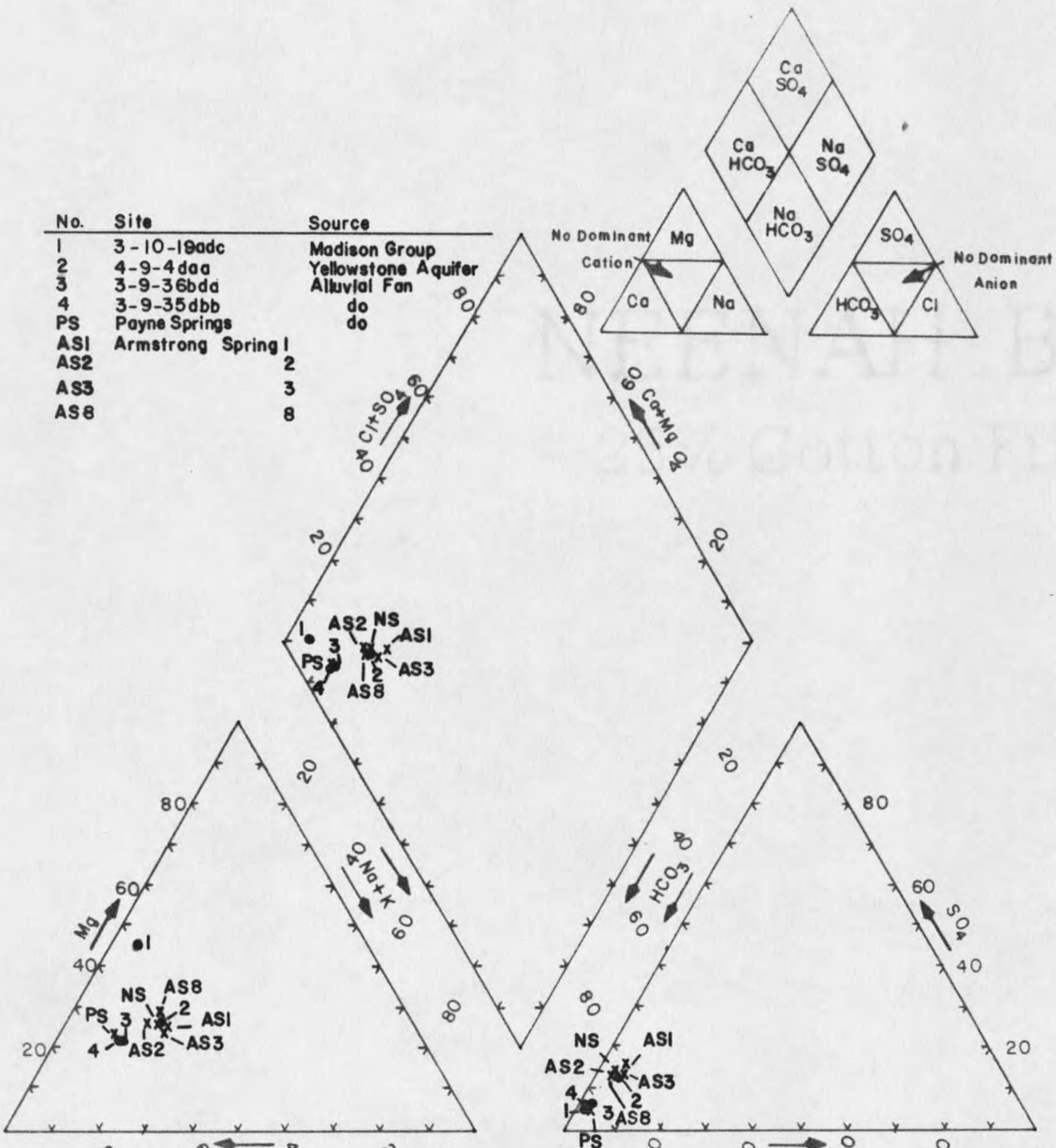


Figure 17. Piper diagram of well and spring samples.

the Madison Group is not a major contributing source to Armstrong and Nelson Springs, if this aquifer contributes water at all.

Another approach to chemically fingerprint the source aquifer is trace element data. Several cations were examined. None unequivocally point to the Madison Group as a source aquifer; however, each is consistent with the Yellowstone aquifer as the source. For example, the SiO_2 concentration of samples collected from Armstrong and Nelson Springs averaged 0.39 meq/l ± 0.06 meq/l (Table 8). The Yellowstone aquifer had a SiO_2 content of 0.44 meq/l. The Madison Group sample had a SiO_2 content of 0.21 meq/l and the SiO_2 content of samples collected from the alluvial fan averaged 0.18 meq/l ± 0.02 meq/l (Table 8). Clearly, only the Yellowstone aquifer is capable of delivering the SiO_2 content present in the springs.

Strontium was thought to be a potentially viable fingerprinting ion in order to determine the presence or absence of Madison Group aquifer leakage. Wyatt (1984) found an increase in the strontium/total dissolved solids (Sr/TDS) ratio in groundwater as Madison Group subcrop was passed. Strontium is released to groundwater when aragonite is converted to calcite during diagenesis, and if present, may suggest the presence of a carbonate source rock (Blatt and others, 1972). However, strontium is also released when strontium-bearing feldspar minerals decompose. Locally, the Tertiary volcanic rocks of the Gallatin Range and tuffaceous sediments in the valley floor may be another source of strontium.

Chemical analyses indicate the strontium argument is a poor indicator for discerning local Madison Group leakage in this area. The Sr/TDS ratio of Armstrong and Nelson Spring samples range from 5.5×10^{-4} to 7.6×10^{-4} (Table 8). The Sr/TDS of the Madison Group sample was 2.6×10^{-4}

(Figure 18). In this area then, the Sr/TDS in the springs exceeds that in the Madison Group rocks by 111 % (Table 8, Figure 18).

The high Sr/TDS ratios of the samples, if anything, support a non-Madison source for the springs. The relatively high Sr/TDS ratios in the groundwater samples collected from the springs and the Yellowstone aquifer may suggest the Yellowstone aquifer receives recharge from the Yellowstone River. Representative water quality samples for the Yellowstone River are taken from the literature (Klarich and Thomas, 1977; United States Department of the Interior, 1983, 1985) (Table 10) due to a lack of available funds for additional water sample analyses. The samples were collected at Carter's Bridge (United States Geological Survey stream gaging station 06192500). The Sr/TDS ratios in samples collected from the Yellowstone River exceed Sr/TDS ratios of each sampled spring (Tables 8 and 10). The Sr/TDS of the Yellowstone River ranges from 7.4×10^{-4} to 9.2×10^{-4} (Table 10) while the springs ranged from 5.5×10^{-4} to 7.6×10^{-4}. Therefore, the Yellowstone River is capable of supplying high Sr/TDS water to the springs. The high Yellowstone River strontium concentration is most likely received from the many thermal springs which are located upstream, in and near Yellowstone National Park (Leonard and others, 1978).

Chemical analyses, discharge, and temperature characteristics of Armstrong Springs supports a lack of Madison Group contribution. Springs which receive vertical recharge from the Madison Group at depth should be expected to be warm and exhibit consistent temperatures. Spring A5 was the warmest and most thermally constant spring monitored (Table 4). Therefore, if the Madison Group is the source for any spring, it is

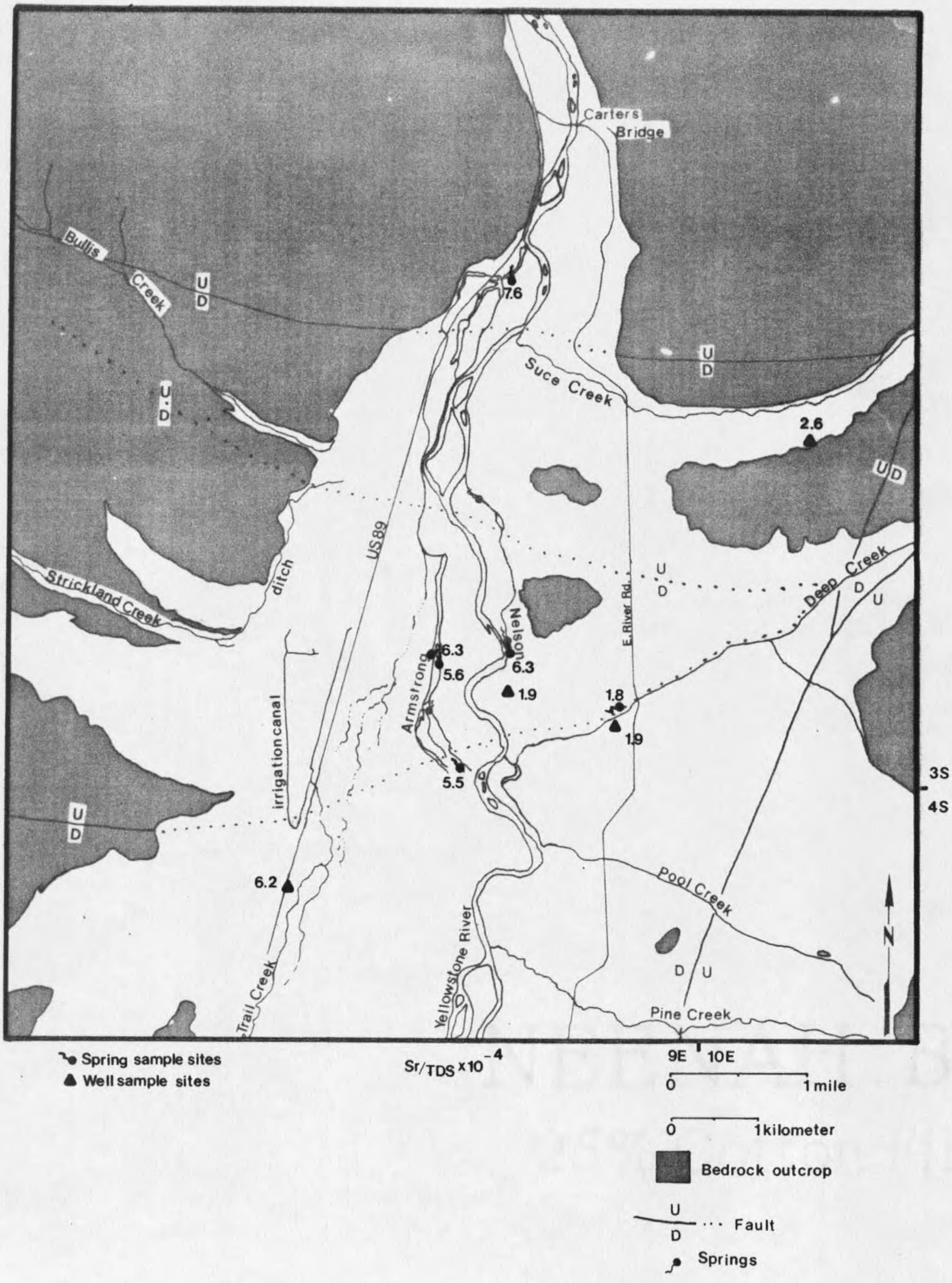


Figure 18. Strontium/total dissolved solids (Sr/TDS) ratios (meq/l/mg/l).

reasonable to expect that spring A5 is the most likely candidate to reflect recharge by the Madison Group. The A5 sample, however, is least similar to the quality of the Madison Group of any spring sampled (Figure 17) and therefore must not represent Madison Group groundwater. Therefore, based upon groundwater gradients, groundwater and spring temperatures, and aquifer and spring chemistries, Armstrong and Nelson Springs are interpreted to represent discharge from the Yellowstone aquifer. The Madison Group aquifer is interpreted to discharge little, if any, groundwater to the springs.

Table 10. Yellowstone River water quality. Source 1 - United States Department of the Interior, 1983. Source 2 - Klarich and Thomas, 1977. Source 3 - United States Department of the Interior, 1985. Ca = calcium (meq/l), Mg = magnesium (meq/l), Na = sodium (meq/l), K = potassium (meq/l), HCO_3 = bicarbonate, SO_4 = sulfate (meq/l), Cl = chloride (meq/l), TDS = total dissolved solids (mg/l), Sr = strontium (mg/l).

SOURCE	Ca	Mg	Na	K	HCO_3	SO_4	Cl	Sr/TDS
1	1.00	0.54	0.78	0.11	1.33	0.69	0.29	7.9×10^{-4}
1	1.00	0.58	0.83	0.11	1.34	0.62	0.29	
1	1.00	0.58	0.87	0.12	1.41	0.62	0.29	8.1×10^{-4}
1	1.00	0.56	0.74	0.10	1.31	0.58	0.25	
1	0.55	0.28	0.27	0.04	0.73	0.23	0.07	7.4×10^{-4}
1	0.70	0.41	0.52	0.06	1.00	0.35	0.17	8.0×10^{-4}
2	0.95	0.46	0.70	0.08	1.50	0.52	0.20	
3	1.05	0.55	0.74	0.10	1.22	0.54	0.25	8.7×10^{-4}
3	1.20	0.65	0.91	0.12	1.39	0.71	0.29	
3	1.20	0.64	1.00	0.12	1.34	0.73	0.34	7.9×10^{-4}
3	0.70	0.38	0.41	0.07	0.79	0.40	0.12	
3	0.55	0.29	0.35	0.05	0.66	0.21	0.10	7.5×10^{-4}
3	0.90	0.50	0.65	0.09	1.09	0.44	0.21	9.2×10^{-4}

The Hydrogeology of Armstrong and Nelson Springs

Two Groundwater Flow Systems

Spring Temperatures and Discharges. Spring temperatures and discharges suggest spring A5 and Armstrong Channel Springs may represent two different groundwater flow systems. Springs which represent the same groundwater flow system exhibit similarities in discharge patterns and temperature patterns (Eakin, 1966; Maxey, 1968; Jacobson and Langmuir, 1974; Rosenthal and Mandel, 1985). Therefore, it follows that if spring A5 and the channel springs do not represent the same flow system, dissimilarities in patterns of discharge and temperatures should indicate so.

Hydrographs of spring A5 and the channel springs reveal both short term fluctuations and peak period discharge differences (Figure 19) (Table 3). Spring A5 increased in discharge from 1.67 m^3/s on March 14, 1987 to a peak of 2.97 m^3/s on July 2, 1987. Channel spring discharges increased from 0.31 m^3/s on March 14, 1987 to a maximum recorded discharge of 0.96 m^3/s on September 23, 1987, two months after spring A5 peaked (Figure 19). Statistical regression between the measured discharges of spring A5 and the channel springs yields a correlation coefficient of only 0.04. The lack of correlation may be due to either a lag time response to recharge fluxes between the two spring groups, or the presence of two different flow systems.

Temperature patterns suggest the lack of correlation in discharge between the two spring groups is due to different flow systems. Similar temperatures between these two groups should be expected if the same flow system is represented because the two groups are in similar position with respect to groundwater flow patterns (Figure 9). The similar position of the two groups suggests groundwater of the same temperature would be discharged. However, temperatures of spring A5 range from 11.5°C to 12°C, with an average of 11.9°C and a variance of 0.2°C (Table 4, Figure 20). The channel springs vary between 8°C and 11°C. In particular, channel spring A3 ranges from 8°C to 11°C, with an average of 9.4°C and variance of 0.9°C (Table 4, Figure 20). Despite the similarity of position of the two spring groups with respect to groundwater flow, the two groups discharge water of contrasting temperatures which suggests different flow systems.

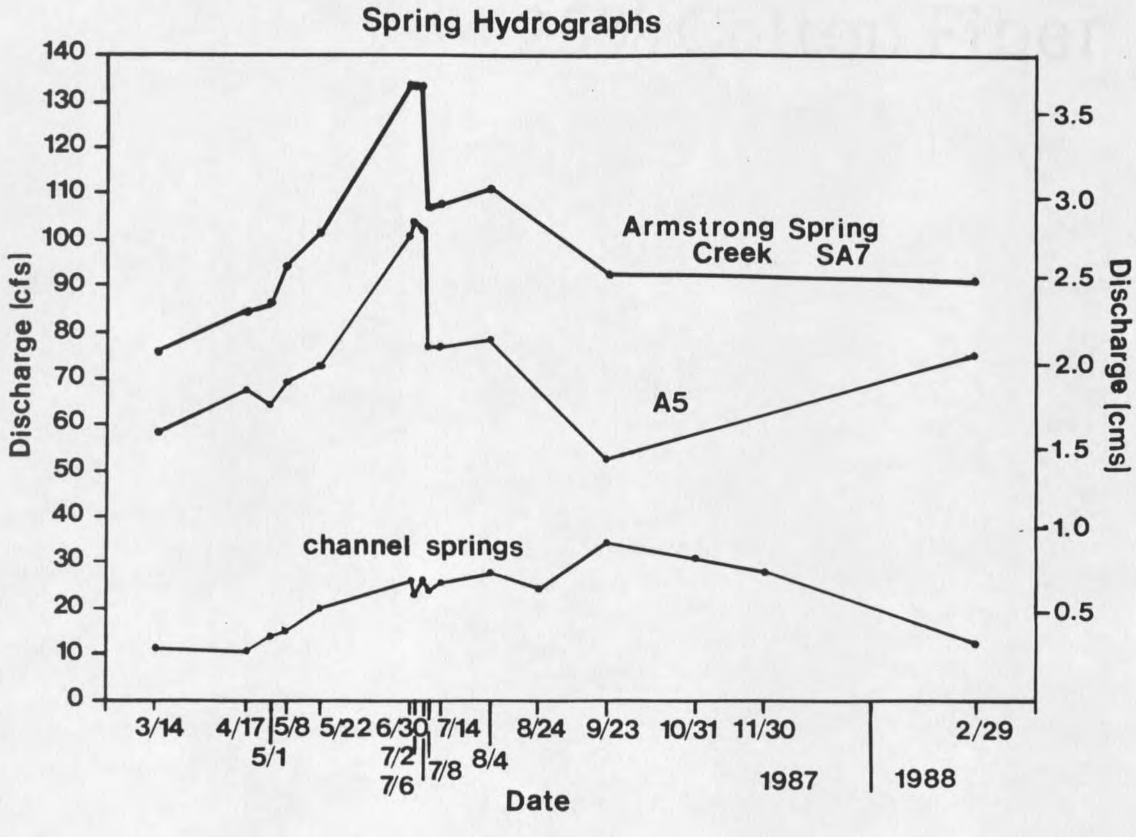


Figure 19. Hydrographs of Armstrong Springs. Discharge based upon direct measurement and estimation from the AS4 rating curve for channel springs.

Finally, temperature patterns of spring A5 and channel spring A3 contrast in their response to seasonal temperatures. Channel spring A3 is used for comparison because it has the most complete temperature record. The temperatures of spring A3 show a general increase in temperature from March through September (Figure 20). The increase in temperatures may reflect climatic warming into the summer season. The 29 year temperature record (1951-1980) for the Livingston area indicates mean monthly temperatures increase from -4.4°C in January to 18.7°C in July (Ruffner, 1985). The warming measured at spring A3 suggests a recharge source may be warming in response to seasonal atmospheric warming which occurs in the study area from January through July.

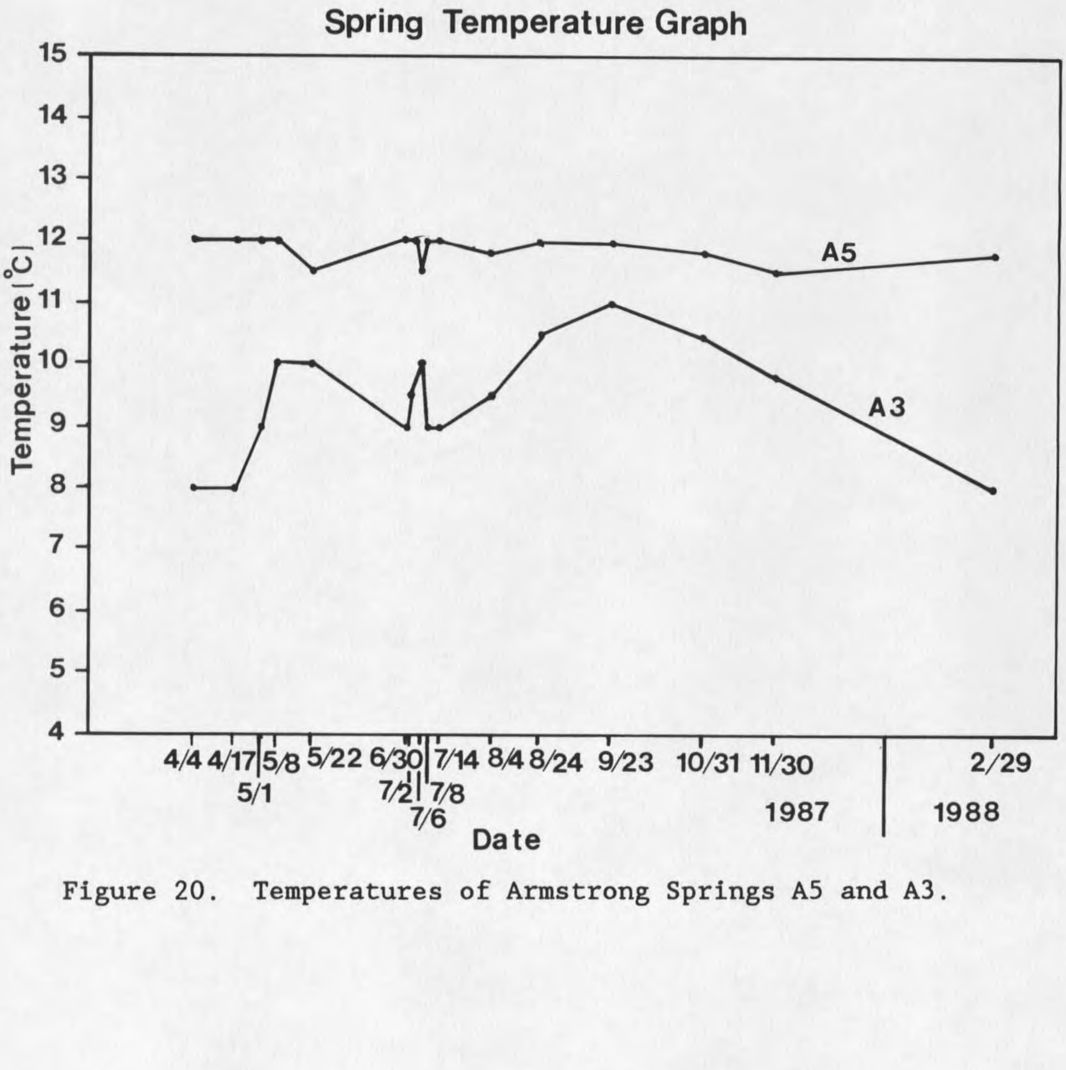


Figure 20. Temperatures of Armstrong Springs A5 and A3.

In contrast, spring A5 shows no response to seasonal atmospheric temperature changes (Figure 20). The temperature variance of spring A5 is only 0.2°C. The consistent, relatively warm temperature of spring A5 points to a groundwater flow system which is different from the flow system represented by the channel springs. The contrast between spring A5 and channel spring A3 with respect to temperature response to seasonal temperature patterns, fully supports different channel spring and spring A5 flow systems.

Geomorphology. The geomorphology of the area also suggests spring A5 and the channel springs may represent two different groundwater flow systems. Spring A5 discharges from the top of a terrace bordering Armstrong Spring Creek. The eastern extent of the terrace is interpreted to be the border between the glacial outwash deposits and floodplain alluvium of the Yellowstone River (Plate 1). The top of the terrace is approximately 3.5 m above the active Yellowstone floodplain. Armstrong Channel Springs and Nelson Springs, with the exception of N1, issue exclusively from the active Yellowstone River floodplain at approximately the same elevation as the adjacent Yellowstone River. Furthermore, the point of origination of the springs which originate from the Yellowstone River floodplain is coincident with the initial point of establishment of a well-developed Yellowstone River floodplain (Plate 1). The elevations of the springs relative to each other and to the adjacent Yellowstone River and the fact that not all springs issue from the same geomorphologically distinct alluvial body suggests the presence of more than one groundwater flow system.

Hydrogeologic Model for Armstrong and Nelson Springs

The hydrogeologic model proposed to explain the formation of Armstrong and Nelson Springs is consistent with the geologic characteristics of Paradise Valley, groundwater and spring temperatures, chemical compositions, and discharges monitored during this study. Armstrong and Nelson Springs represent a regional and local groundwater discharge area for the unconfined Cenozoic Yellowstone aquifer. The development of the discharge area is directly related to the downgradient thinning of the Yellowstone aquifer.

Regional and Local Groundwater Flow Systems

The springs discharge from the Yellowstone aquifer. The aquifer is probably recharged from the Yellowstone River, precipitation on the aquifer outcrop, the alluvial fan aquifer, and influent streams (Figures 9 and 10). However, the lack of similarity in total discharge, discharge patterns, and temperature patterns between spring A5 and the Armstrong Channel Springs and Nelson Springs as well as geomorphic evidence indicates that spring A5 represents a different groundwater flow system from that of Armstrong Channel Springs and Nelson Springs. Additional study is necessary in order to determine more precisely the paths and intricacies of the two groundwater flow systems, but some inferences can be made from the data collected during this study.

Spring A5 Groundwater Flow System. The groundwater flow system which discharges at spring A5 is interpreted to represent a regional groundwater flow system. Spring A5 discharges from the glacial outwash deposits of the Yellowstone aquifer approximately 3.5 m above the Yellowstone River floodplain. The Yellowstone River is likely the only recharge source capable of maintaining the large discharge of spring A5. The positioning of the spring at least 3.5 m above the Yellowstone River, however, makes it unlikely that the Yellowstone River locally recharges the glacial outwash deposits from which spring A5 discharges. The positioning of spring A5 above the Yellowstone River, on the other hand, does not rule out regional recharge from the Yellowstone River.

The Yellowstone River may regionally recharge the glacial outwash deposits from which spring A5 discharges. South of the study area, the glacial outwash deposits directly border the river from the southern extent of the valley to Pool Creek (Plate 1). Possibly as much as 9 km or more south of the study area (the extent of the highest glacial outwash surface), the Yellowstone River is envisioned to recharge the glacial outwash deposits. The Yellowstone River is approximately 53 m in elevation above spring A5 at that position. The elevation of the river provides the head for the flow system to discharge downstream at spring A5 3.5 m above the Yellowstone River floodplain. The influent Yellowstone River recharge migrates north within the glacial outwash deposits, and possibly deeper alluvium, as underflow and mixes with groundwater from local recharge sources (Figure 21). Eventually, the groundwater is forced towards the surface in the Armstrong and Nelson Springs area due to sufficient aquifer thinning at the positions of the Laramide faults (Figure 21). The regional extent and coarse-grained nature of the glacial outwash deposits provides the mechanism for the large discharge of spring A5. Spring A5 contributes approximately 76 % of the discharge measured at SA7 on Armstrong Spring Creek (Plate 3, Table 3, Figure 19). Armstrong Spring A5 also discharges 47 % more than the total combined flow of Nelson Springs (Table 3). The glacial outwash deposits represent an appreciable groundwater reservoir which, due to the probable high-energy environment in which it was deposited, has highly transmissive characteristics which promote the large spring A5 discharge. The regional-scale groundwater flow paths also provide the mechanism for the relatively warm, constant temperature of the groundwater discharged at spring A5 when compared to the temperatures of Armstrong Channel Springs and Nelson Springs (Table 4).

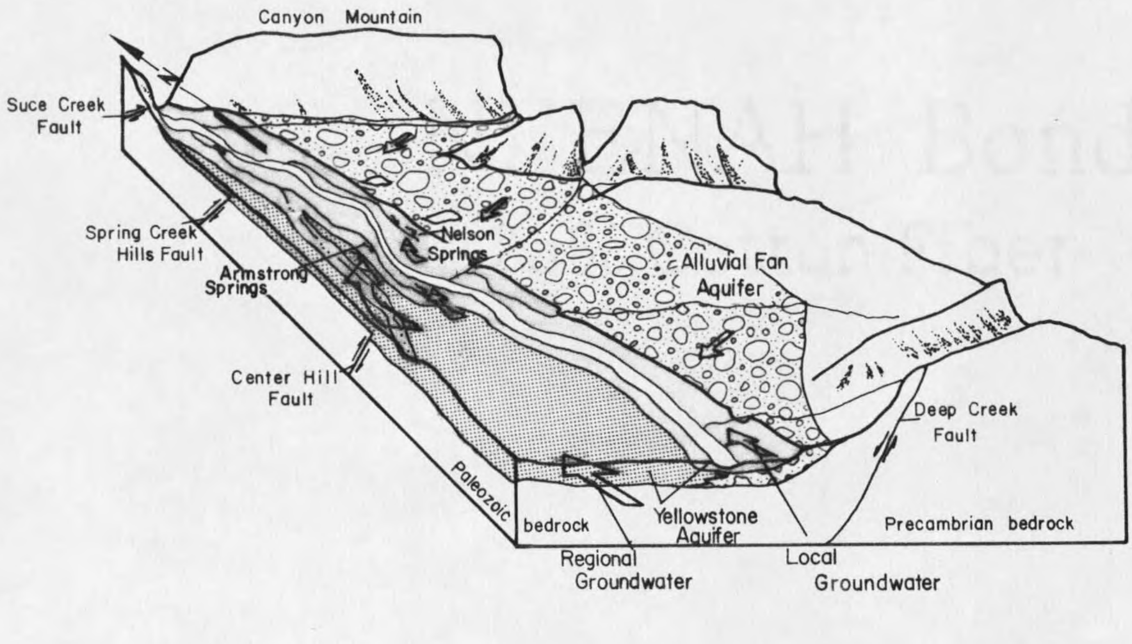


Figure 21. Diagrammatic interpretation of the regional and local groundwater flow systems discharging at Armstrong and Nelson Springs. (Not to scale.)

The temperature of the groundwater discharged at spring A5 is virtually constant (Figure 20). Furthermore, the temperature of spring A5 is consistently the warmest of any spring monitored during this study

(Table 4). The thermally constant nature and the relatively warm temperature of this spring suggests the associated groundwater flow system is not influenced by seasonal fluctuations in the mean atmospheric temperature or temperatures of surface water recharge. The lack of seasonal groundwater temperature fluctuations is consistent with regional groundwater flow systems (Maxey, 1968). The constant temperatures are probably due to substantial residence time of the groundwater in the aquifer prior to discharge from the aquifer. In short, the groundwater has sufficient time to equilibrate with the base aquifer temperature prior to discharge from the aquifer.

<u>Armstrong Channel Springs and Nelson Springs Groundwater Flow System</u>. The groundwater flow system which discharges at the Armstrong Channel Springs and Nelson Springs is interpreted to represent a local groundwater flow system (Figure 21). The channel springs issue from the Yellowstone River floodplain at approximately the same elevation as the adjacent Yellowstone River (Roberts, 1964a). Also, the origination point of the springs coincides with the point of broadening of the Yellowstone River floodplain (Plate 3). Therefore, a direct (localized) hydrologic connection between the Yellowstone River and Armstrong Channel Springs and Nelson Springs is suggested.

The discharge of Armstrong Channel Springs and Nelson Springs is small when compared to that of spring A5. The small discharge of the springs is consistent with the local groundwater flow system interpretation and is also related to the floodplain environment from which they issue. The local groundwater flow system is interpreted to be primarily comprised of only the deposits of the Yellowstone River floodplain and the alluvial fan aquifer (Figure 21), both of which are composed of poorly-sorted deposits which reflect the fluctuating depositional energy under which the units were deposited. The limited outcrop area of these deposits, the position of these deposits primarily in the immediate vicinity of the springs (Plate 3), and the poor sorting of these deposits suggests only a local flow system with a small discharge capacity should be developed.

The seasonally dependent temperatures of the Armstrong Channel Springs and Nelson Springs (Figure 20, Table 4) are also indicative of a local groundwater flow system (Maxey, 1968). The temperature of the groundwater within the local flow system is influenced indirectly within the aquifer by atmospheric temperatures via seasonal warming and cooling of the shallow aquifer materials. The short flow paths do not provide adequate groundwater residence times for the aquifer materials and recharge waters to equilibrate. The temperature of the groundwater is also directly influenced by the temperature of the recharge waters which also warm and cool in response to fluctuations of atmospheric temperatures. Therefore, the shallow groundwater flow system warms and cools in response to atmospheric temperature changes; the effects of which are ultimately displayed as the fluctuations in the temperatures of the Armstrong Channel Springs and Nelson Springs.

CONCLUSIONS

Three aquifers have been identified in the northern Paradise Valley: the Yellowstone aquifer, the alluvial fan aquifer, and the Madison Group aquifer. Each of the three aquifers is present in the immediate vicinity of Armstrong and Nelson Springs; however, only the Yellowstone aquifer is interpreted to be the source aquifer for the springs. All sampled Armstrong and Nelson Springs issue water of essentially identical chemical composition. The composition of the spring waters closely matches that of the Yellowstone aquifer. The springs and the three aquifers yield calcium bicarbonate-type water, but only the springs and the Yellowstone aquifer water are enriched in SiO_2 and exhibit a high Sr/TDS ratio.

The Yellowstone aquifer is a wedge-shaped alluvial body which thins from approximately 243 m, 12 km south (upstream) of the study area to less than 15.5 m at the northern (downstream) terminus of the Paradise Valley and study area. The origin of Armstrong and Nelson Springs is directly related to the downgradient thinning of the Yellowstone aquifer. Groundwater migrates to the south as underflow in the upstream and middle sections of the valley. Groundwater is directed to the surface at Armstrong and Nelson Springs as the volume of groundwater the aquifer is able to transmit is diminished due to the downgradient reduction in aquifer cross-sectional area. The groundwater discharges at positions which are topographically low with respect to that flow system.

Armstrong and Nelson Springs are actually a spring system composed of many marshy areas, individual springs, and clusters of springs. The number and size of individual springs, and accordingly spring discharges, increase through the spring and summer seasons and decrease in fall and winter presumably in response to seasonal Yellowstone aquifer water-table fluctuations. The positioning of the springs correlates primarily only with topography and broadly with the traces of local faults and stratigraphic contacts. However, the springs are interpreted to be depression springs and not fault or stratigraphic springs which is fully consistent with observed seasonal variations in spring size, number, and discharge.

Armstrong Springs issue from two distinct alluvial bodies. The master Armstrong Spring (A5) and several marshes discharge approximately 3.5 m above the current Yellowstone River floodplain from the top of a terrace composed of glacial outwash deposits. The majority of Armstrong Springs (channel springs) and marshes discharge from the Yellowstone River floodplain. Nelson Springs also discharge from the Yellowstone River floodplain. The issuance of the springs from two distinct alluvial bodies and the elevational contrasts suggest the glacial outwash springs (A5) and the floodplain springs (channel springs) represent separate groundwater flow systems.

Armstrong Spring (A5) represents a regional groundwater flow system. Spring A5 discharges approximately 76 % of the measured Armstrong Spring Creek flow, is consistently the warmest of any spring, and is essentially invariable in discharge temperature. Spring A5 discharges from extensive glacial outwash deposits which are continuous upstream at least 9 km.

Spring A5 discharges 3.5 m above the Yellowstone River. Finally, the Yellowstone River is probably the only local recharge source capable of maintaining the large discharge of the spring. All factors point to a regional groundwater system.

The springs which issue from the Yellowstone River floodplain (Armstrong channel springs and Nelson Springs) represent a local groundwater flow system. The springs, though large in number, discharge substantially less water than spring A5. Armstrong Channel Springs discharge only 24 % of the measured Armstrong Spring Creek flow. Nelson Springs discharge on average only 53 % of that of spring A5. The temperatures of the springs fluctuate seasonally. The springs issue from the lowest position in the valley, at essentially the same elevation as that of the adjacent Yellowstone River. The area where the Yellowstone River floodplain first widens and becomes well developed coincides with the positions of the most upstream Armstrong and Nelson Springs. Therefore, the Yellowstone River may locally influence the springs.

In order to test the conclusions drawn from this study, future endeavors should focus on relationships between the stage of the Yellowstone River, Yellowstone aquifer water-table elevations, spring discharges. Hydrogeologic studies should also focus on the Yellowstone aquifer south of this study area to test the interpretation of local and regional flow systems within the Yellowstone aquifer. Clearly, from the paucity of hydrogeologic data for the valley and the complex Cenozoic structural and glacial history of the valley, the possibilities for hydrogeologic study in the valley are limited only by imagination.

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APPENDICES

APPENDIX A

DRILLERS' WELL REPORTS

Table 11. Records of study area wells.

Well	Owner	Year Drilled	Drilling Method	Surface Elevation (m)	Casing and Screen Data: Casing or Screen	Diameter (m)	Depth (m) (from)	Depth (m) (to)
3-9-12cca	Dana			1392				
3-9-14dbb	Dana			1405				
3-9-14ddb	Dana			1405				
3-9-22cbd	Brawner	1961		1425	c	0.15	0	23
					s			open end
3-9-23dcc	Nelson	1963	Churn	1404	c	0.15	0	12.2
					s	0.15	11.6	12.2
3-9-23dcd	Nelson	1977		1413				
3-9-27bbb	Ohair	1985		1420	c	0.15	0	24
					s	0.15	17	20
3-9-28baa	Spring Creek Hills	1987		1536				
3-9-28bba	Spring Creek Hills	1971	Rotary	1535	c	0.15	0	64
					c	0.13	58	113
					s	open hole	113	128
3-9-28bbb	Spring Creek Hills	1981	Cable	1535	c	0.15	0	76
					s	0.15	59	76
								open end
3-9-32daa	Rose	1980	Forward Rotary	1474	c	0.15	0	18
					c	0.375	18	43
					s	0.375	24	31
					s	0.375	37	43
								open end

Table 11. Records of study area wells-Continued.

Well	Owner	Year Drilled	Drilling Method	Surface Elevation (m)	Casing and Screen Data: Casing or Screen	Diameter (m)	Depth (m) (from)	Depth (m) (to)
3-9-33bdd	Ohair	1976	Cable	1430	c	0.15	0	11
					s			open end
3-9-33dcb	Smith			1433				
3-9-35acc	Jumping Rainbow Ranch	1968	Churn	1420	c	0.15	0	21
					s	0.15	21	23
3-9-35baa	Jumping Rainbow Ranch	1968	Churn	1414	c	0.20	0	11
					s	0.20	11	15
3-9-35dbb	McAdams			1423	c	0.15	0	18
					s	0.15	18	20
3-9-35daa	Carter	1985	Forward Rotary	1447	c	0.15	0	28
					s			open end
3-9-36adc	Viers	1985		1502	c	0.15	0	37
					s			open end
3-9-36bda	Payne	1965	Churn	1477	c	0.15	0	11
					s			open end
3-9-36bdd	Forrest	1976	Cable	1468	c	0.15	0	36
					s			open end
3-9-36cab	Jones			1466				
3-9-36cbc	Johnson	1981	Forward Rotary	1452	c	0.15	0	31
					s			open end
3-9-36ccb	Speare	1983	Forward Rotary	1451	c	0.15	0	28

Table 11. Records of study area wells-Continued.

Well	Owner	Year Drilled	Drilling Method	Surface Elevation (m)	Casing and Screen Data: Casing or Screen	Diameter (m)	Depth (m) (from)	Depth (m) (to)
3-9-33bdd	Ohair	1976	Cable	1430	c	0.15	0	11
					s			open end
3-9-33dcb	Smith			1433				
3-9-36ccc	McGrath	1978	Cable	1449	c	0.15	0	40
					s	0.15	30	31
								open end
3-9-36cdb	Kneuschel	1987		1476				
3-9-36dbb	Scanson			1481				
3-9-36dbd	Melvin	1977		1492				
3-9-36dca	Garby	1977		1497				
3-9-36dcb	Waddington			1483				
3-10-19abc	Lewis	1974	Cable	1514	c	0.15	0	28
					s	0.15	24	25
								open end
3-10-19aca	Priest	1984		1521				
3-10-19adc	Bernthal	1985	Air Rotary	1558	c	0.20	0	116
					s	0.20	104	116
3-10-19bac	Broderick	1974	Cable	1501	c	0.15	0	28
					s	0.15	20	20.5
								open end
3-10-19bdb	Campanella	1983	Forward Rotary	1484	c	0.15	0	24

Table 11. Records of study area wells-Continued.

Well	Owner	Year Drilled	Drilling Method	Surface Elevation (m)	Casing and Screen Data: Casing or Screen	Diameter (m)	Depth (m) (from)	Depth (m) (to)
3-9-33bdd	Ohair	1976	Cable	1430	c	0.15	0	11
					s			open end
3-9-33dcb	Smith			1433				
3-10-31aba	Bowers	1983	Forward Rotary	1570	c	0.20	0	24
					c	0.15	24	38
					s	0.15	24	38
								open end
3-10-31abd	Abbott			1572				
3-10-31ada	Bridges	1979	Cable	1611	c	0.15	0	18
					s			open end
3-10-31bba	Bowles			1513				
3-10-31bbd	Morgan	1982		1526				
3-10-31cab	Nelson	1978	Forward Rotary	1555	c	0.15	0	29
					s			open end
3-10-31cba	Nelson	1973	Cable	1550	c	0.15	0	21
					c	0.10	6	27
					s	0.10	21	27
3-10-31cbb	Hammes	1977	Cable	1535	c	0.15	0	3
					c	0.15	3	27
					s			open end
3-10-31cbc	Duraham	1979		1530				
3-10-31cca	Lombard			1542				

Table 11. Records of study area wells-Continued.

Well	Owner	Year Drilled	Drilling Method	Surface Elevation (m)	Casing and Screen Data: Casing or Screen	Diameter (m)	Depth (m) (from)	Depth (m) (to)
3-9-33bdd	Ohair	1976	Cable	1430	c	0.15	0	11
					s			open end
3-9-33dcb	Smith			1433				
3-10-31cdb	Kovar	1987	Forward Rotary	1558	c	0.15	0	43
					s	0.15	35	40
								open end
3-10-31dbb	Bergsing	1978	Cable	1587	c	0.15	0	27
					c	0.15	27	30
					s			open end
3-10-31dcd	Chase	1981	Forward Rotary	1611	c	0.15	0	3
					c	0.15	3	26
					s			open end
4-9-1baa	Bandstra	1976	Cable	1467	c	0.15	0	33
					s			open end
4-9-1bad	Vietz	1967	Cable	1472	c	0.15	0	35
					s			open end
4-9-1cad	Kamps	1978	Forward Rotary	1480	c	0.15	0	40
					s			open end
4-9-1dad	Berg			1506				
4-9-4daa	Lindeman	1978	Cable	1434	c	0.15	0	21
					s			open end
4-9-4dda	Payne	1977		1436	c	0.15	0	22
					s			open end

Table 11. Records of study area wells-<u>Continued</u>.

Well	Owner	Year Drilled	Drilling Method	Surface Elevation (m)	Casing and Screen Data: Casing or Screen	Diameter (m)	Depth (m) (from)	Depth (m) (to)
3-9-33bdd	Ohair	1976	Cable	1430	c s	0.15	0	11 open end
3-9-33dcb	Smith			1433				
4-9-9dac	Jordan	1975	Cable	1440	c s	0.15	0	22 open end
4-9-10caa	Frelich	1978	Cable	1437	c s	0.15	0	20 open end
4-9-11aad	Durgan	1981	Forward Rotary	1469	c s	0.15	0	32 open end
4-9-11abb	Durgan	1971	Forward Rotary	1439	c s	0.15	0	18 open end
4-9-11bdb	Pollock	1982	Cable	1425	c s	0.15	0	9 open end
4-9-11cdd	Paradise Valley KOA	1966	Churn	1428	c s	0.15	0	12.3 open end
4-9-12bbd	Carter	1984	Forward Rotary	1474	c s	0.15	0	43 open end
4-10-5bbc	USDA Forest Service			1684				
4-10-6bbd	White	1974	Cable	1541	c s	0.15	0	32 open end

APPENDIX B

LITHOLOGIC LOGS

Table 12. Lithologic logs for wells utilized in the study.

Well 3-9-23dcc

Owner: Ed Nelson
Driller: Jones Drilling

	From (m)	To (m)
Top soil and sand	0	1
Gravel	1	7
Sand and fine gravel	7	10
Sand	10	12
Sand and fine gravel	12	12.2

Well 3-9-22cbd

Owner: Quincy Brawner

	From (m)	To (m)
Soil and boulders	0	1
Gravel (clay mixed)	1	6
Rock (hard-dry)	6	18
Rock and water	18	23

Well 3-9-27bbb

Owner: Ohair Ranch

	From (m)	To (m)
Topsoil	0	2.5
Sand and gravel	2.5	22
Hard blue shales	22	30

Well 3-9-28bba

Owner: Spring Creek Hills
Driller: Digger Dals

	From (m)	To (m)
Limestone gravel	0	24
Black shale	24	113
Sand, water @ 20 gpm (1×10^{-3} m^3/s)	113	128

Well 3-9-28bbb

Owner: Glen Alan
Driller: Van Dyken Drilling

	From (m)	To (m)
Dirt and rock	1	6
Soft shale	6	9

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)
Well 3-9-28bbb-Continued.		
Hard shale	9	37
Clay and rock layers	37	38
Clay	38	41
Rock	41	42
Hard rock	42	44
Shale	44	47
Limestone	47	49
Soft limestone	49	53
Medium hard limestone	53	59
Very hard limestone	59	62
Shale	62	66
Very hard shale	66	68
Hard shale	68	76
Well 3-9-32daa		
Owner: Roy Rose Driller: Hillman Drilling		
Gravel, clay	0	1
Well 3-9-32daa-Continued.		
Yellow clay	1	12
Red clay	12	15
Hard red clay	15	18
Hard grey clay	18	27
Little water, soft shales	27	34
hard shale, soft spots	34	43
Well 3-9-33bdd		
Owner: Ohair Ranch Driller: Hillman Drilling		
Soil, sand and gravel	0	11
Well 3-9-35acc		
Owner: Paul McAdams Driller: Jones Drilling		
Clay	0	3
Gravel	3	8
Clay	8	9

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)
Well 3-9-35acc-Continued.		
Sand and gravel	9	17
Sandy clay	17	18
Sand	18	19
Sand and fine gravel	19	19.2
Sand	19.2	20
Sand and gravel	20	23
Well 3-9-35baa		
Owner: Paul McAdams Driller: Jones Drilling		
Topsoil and gravel	0	2
Sand and gravel	2	15
Well 3-9-35dbb		
Owner: Paul McAdams Driller: Jones Drilling		
Gravel	0	6
Clay	6	8
Well 3-9-35dbb-Continued.		
Gravel	8	16
Clay	16	16.5
Sand	16.5	18
Sand and fine gravel	18	20
Well 3-9-35dda		
Owner: Howard Carter		
Topsoil	0	1
Sand and gravel	1	28
Well 3-9-36adc		
Owner: Don Viers		
Topsoil	0	1
Coarse overburden	1	27
Glacial wash, boulders, rocks, sand and gravel	27	37

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)
Well 3-9-36bda-Continued.		
Owner: James Payne Driller: Hulbert		
Soil	0	0.3
Soil and rock	0.3	1
Gravel	1	2
Gravel	2	5
Clay and rock	5	6.7
Gravel	6.7	7.3
Clay and rock	7.3	10
Gravel and water, coarse	10	12
Well 3-9-36bdd		
Owner: Ray W. Forrest Driller: Hillman		
Rock	0	2
Granite rock	2	16
Sand	16	24
Sand and gravel	24	35
Well 3-9-36bdd-Continued.		
Gravel	35	36
Well 3-9-36cbc		
Owner: Vernon Johnson Driller: Hillman		
Clay	0	1
Large boulders and gravel	1	18
Gravel and sand	18	24
Large boulders	24	26
Gravel and sand	26	30
Well 3-9-36ccb		
Owner: Robert L. Speare Driller: Hillman		
Topsoil	0	1
Sand, gravel, small boulders	1	28

Table 12. Lithologic logs for wells utilized in the study-Continued.

Well 3-9-36ccc

Owner: Stephen McGrath
Driller: Van Dyken

	From (m)	To (m)
Topsoil	0	1
Coarse claybound gravel	1	2
Boulders	2	4
Claybound gravel	4	8
Tight claybound gravels	8	10
Claybound gravels	10	23
Claybound coarse gravel and water	23	26
Silty claybound sand and gravel	26	29
Claybound sandy gravel	29	30
Gravel, some sand, 200gpm	30	30.5
Claybound gravel and sand	30.5	37

Well 3-10-19abc

Owner: Millard J. Lemon
Driller: Lindsay

	From (m)	To (m)
Topsoil	0	2
Gravel, clay	2	5
Clay, boulders	5	10
Illcomposed rock	10	15
Bedrock	15	28

Well 3-10-19bac

Owner: Oneita Broderick
Driller: Lindsay

	From (m)	To (m)
Topsoil	0	2
Gravel, clay	2	6
Illcomposed rock	6	13
Bedrock	13	28

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)		From (m)	To (m)
Well 3-10-19aca			Well 3-10-19aca-continued.		
Owner: Brian Priest			Clay	39	40
Topsoil	0	1	Rock ledge	40	43
Claybound gravel	1	9	Claybound gravel	43	44.5
Boulders	9	10	Rock ledge	44.5	45
Clay	10	11	Claybound gravel	45	46
Claybound gravel	11	14	Rock ledge	46	47
Boulders	14	15	Claybound gravel	47	56
Claybound gravel	15	22	Rock	56	56.4
Rock ledge	22	23	Claybound gravel	56.4	63
Gravel and clay	23	26	Rock ledge	63	63.4
Rock ledge	26	27	Claybound gravel	63.4	70
Clay, gravel, boulders	27	28	Rock	70	72
Claybound gravel	28	30	Claybound gravel-wet	72	76
Rock ledge	30	31	Rock	76	77
Sandy clay	31	33	Cemented gravel	77	78
Rock	33	34	Claybound gravel	78	80
Claybound gravel	34	37	Red clay and rock	80	92
Boulders	37	39			

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)
Well 3-10-19adc		
Owner: Peter Bernthal Driller: Potts		
Black topsoil	0	6
Fractured rock, clay streaks, brown and red	6	16
Dark gray to brown rock, medium hard	16	21
Gray to dark brown rock	21	24
Hard fractured limestone	24	103
Soft fractured limestone	103	105
Cavernous limestone	105	116
Well 3-10-19bdb		
Owner: Dr. R.A. Campanella Driller: Briggs		
Sand, gravel, clay	0	24
Well 3-10-31aba		
Owner: Tim Bowers Driller: Rock Creek		
Topsoil	0	0.3
Gravel	0.3	24
Gravel	24	38
Well 3-10-31ada		
Owner: Jeff Bridges Driller: Hillman		
Sand, gravel, clay, boulders	0	18
Well 3-10-31cab		
Owner: Mike Nelson Driller: Hillman		
Clay	0	1
Boulders, gravel	1	29

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)
Well 3-10-31cba		
Owner: Wayne Nelson Driller: Van Dyken		
Soil	0	1
Boulders	1	5
Gravel, clay	5	20
Sand, gravel, boulders and water	20	27
Well 3-10-31cbb		
Owner: George Hammes Driller: Hillman		
Clay	0	1
Rock, sand, gravel	1	5
Sand, gravel	5	9
Sand, gravel, rock	9	14
Sand, gravel	14	22
Sand, clay	22	25
Sand, gravel, rock	25	27
Well 3-10-31dbb		
Owner: Arne T. Bergsing Driller: Hillman		
Sand, gravel, rock	0	30
Well 3-10-31dcd		
Owner: Alston Chase Driller: Hillman		
Clay	0	1
Large boulders and clay	1	5
Gravel and sand	5	19
Large boulders and gravel	19	21
Gravel and sand	21	26
Well 4-9-1baa		
Owner: Anthony Bandstra Driller: Van Dyken		
Topsoil	0	0.3
Gravel, small boulders	0.3	3

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)		From (m)	To (m)
Well 4-9-1baa-Continued.			Well 4-9-1bad		
Gravel	3	4	Owner: Al Vietz Driller: Van Dyken		
Boulders	4	5			
Gravel	5	6.7	Soils, clay	0	0.3
Boulders	6.7	7	Sand, gravel and large boulders	0.3	10
Tight gravel, water	7	11			
Gravel, clay	11	14	Sand and gravels wet 110-114	10	35
Gravel	14	15			
Gravel	15	17	Clean sand and gravels water	35	37
Fine gravel	17	18			
Sand, gravel	18	20			
Sand, gravel	20	21	Well 4-9-1cad		
Loose gravel, sand	21	21.3	Owner: George Kamps Driller: Hillman		
Sand, gravel	21.3	24			
Gravel	24	26	Gravel and clay	0	6
Sandstone	26	27	Boulders and gravel	6	9
Gravel with rusty water	27	29	Gravel	9	15
Sand and gravel	29	33	Boulders	15	18
			Gravel with sand	18	26

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)
Well 4-9-1cad-Continued.		
Boulders	26	27
Gravel with sand	27	34
Well 4-9-4daa		
Owner: Diane Lindeman		
Driller: Hillman		
Gravel and clay	0	6
Sand and gravel	6	21
Well 4-9-4dda		
Owner: Clark Payne		
Driller: Hillman		
Sand, gravel, rock	0	6
Gravel, sand	6	22
Well 4-9-9dac		
Owner: Calvin Jordan		
Gravel	0	5
Well 4-9-9dac-Continued.		
Clay and sand	5	11
Sand	11	16
Sand and gravel	16	22
Well 4-9-10caa		
Owner: John A. Frelich		
Topsoil	0	0.15
Claybound gravel- boulders	0.15	2
Loose sand, fine gravel	2	7
Loose claybound gravel	7	12
Sand, some gravel	12	15
Sand	15	18
Tight sand and gravel	18	19
Sand, some gravel	19	19.8
Gravel and sand	19.8	20.4

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)
Well 4-9-11aad		
Owner: Larry J. Durgan		
Sandy Clay	0	1
Large boulders, sand and gravel	1	31
Well 4-9-11abb		
Owner: Larry j. Durgan Driller: Rock Creek		
Topsoil	0	0.3
Gravel	0.3	18
Well 4-9-11bdb		
Owner: Paul Pollock		
Sand and gravel	0	9
Well 4-9-11cdd		
Owner: Paradise Valley KOA Driller: Jones		
Sand	0	1
Gravel	1	5
Sand and boulders	5	9
Sand and gravel	9	12
Well 4-10-12bbd		
Owner: Howard Carter Driller: Hillman		
Gravel and boulders	0	7
Gravel and smaller boulders with some clay	7	33
Large boulders	33	35
Gravel, boulders with some clay	35	43

Table 12. Lithologic logs for wells utilized in the study-Continued.

	From (m)	To (m)
Well 4-10-6bbd		
Owner: White Driller: B&H		
Topsoil	0	1
Big boulders and gravel	1	11
Sand and gravel	11	32

APPENDIX C

DRILLERS' WELL REPORTS

Table 13. Tests reported in drillers' well reports.

Well	Test Date	Static Level (m)	Pumping Level (m)	Yield (x 10^{-3} m^3/s)	Test Time (hr)	Test Method
3-9-22cbd	10/25/61	15.2	18.3	0.9		Jet
3-9-23dcc	11/16/63	2.3	12.2	1.8	1	
3-9-27bbb	01/07/85	10.7	14.5	5.3	7	
3-9-28bba	05/13/71	51.8	91.4	1.0	4	Air
3-9-28bbb	02/18/81	59.4	68.6	0.9	1	Bailer
3-9-32daa	09/25/80	12.2	42.7	1.2	1	Air
3-9-33bdd	09/03/76	4.9	5.5	2.5	1	Bailer
3-9-35acc	02/09/68	7.3	18.2	5.7	2.5	Pump
3-9-35baa	02/22/68	0.2	10.1	11.7	5.5	Pump
3-9-35dbb		6.4	18.0	3.7	6	Pump
3-9-35dda	09/29/87	22.3	28.3	0.9	1	Air
3-9-36adc	04/10/85	21.3	29.0	1.5		Air
3-9-36bda	04/29/65	6.4	7.3	1.3	8	Pump
3-9-36bdd	03/16/76	29.3	30.5	1.3	1	Bailer
3-9-36cbc	07/17/81	12.2		0.6		
3-9-36ccb	05/17/83	20.1		1.9		
3-9-36ccc	04/26/78	22.9	29.0	1.9	1	Bailer
3-9-36dbd	1977			1.3		
3-9-36dca	1977			2.5		
3-10-19abc	07/17/74	19.2	26.2	0.6	2	Bailer
3-10-19aca	03/24/84	97.5	106.1	1.5	1	Bailer
3-10-19adc	04/04/85	89.9	90.2	1.3	4	Pump
3-10-19bac	07/09/74	14.0	24.4	0.8	2	Bailer
3-10-19bdb	07/28/83	19.5	24.4	0.6	2	Air
3-10-31aba	02/21/83	24.4	36.6	0.5	2	Air
3-10-31ada	10/12/87	10.7	16.8	0.7	1	Air
3-10-19bbd	1982			1.9		
3-10-19cab	09/22/78	8.23	25.0	1.3	1	
3-10-31cba	11/07/73	19.8	22.9	0.9	2	

Table 13. Tests reported in drillers' well reports-Continued.

Well	Test Date	Static Level (m)	Pumping Level (m)	Yield (x 10^{-3} m^3/s)	Test Time (hr)	Test Method
3-10-19cbb	09/30/87	21.3	24.4	1.0	1	Bailer
3-10-31cbc	1979			2.5		
3-10-19cdb	09/28/87	26.8	38.1	1.3	1	Air
3-10-31dbb	02/27/78	25.3	28.3	0.6	1	Bailer
3-10-31dcd	08/17/81	14.3	25.9	0.9	1	Air
4-9-1baa	09/04/1976	7.5	22.7	1.9	1	Bailer
4-9-1bad	06/19/67	12.2	21.3	2.3	2	Bailer
4-9-1cad	09/11/78	15.2	27.4	1.9	1	Air
4-9-4daa	05/01/78	15.2	18.3	1.3	1	Bailer
4-9-4dda	03/04/77	18.6	18.6	1.6	1	Bailer
4-9-4dac	10/17/75	16.8	16.5	1.6	1	Bailer
4-9-10caa	09/29/87	12.2	18.3	1.6	1	Bailer
4-9-11aad	10/08/81	15.8	32	1.3	1	Air
4-9-11abb	1971	15.2	17.7	1.9	2	Air
4-9-11bdb	03/30/82	4.0	4.9	1.3	1	Bailer
4-9-11cdd	06/28/66	2.4	12.3	2.5	1	Pump
4-9-12bbd	04/20/84	33.2	41.1	0.7	1	Air
4-10-5bbc				2.1	1	
4-10-6bbd	11/15/74	12.2	4.7	2.5	1.5	

PLATE I

Quaternary

Recent Alluvium (Yellowstone Aquifer)

Landslide Deposits

Alluvial Fan (Alluvial Fan Aquifer)

Glacial Outwash (Yellowstone Aquifer)

Glacial Till

Pediment

Tertiary Volcanics

Cretaceous

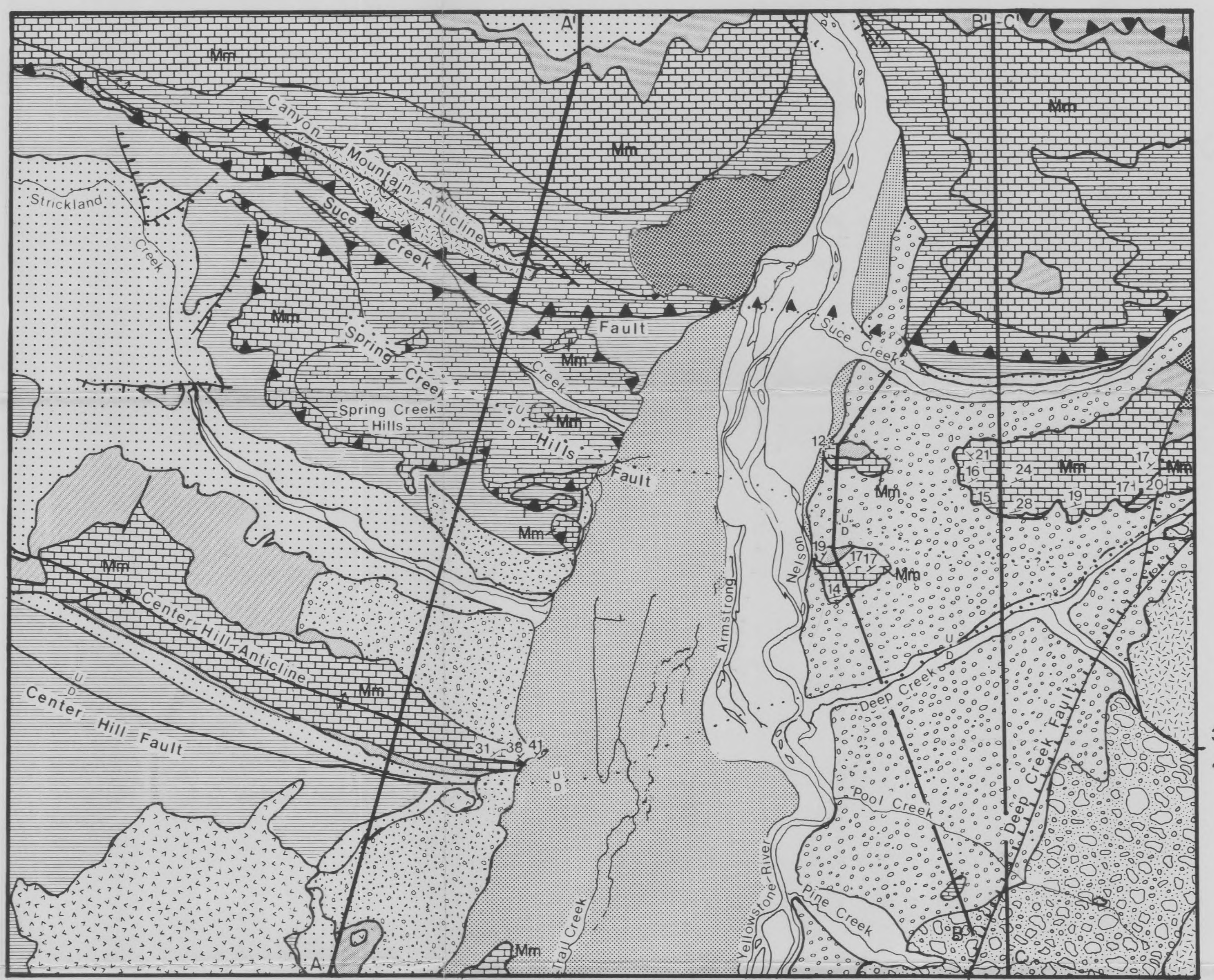
Jurassic

late Mississippian - Pennsylvanian

Mississippian Madison Group (Madison Group Aquifer)

Cambrian - early Mississippian

preCambrian



Thrust Fault (teeth on upper plate)

Reverse Fault U · upthrown side D · downthrown side

Normal Fault bars on downthrown side

Anticline

Overturned Anticline

lines of cross - section

Strike and Dip supplemental to Roberts 1964a, b

0 1km 1mi

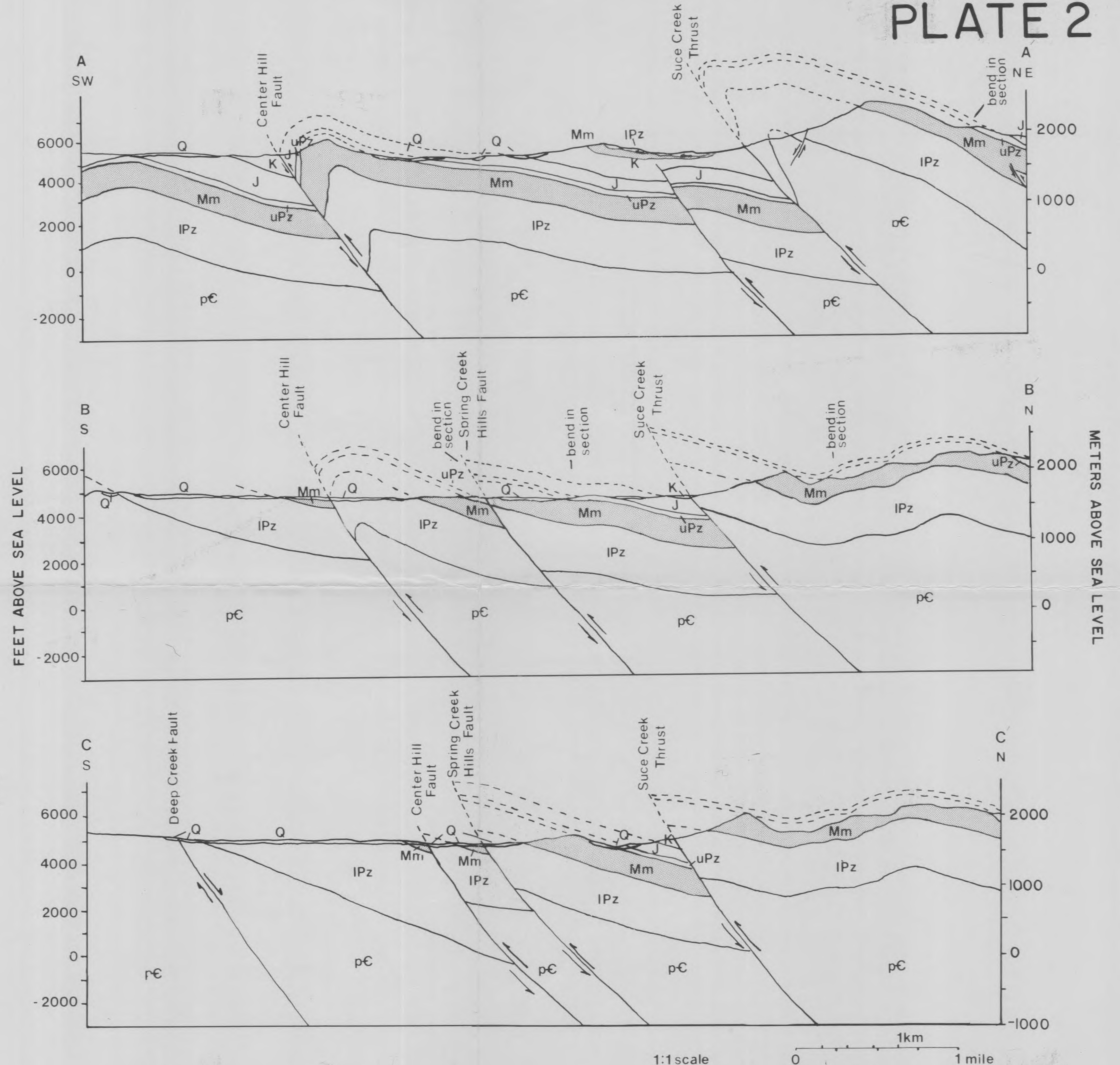
N

Geologic Map of Armstrong and Nelson Spring Creek Area, Park County, Montana

after Roberts 1964a, 1964b

modified by William Clarke, 1991

PLATE 2



Geologic Cross-Sections of Armstrong and Nelson Spring Creek Area, Park County, Montana

by William Clarke, 1991

PLATE 3

ARMSTRONG and NELSON SPRING CREEKS
PARK COUNTY, MONTANA

by William Clarke 1991