Ground Water Movement through the Highly-Fractured Crystalline Core of a Large-Scale, Foreland Uplift, Royal Gorge Arch, South-Central Colorado

Timothy L. Clarey
Western Michigan University

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GROUND WATER MOVEMENT THROUGH THE HIGHLY-FRACTURED
CRYSTALLINE CORE OF A LARGE-SCALE, FORELAND UPLIFT,
ROYAL GORGE ARCH, SOUTH-CENTRAL COLORADO

by

Timothy L. Clarey

A Dissertation
Submitted to the
Faculty of The Graduate College
in partial fulfillment of the
requirements for the
Degree of Doctor of Philosophy
Department of Geology

Western Michigan University
Kalamazoo, Michigan
April 1996
GROUND WATER MOVEMENT THROUGH THE HIGHLY-FRACTURED CRYSTALLINE CORE OF A LARGE-SCALE, FORELAND UPLIFT, ROYAL GORGE ARCH, SOUTH-CENTRAL COLORADO

Timothy L. Clarey, Ph.D.
Western Michigan University, 1996

The objectives of this study were twofold: (1) to examine the well-exposed, basement core of the Royal Gorge arch as a large-scale, field test of ground water flow through a fractured, crystalline, aquifer system; and (2) to devise a method to more accurately determine the direction of maximum aperture of a fractured aquifer through structural and hydrogeologic analysis.

The recognition of a preferred direction of maximum aperture in a fractured aquifer remains an important hydrogeologic consideration because flow through a fracture is proportional to the cube of the aperture width. Previous studies of hydraulic conductivity directions using fracture orientation techniques have met with varying degrees of success as many of the systems were complicated by multiple deformations, making the maximum stress direction, and corresponding maximum aperture direction, difficult to decipher.

Fracture orientations, fracture spacing, and other indicators of brittle deformation were collected at 565 outcrops within the basement core of Royal Gorge arch, and in the overlying sedimentary units. Over 500 foliation and bedding orientations were analyzed via stereographic projection, dividing the uplift into 10 homogeneous, structural domains. Arrays of slickensides were examined in each domain, resulting in determination of the stress field which caused the last episode of
brittle deformation. The orientation of extension fracturing, and corresponding maximum hydraulic conductivity, was then defined in each domain. Fifty water samples were collected and analyzed for stable isotopic signature, hydrogeochemical parameters, and ground water quality. Water well, spring, stream, isotopic, and hydrogeochemical data were used to define the water table surface within the aquifer and to define flow directions.

Results indicate that fracture orientation, fracture spacing, and the direction of maximum stress depend strongly on pre-existing basement fabric orientation and lithology. Ground water flow lines generally trend S 45-70 W, nearly identical to the direction of maximum stress interpreted for Laramide time. This direction also parallels the elevation gradient of the arch. All isotopic, hydrogeochemical, and potentiometric data are consistent with the interpretation that ground water is preferentially migrating along fractures parallel to the last operative direction of maximum stress.
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ACKNOWLEDGMENTS

First, I wish to thank my beautiful wife, Barbara Ann, who essentially did this research with me, and to her I owe this degree. Not only did she do field work by my side, but she ran the entire household, singlehandedly, for the last three years, allowing me to complete this degree requirement.

I also wish to thank my children, Ryan Timothy and Erin Catherine, for putting up with my stressed state of mind. Hopefully, I will be able to spend more evenings and weekends with my family once again.

Next, I wish to express my appreciation to my committee, Dr. Ronald Chase, Dr. Duane Hampton, Dr. R.V. Krishnamuthy, Dr. Alan Kehew, Dr. Christopher Schmidt, and Dr. Michael Stoline, for their guidance and support throughout.

My sincere thanks also goes out to the various organizations which partially funded this research: the American Association of Petroleum Geologists, the Geological Society of America, the WMU Graduate College Research Fund, the WMU Research Fellows Program, the Colorado Scientific Society Ogden Tweto Memorial Fund, the Wyoming Geological Society J.D. Love Field Scholarship, Chevron U.S.A., and the National Ground Water Association.

Special thanks also go out to Dr. W. Thomas Straw for supporting me both financially and emotionally during my return to college.

Finally, and above all, I would like to acknowledge the Lord Jesus for helping me every step of the way. He, and He alone, has created the world's geology, so much of which remains a mystery to man and is left to us to decipher. The following quotes from the Bible testify of Him (I Corinthians 1:20, 21, 25; I Timothy 6:20, 21; II
Acknowledgments - Continued

Timothy 4: 3, 4; II Peter 3: 4-7, NIV):

"Where is the wise man? Where is the scholar? Where is the philosopher of this age? Has not God made foolish the wisdom of the world? For since in the wisdom of God the world through its wisdom did not know Him, God was pleased through the foolishness of what was preached to save those who believe. For the foolishness of God is wiser than man's wisdom, and the weakness of God is stronger than man's strength."

"Timothy, guard what has been entrusted to your care. Turn away from godless chatter and the opposing ideas of what is falsely called knowledge, which some have professed and in so doing have wandered from the faith."

"For the time will come when men will not put up with sound doctrine. Instead, to suit their own desires, they will gather around them a great number of teachers to say what their itching ears want to hear. They will turn their ears away from the truth and turn aside to myths."

"They will say, 'Where is this coming He promised? Ever since our fathers died, everything goes on as it has since the beginning of creation.' But they deliberately forget that long ago by God's word the heavens existed and the earth was formed out of water and by water. By these waters also the world of that time was deluged and destroyed. By the same word the present heavens and earth are reserved for fire, being kept for the day of judgment and destruction of ungodly men."

Timothy L. Clarey
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CHAPTER I

INTRODUCTION

Historical Background of Problem

In recent years, research has increased considerably in the area of ground water flow in fractured crystalline rock. Literally, tens of thousands of residents across the United States rely on unconfined, fractured, crystalline aquifers for their drinking water supply. Studies for nuclear waste disposal in the subsurface also have considered repositories in fractured crystalline rock (Johnson and DePaolo, 1994; Sinnock and others, 1987; Benson and others, 1983). The need to monitor the performance of such disposal systems (and preserve drinking water supplies) has clearly demonstrated the importance of understanding ground water flow through basement complexes.

Fractured rocks are exposed or thinly-covered in about one-third of the world's landmasses, yet the understanding of their hydrogeology is still poorly documented (Fontes, 1980). The majority of hydrogeological studies concentrate on flow systems within a relatively homogeneous, porous medium, relying on established methods to describe the ground water movement. Flow studies of fractured rocks however, require detailed three-dimensional control as a consequence of the heterogeneity. Small-scale studies involving hydrogeologic testing and tracer experiments have repeatedly documented the highly discontinuous nature of fracture systems. As a result, preferential flow may become channeled along dominant fractures related to the strain prescribed by past or present stress fields acting on the rocks. Research by
Neretnieks (1985) at the Stripa mine in Sweden demonstrated the extreme localization of ground water flow due to channeling. One-third of the total flow was determined to have entered the mine from only 2% of the area monitored. Over 70% of the mine surface showed no ground water flow at all.

Tsang and Tsang (1989) and Moreno and Tsang (1994) recently published research which documented marked flow channeling in strongly heterogeneous media. They concluded that fluids tend to flow along least resistive pathways and with widely varying velocities. Unfortunately, these papers, like many others (e.g., Rubin, 1990; Thompson and Gelhar, 1990; Vomvoris and Gelhar, 1991), have concentrated on numerical simulations and not on actual field studies. Recent literature reviews of "real-world" examples have been largely confined to the Borden test site in Canada (Zhang and Neuman, 1990; Graham and McLaughlin, 1991; Rajaram and Gelhar, 1991), the Stripa mine in Sweden (Dverstorp and Andersson, 1989; Dverstorp and others, 1992; Neretnieks, 1993), and the Fanay-Augeres mine in France (Cacas, Ledoux, de Marsily, Barbreau, and others, 1990; Cacas, Ledoux, de Marsily, Tillie, and others, 1990; Long and Billaux, 1987). These field-based reports have been primarily based on numerical models and statistical examinations of the field sites, and limited to moderately heterogeneous media as opposed to strongly heterogeneous, fractured systems (Moreno and Tsang, 1994). They are also, almost exclusively, small-scale studies, usually within the range of 10s or 100s of meters in length and width.

Hydraulic conductivity in a fractured crystalline aquifer is determined by a combination of aperture width and fracture density. Knowledge of aperture width is critical when estimating ground water flow through fractures. Unfortunately, no adequate method exists with which to quantify the original aperture width prior to mineralization, in-filling, or later stress field adjustments. Villas and Norton (1977)
assumed that present-day openings provide reasonable estimates of paleoaperture values, finding no systematic changes in aperture width with depth or with lateral extent in the Mayflower stock, Park City, Utah. Haynes and Titley (1980) concluded that it is still necessary to assume constant apertures with time. They suggested that fracture density alone provides the best estimate of relative permeability of a fractured rock mass, leaving aperture changes out of their analysis because of the difficulty in quantifying this factor through time.

The "cubic law" states that for a given head gradient, flow through a fracture is proportional to the cube of the aperture width (Domenico and Schwartz, 1990). Although the validity of the cubic law has been demonstrated by many authors for fresh fracture surfaces (Witherspoon and others, 1980; de Marsily, 1986), it has been difficult to apply to older and weathered fracture systems (Neretnieks, 1993). Because many older fractures are filled with chemical precipitates and clays, the cubic law is invalidated without adjusting for variable fracture fill. Abelin and others (1985) found that the same fracture in the Stripa mine had very different hydraulic conductivities in boreholes only 0.5 m apart, implying a highly variable aperture width and/or partially clogged fractures. The large variation in hydraulic conductivity caused groundwater to travel in a nonuniform manner.

Most crystalline rocks consist of impermeable matrix blocks bounded by fracture planes. The interconnected porosity within blocks is generally very low: between 0.1% and a few percent (Grisak and Pickens, 1980). Typical hydraulic conductivities of the matrix blocks are less than $10^{-9}$ cm/sec, allowing little ground water to flow except through fractures (Gale, 1982). Many authors note that the permeability of fractured media decreases with depth, partly because fractures close with increasing effective stress and partly because of mineralization below the water.
Snow (1982) however, reported that workers in the Rand mines of India have observed copious amounts of water moving along faults to depths of over one kilometer. Field studies in eastern Africa indicate that, although fractures become less frequent with depth and finer in aperture, fluid pore pressures can be propagated along fractures to depths exceeding 3.5 km (Snow, 1982).

Research Objectives

The present study is a large-scale examination of ground water migration through an exceptionally well-exposed, fractured, crystalline aquifer in south-central Colorado (Figure 1). The research serves as a field test of theories of fractured flow which heretofore have only rarely been extrapolated beyond small-scale studies. Precambrian-age basement outcrops in the Royal Gorge arch are well-exposed in three-dimensions and cover approximately 60 km² areally, and 540 m vertically. A near perfect cross section is cut by the Arkansas River through the arch, exposing a canyon over 360 m deep into the crystalline core.

An additional objective is to devise a method to more accurately determine the direction of maximum aperture of the fractured crystalline system. According to theory, this direction should correspond with the direction of maximum hydraulic conductivity. Detailed structural analysis of the uplift, including kinematic analysis and stress field determination, is necessary to attain this goal.

As a consequence of the structural analysis process, insight has been gained concerning the structural evolution of the Royal Gorge arch. Although the uplift is primarily thought of as a "Laramide-age" event, it has a long and relevant history of deformation extending back into Precambrian time. An examination of the events creating the fracture system within the Royal Gorge arch was necessary prior to
Figure 1. Precambrian-age Rock Exposures in Colorado Showing the Location of the Study Area in Relation to the Front Range.

hydrogeologic study. This also included petrographic analyses to determine the mineralogies present in the aquifer matrix and to examine fractures for secondary mineralization. Accomplishment of this goal has also extended the research findings of Chase and others (1993) and Schmidt and others (1993) to a larger, mountain-sized, scale.

The ultimate goal of the project is the integration of detailed structural, hydrologic, and geochemical data into a three-dimensional, ground water flow analysis

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of the crystalline core of the Royal Gorge arch. The study will also serve as a typical example of ground water activity in the exposed cores of mountains or eroded ancient mountains.

To accomplish these objectives, extensive field work at the Royal Gorge arch was conducted during the summers of 1993 and 1994. The necessary three-dimensional fracture control was established by examining 565 outcrops in the Precambrian core of the uplift and in the surrounding sedimentary units. Fifty water samples were collected from available water wells, springs, and streams and analyzed for major ion concentrations and stable isotope ratios to varying degrees. A subset of the water samples was also sent to the University of Waterloo, Ontario, Canada for tritium analysis. Detailed methodologies for the structural analysis, water chemistry analysis, isotopic analysis, and additional analyses are listed below under appropriate chapter headings.

Location and Setting

The Royal Gorge arch is located in Fremont County, Colorado, immediately west of Canon City (Figures 1 and 2). U.S. Highway 50 gives excellent access to the study area, and bounds the east and north flanks of the arch. Temple Canyon Road and Copper Gulch Road (both gravel county roads) follow the south flank and west flank, respectively, and also intersect Highway 50 at the small community of Parkdale and Canon City. These three roads collectively encircle the Royal Gorge arch and allow direct access to the sedimentary units bounding the uplift. The Royal Gorge Bridge Road and associated "picnic" roads give additional access to the interior of the uplift, connecting Highway 50 and Copper Gulch Road.
Relief is generally rough throughout the study area. The elevation varies from 7233 feet (2212 m) above sea level at Fremont Peak to 5357 feet (1638 m) above sea level at the Arkansas River west of Canon City.

Colorado House Bill 35 has effectively stopped large land developments in Fremont County due to the unavailability of domestic water. Likewise, the growth of mobile home areas has ceased recently because of water supply problems. Most of the small development is occurring on the perimeter of Canon City, where piped domestic water is still available from the Arkansas River.

Along Highway 50, between Canon City and Parkdale, development is also dependent on water availability. Most of the domestic wells along this route tap fractures in the Precambrian core of Royal Gorge arch. Whenever larger developments, such as campgrounds, are not able to supply their water needs through wells, they are forced to truck in additional water from Canon City at considerable yearly expense. Most single-family dwellings within the study area are able to supply their water needs via the fractured aquifer, but commonly employ either a cistern or a holding tank to maintain a steady supply.

Climate and Precipitation Data

Canon City receives approximately 13 inches (33 cm) of precipitation per year, according to data supplied by the National Climatic Center, Asheville, North Carolina to the Soil Conservation Service. This leaves agriculture in the area very dependent on irrigation from the Arkansas River, whence the municipal water is also derived. Approximately, 65% of the annual precipitation falls between April and September, coinciding with the growing season for most crops. Most of the summer precipitation is from thunderstorms which occur about 60 days per year.
The National Climatic Center also reports that the average winter temperature at Canon City is 38 degrees F (3 degrees C), and the average daily minimum temperature in winter is 24 degrees F (-4 degrees C). In summer, the average temperature is 73 degrees F (23 degrees C) at Canon City, while the average daily maximum temperature is 84 degrees F (29 degrees C). The average frost free season is 167 days.

Canon City averages about 37 inches (94 cm) of snowfall during winter, but on average, there is greater than one inch of snow on the ground for only two days per year.

Average relative humidity is 60% at dawn, dropping to about 40% by midaftemoon. The percentage of possible sunshine is 80% in summer and 75% in winter. Prevailing winds are from the north-northeast, with the highest average windspeed of 12 miles per hour (19 km/hr) occurring in the spring.
CHAPTER II

GEOLOGICAL SYNOPSIS OF THE STUDY AREA

Geologic Setting

The Royal Gorge arch is one of three large, basement cored uplifts at the southern end of the Front Range, Colorado (Figure 3). All of these uplifts plunge to the south, and each contains well-exposed basement and sedimentary cover rocks. The Royal Gorge arch lies at the juncture of the southeast-plunging Wet Mountains to the south, and the continuation of the south-plunging Front Range to the north.

The basement core of the Royal Gorge arch has been exposed to multiple deformation events. Sorting out the resultant basement features and fabrics will provide information about the structural evolution, and also assist in defining ground water movement through the present aquifer.

Geologic History

Precambrian History

The Precambrian rocks of Colorado have been described by Tweto (1977) and (1980b) as a gneissic complex (dated at about 1800 Ma), three ages of granitic rocks (dated at 1700, 1400, and 1000 Ma), and three sedimentary or sedimentary and volcanic units (one dated between 1720-1460 Ma, one dated at 1400 Ma, and one undated, but assumed to be 1000 Ma).

Braddock (1970) and Taylor (1975) described three episodes of folding in
Proterozoic rocks of the Front Range. Foliation is thought to have developed during the first episode, and was subsequently modified by the later events (Tweto, 1980b). A second episode caused refolding of the first folds in the central Front Range (Taylor, 1975). The second episode was also accompanied by the emplacement of granitic intrusives (Tweto, 1980b). The third episode of folding caused formation of shear...
zones during, and following, the emplacement of additional granitic intrusives (Tweto and Sims, 1963; Abbott, 1972; and Taylor, 1975).

Numerous 1700 Ma age granitic plutons in the Front Range and Wet Mountains suggest that the overall basement complex in this part of Colorado is at least 1700 Ma old or older (Bickford, 1988). Plutons of this age were also found in the Arkansas River Canyon (the Royal Gorge arch) at the northern end of the Wet Mountains (Bickford and others, 1984). Bickford (1988) also notes that major pegmatite swarms in the northern Wet Mountains and other parts of Colorado all have a similar age of 1450 Ma. These pegmatites were also intruded into the 1700 Ma basement rock near the Royal Gorge arch.

Phanerozoic History

There were three major tectonic episodes in south-central Colorado during the Phanerozoic (Tweto, 1980a): (1) uplift of the Ancestral Rockies in the late Paleozoic, (2) development of uplifts and basins during the late Mesozoic-early Cenozoic Laramide orogeny, and (3) Neogene (Miocene and Pliocene) block faulting associated with the Rio Grande rift. Gerhard (1967) also speculated on tectonic activity prior to Early Ordovician time, which may have removed sedimentary rocks of Cambrian age (Sawatch Quartzite) in the study area.

Ancestral Rocky Mountains

In late Paleozoic time, large uplifts developed in Colorado, as part of the Ancestral Rocky Mountains (Tweto, 1980a). The uplifts were interpreted to be part fault-bounded and part upwarps of the crystalline basement. Tweto (1980a)
believed the uplifts had the profiles of broad cuestas, with the steep side to the southwest and only upwarped at a moderate angle to the east. De Voto (1980) reports that during the Pennsylvanian Period alone, mountain building in Colorado resulted in ranges with up to 10,000 feet (3000 m) of relief. The late Pennsylvanian landscape was interpreted to have contained uplifted mountainous areas roughly coincident with many of the present-day (Laramide) uplifts, including the Front Range (Royal Gorge arch) and the Wet Mountains (south of the Royal Gorge arch) (De Voto, 1980).

The differential preservation and extent of many of the early Paleozoic sedimentary units on the flanks of the Royal Gorge arch are believed to be caused by the uplifted Ancestral Rocky Mountains. Mississippian carbonates, that were thought to have covered much of Colorado (Tweto, 1980a), were totally removed from the southern Front Range and the Royal Gorge arch. Figure 2 shows a considerable difference in the preserved stratigraphic column from the east flank to the west flank of the Royal Gorge arch. This difference is interpreted to be a result of partial uplift of the arch during the Pennsylvanian Period (Ancestral Rocky Mountains). The east flank has a considerable section of Ordovician and Permian-Pennsylvanian age rocks preserved beneath the Jurassic section. The west side, by contrast, has Jurassic age rocks in direct contact with the basement, with no Permian-Pennsylvanian rocks, and only very limited Ordovician rocks to the south (Figure 2).

**Laramide Orogeny**

The Laramide orogeny (Late Cretaceous to Early Eocene) is responsible for most of the present mountainous regions of Colorado, including a major episode of uplift at the Royal Gorge arch. Tweto (1980a) reports that many uplifts were
rejuvenations of earlier Precambrian and/or Paleozoic (Ancestral Rocky Mountains) tectonic features, as appears to be the case at the Royal Gorge arch. Faulting and folding of the Cretaceous section on both flanks of the arch is a direct result of the Laramide orogeny. Other uplifts in Colorado were newly created during this orogeny in what were the sediment-filled troughs between uplifts of the Ancestral Rocky Mountains.

**Neogene Tectonism**

During the Miocene and Pliocene, many of the uplifts in Colorado were rejuvenated again by an episode of extensional block-faulting that extended through the "mountainous heart of the Laramide uplift" (Tweto, 1979). Basin-and-range style faulting, associated with the Rio Grande rift, followed a wide belt trending north-northwest across Colorado (Tweto, 1980a).

The Rio Grande rift is a graben system that extends from New Mexico through the upper Arkansas River valley to near Leadville, Colorado. Additional, but narrower rifting extends almost to the Wyoming border (Tweto, 1980a). The rift system passes about 25 km to the west of Royal Gorge arch and immediately west of the Wet Mountains.

Tweto (1980a) states that many Colorado mountainous regions, including the Wet Mountains, were reelevated as horsts during the Neogene tectonism. He notes that both regional uplift and differential uplift accompanied the block-faulting, with some Laramide uplifts being elevated at least 600 m. Epis and others (1980) report that displacements due to block-faulting (both uplift and downdrop) varied between 1000 m and 6000 m across Colorado. Oligocene rocks south of Canon City were vertically offset by as much as 1080 m (Epis and others, 1980).
Precambrian and Phanerozoic Lithologies

Precambrian Lithologies of the Royal Gorge Arch

Nearly 400 Precambrian outcrops were examined within the 60 km² exposure area of the Royal Gorge arch. A total of over 100 rock specimens were collected, from which 31 petrographic thin-sections were prepared and analyzed, to assist in the lithologic analysis. Five thin-sections were point-counted to confirm the estimated mineral percentages. Appendix A lists the formal rock names and percentages of minerals identified in each thin-section, including which thin-sections were point-counted.

Two basic lithologies appear to dominate the Precambrian core of the uplift (see Figure 2). A well-foliated, quartzofeldspathic gneiss makes up most of the southern two-thirds of the basement core, whereas, a weakly-foliated granodiorite makes up most of the basement to the north. Rb-Sr dates of 1720 Ma have been reported by Taylor and others (1975) for Precambrian rocks from Royal Gorge arch (Carl Hedge, written communication). This age is consistent with the research by Bickford (1988) and Bickford and others (1984) for rocks in the southern Front Range and Wet Mountains.

Quartzofeldspathic Gneiss

This lithology consists of a well-foliated, quartzofeldspathic, and commonly migmatitic gneiss, with alternating layers of plagioclase-microcline-quartz gneiss (as sample #8, Appendix A) and plagioclase-biotite-quartz gneiss (as sample #300, Appendix A). Local sillimanite and garnet indicate that the unit reached the amphibolite metamorphic facies. Rocks from this unit are typically pink to light brown.
or gray-tan, medium-grained, hypidiomorphic, granular, with a low color index (less than 15). The plagioclase An% is commonly in the low 20s. Foliation is strongly developed by alignment of biotite, muscovite, and even quartzofeldspathic layers.

Thin bands (less than 5 meters) of highly schistose layers are found locally, and are concordant with the surrounding foliation of the gneiss (Figure 2). These layers are typically light red-brown to light purple in color, rich in muscovite, biotite, and quartz, with varying amounts of sillimanite (as sample #7, Appendix A).

Local veins of calcite were identified within fractures on four of the thin-sections from the quartzofeldspathic gneiss (samples #4, #5, #13, and #24, Appendix A). Clay minerals within fractures, presumably from silicate alteration, were also observed on two thin-sections (samples #5, and #7, Appendix A). These observations indicate that ground water has altered the mineralogy locally, causing dissolution and reprecipitation along fractures within the quartzofeldspathic gneiss.

The quartzofeldspathic gneiss is interpreted by Taylor and others (1975) as forming from a mixed sedimentary and volcanic precursor, containing flows, tuffs, and sedimentary interlayers. Based on the above radiometric-age data, the unit is thought to be at least 1700 Ma.

Granodiorite

This unit typically consists of a poorly-foliated, quartz-biotite-plagioclase (oligoclase-andesine) granodiorite (as sample #28, Appendix A). Rocks from this unit are often green-gray to dark pink to brown, medium to coarse-grained, porphyritic (phenocrysts of microcline 1 cm in length are common), with a variable color index (20-40). The plagioclase An% is generally in the mid-30s, but shows more variation than the quartzofeldspathic gneiss. Foliation is best developed in the southern part of
the granodiorite unit, where it contacts the quartzofeldspathic gneiss. North of this contact, the foliation diminishes rapidly and becomes weakly to non-foliated over a distance of 1-2 km. When foliation is developed, it is marked by the alignment of biotite, hornblende, and other platy mafic minerals.

In the extreme northwest part of the study area (Figure 2), the granodiorite contains several, small (less than 0.5 km in diameter) "intrusives" of poorly-foliated granite (as sample #31, Appendix A). These granites are interpreted to be associated with a local increase in the number of granitic pegmatites (Figure 2), and may be related to their intrusion. They are typically pink, medium-grained, often porphyritic (microcline phenocrysts), rich in quartz and microcline, with color indices less than 20.

Bands of thin (less than 5 meters), highly schistose layers are spread along the southern contact of the granodiorite (Figure 2). The schists appear to be concordant to the granodiorite foliation, and vary from a dark gray, plagioclase-biotite-rich variety (as sample #23, Appendix A) to a blue-gray, muscovite-quartz-rich variety (as sample #26, Appendix A).

Taylor and others (1975) interpret the granodiorite to be equivalent to the Boulder Creek Granodiorite (1700 Ma). They also note that the unit is more foliated at the margins, generally concordant to the enclosing gneisses, and can become almost massive in the interior of the plutons.

Other Precambrian Lithologies

Microcline and quartz-rich pegmatites are locally abundant throughout the basement complex of Royal Gorge arch (Figure 2). Most of these pegmatites are fairly concordant with the foliation of the gneiss and/or granodiorite. They are pink in color, very fractured (almost randomly), with very low color indices (less than 5).
Most of the pegmatites are less than 10 m thick, but can be up to 50 m thick. The greatest abundance of pegmatite swarms occurs near the granodiorite-gneiss contact and in the extreme northwest part of the granodiorite lithology (Figure 2). The pegmatites are interpreted to be equivalent to the 1450 Ma pegmatite swarms discussed by Bickford (1988).

A small, isolated exposure of tonalite (less than 0.25 km diameter) was identified at the granodiorite-gneiss contact with the Arkansas River (Figure 2). The tonalite is olive-green, medium-grained to porphyritic, rich in chlorite-quartz-plagioclase, having a color index in the 20's (as sample #22, Appendix A). The tonalite was not observed elsewhere along the mapped lithologic contact.

Several minor exposures of hornblendite were also identified within the granodiorite unit. The hornblendite is typically green-gray to dark gray, medium-grained to porphyritic, rich in hornblende and orthoclase, with a color index of about 60 (as sample #19, Appendix A). Foliation was locally, well-developed in this lithology.

A single, limited exposure of alkali feldspar granite was mapped near the top of YMCA Mountain, within the quartzofeldspathic gneiss (Figure 2). This lithology is pink, fine-grained, granular, rich in biotite, quartz, and microcline, with a color index of about 15 (as sample #15, Appendix A).

Several thin (less than 5 m) outcrops of mafic-rich, hornblende-amphibolite were observed within the quartzofeldspathic gneiss. The amphibolite is typically dark-green, rich in quartz-plagioclase-hornblende (about 60% hornblende), containing abundant porphyroblasts of hornblende up to 3 mm in length (as sample #6, Appendix A). This lithology appears to be concordant with the foliation in the granitic gneiss,
Several thin (less than 5 m) outcrops of mafic-rich, hornblende-amphibolite were observed within the quartzofeldspathic gneiss. The amphibolite is typically dark-green, rich in quartz-plagioclase-hornblende (about 60% hornblende), containing abundant porphyroblasts of hornblende up to 3 mm in length (as sample #6, Appendix A). This lithology appears to be concordant with the foliation in the granitic gneiss, and is probably related to concentrations of mafic-rich, migmatitic zones within the gneiss.

Taylor and others (1975) mapped a 0.5-1.0 km zone of quartz diorite between the granodiorite unit and quartzofeldspathic gneiss unit. This lithology was not observed along the contact zone, except as represented by the single exposure of tonalite discussed above. Because Taylor and others (1975) named this unit prior to the popularization of the present IUGS (1973) system, they called the lithology quartz diorite as opposed to tonalite. However, the extent of the lithology appears to have been extrapolated along the contact on the basis of inference and not direct field evidence. Taylor and others (1975) noted in their lithologic description that quartz diorite is commonly found as a mafic shell along the outer margins of granodiorite plutons. Because they observed this relationship elsewhere, they interpreted a "margin" of quartz diorite along the edge of the granodiorite at the Royal Gorge arch.

Phanerozoic Lithologies of the Royal Gorge Arch

The Phanerozoic rocks around Royal Gorge arch have been mapped and discussed by Mann (1957), Scott (1977), Scott and others (1978), and Taylor and others (1975). The oldest Phanerozoic rocks are interpreted by Taylor and others (1975) to be Cambrian-age, gabbro dikes. These mafic intrusions are thought to represent extensions of the the McClure Mountain-Democrat Creek igneous complex
throughout the basement core of the uplift, usually discordantly, and often follow near-horizontal joints in the basement. Taylor and others (1975) show only the mapped extent of three of these dikes, although they are far more common. The dikes are usually less than 10 m thick, but can be up to 25 m thick, as are the dikes mapped by Taylor and others (1975). Offsets of the dikes were used as field evidence of faulting and shearing in the basement.

The Paleozoic rocks of the Royal Gorge arch (Table 1) consist of several Ordovician sedimentary units (Manitou Limestone, Harding Sandstone, and Fremont Dolomite) and the Pennsylvanian and Permian Fountain Formation. These rock units are limited to the eastern and southern flanks of the Royal Gorge arch because of differential erosion associated with the Ancestral Rocky Mountains. The Manitou Limestone is in depositional contact with the crystalline basement in only the extreme northeast, whereas the Harding Sandstone rests directly on Precambrian rocks in both the east and south. The Fremont Dolomite is confined to the eastern flank of the Royal Gorge arch, and always in depositional contact with the underlying Harding Sandstone.

The Pennsylvanian to Permian Fountain Formation is irregularly distributed on the eastern flank of the Royal Gorge arch, similar to the Fremont Dolomite. In contrast however, the thickness varies considerably (0-606 m), generally becoming thicker to the north.

Although observed elsewhere in the southern Front Range, the Permian to Triassic-age Lykins Formation and the Mississippian-age Williams Canyon Limestone were not mapped at the Royal Gorge arch, and were not included in the stratigraphic column (Table 1). Both of these formations are locally quite thin (usually less than 10
<table>
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<tr>
<th>Age</th>
<th>Map Unit</th>
<th>Rock Unit</th>
<th>Description</th>
<th>Thickness (m)</th>
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<tr>
<td>Paleocene</td>
<td>Tpc</td>
<td>Poison Canyon Fm.</td>
<td>Lower: coarse sandstone and conglomerate. Upper: interlayered shale, siltstone, and sandstone.</td>
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<tr>
<td>Paleocene-Cretaceous</td>
<td>TKr</td>
<td>Raton Fm.</td>
<td>Massive sandstone.</td>
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<td>Cretaceous</td>
<td>Kvt</td>
<td>Vermejo Fm.</td>
<td>Sandstone interlayered with shale and coal. Thin-bedded to massive sandstone.</td>
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<tr>
<td></td>
<td></td>
<td>Trinidad Ss.</td>
<td>Thinn-bedded to massive sandstone.</td>
<td>12-27</td>
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<tr>
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<td>Niobrara Fm.</td>
<td>Calcareous shale and chalk.</td>
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<td></td>
<td></td>
<td>Smoky Hill Sh.</td>
<td>Thick-bedded limestone.</td>
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<td>Fort Hays Ls.</td>
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<td>Carlile Sh.</td>
<td>Shale, chalky shale, and Codell Ss.</td>
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<td>Greenhorn Ls.</td>
<td>Massive limestone with shale partings.</td>
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<td>Graneros Sh.</td>
<td>Hard shale with local limestone.</td>
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<tr>
<td>Cretaceous</td>
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<td>Dakota Ss.</td>
<td>Fine-grained, cross-bedded sandstone.</td>
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<td></td>
<td>Purgatorioire Fm.</td>
<td>Massive sandstone and interlayered shale and sandstone.</td>
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Table 1-Continued

<table>
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<tr>
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<th>Rock Unit</th>
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<th>Thickness (m)</th>
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<td>Morrison Fm.</td>
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<td>Ralston Creek Fm.</td>
<td>Conglomerate, gypsum, siltstone, sandstone, and limestone.</td>
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<td>Pennsylvanian PPf and Permian</td>
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<td>Fountain Fm.</td>
<td>Red arkose, conglomerate, and sandstone with local shale and limestone.</td>
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<td>Ordovician</td>
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<td>Harding Ss.</td>
<td>Fine-grained sandstone with interlayered shale.</td>
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<td>Om</td>
<td>Manitou Ls.</td>
<td>Interlayered limestone and cherty dolomite.</td>
<td>5</td>
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* Modified from R. B. Chase (personal communication, 1995), Scott (1977), and Scott and others (1975).
where observed. However, at Twin Mountain, immediately northeast of the Royal Gorge arch (Figure 2), the Williams Canyon Limestone becomes 15 m thick.

The Jurassic Morrison and Ralston Creek Formations are in depositional contact with the crystalline basement along the western flank and crest of the Royal Gorge arch. As previously discussed, the stratigraphic column is quite variable across the uplift, reflecting the influence of at least the Ancestral Rocky Mountains deformational event. The presence of a few, fairly flat-lying, remnants of Jurassic rocks along the crest of the Royal Gorge arch (Figure 2) imply that these units are the oldest sediments to have covered the entire uplift prior to the Laramide deformational event.

The Cretaceous Dakota Sandstone and Purgatoire Formation together form a continuous hogback on the eastern flank of the arch and a cliff on the more gently-dipping western flank of the arch. Where observed, the Dakota-Purgatoire Formation is always in depositional contact with the underlying Morrison Formation.

Structural Features and Distributions

Foliation Patterns

Foliation patterns in the Precambrian rocks are shown on Figure 2. Foliations in the northern part of the Royal Gorge arch (in the granodiorite) are generally steep or near vertical, and trend fairly consistently northeast-southwest. The degree of foliation development in the granodiorite is varied. In the northern part of the granodiorite, foliations are weakly developed or not developed, whereas to the south, foliations are commonly well developed.
In the southern part of the Royal Gorge arch (in the quartzofeldspathic gneiss), foliations are uniformly well developed. The foliation pattern in the quartzofeldspathic gneiss suggests a broad antiform, with foliations trending northeast-southwest on the western flank of the arch and northwest-southeast on the eastern flank of the arch. This pattern is also illustrated by the strikes of the schistose layers within the quartzofeldspathic gneiss (Figure 2). The dip angle of foliations in the quartzofeldspathic gneiss is also locally more variable than the granodiorite.

**Fold Patterns**

The Royal Gorge arch is a broad, asymmetric, anticlinal fold, with sedimentary rocks dipping steeply on the eastern flank and gently on the western flank. Smaller folds in the overlying sedimentary units are developed on the western flank of the arch (Webster Park graben) and along the northwestern margin of the arch.

Although the Royal Gorge arch plunges to the south, it appears slightly warped about an east-west axis in the vicinity of the Gorge fault (Figure 2). Part of the reason for this apparent "warping" is that the overall folding of the arch is broader north of the Gorge fault and tighter south of the fault.

**Faults**

The mapped extent of the Royal Gorge arch is bounded by the Wet Mountains fault system to the south and the Mikesell Gulch fault to the north (Figure 2). Both of these faults are fairly steep and extensive, and delineate the arch either structurally and/or topographically from adjacent geologic features.
The steeply-dipping Gorge fault (including one splay) trends nearly east-west, offsetting both the sedimentary units and the Precambrian basement on the western flank of the arch. A similar, but smaller, unnamed fault about 2 km south of the Gorge fault also trends east-west. These two faults appear to lose displacement to the east, until no offset is observed along the crest of the arch. Both faults also appear to follow the strike of adjacent Precambrian foliations.

An unnamed, northwest-southeast-trending fault connects the Gorge fault to the Wet Mountains fault system. This fault dips steeply to the east and shows reverse motion.

Several small, northeast-southwest-trending faults are observed on the northeastern margin of the arch. These faults also parallel the strike of local foliations and offset the overlying sedimentary units.

Joints

Joints in the Precambrian basement of the Royal Gorge arch are commonly near-vertical and prevalent. Joints are more closely-spaced in the quartzofeldspathic gneiss (usually less than 1 m apart), as opposed to the granodiorite (often greater than 1 m apart). The density of joint sets also appears to increase near faults, for example, the Mikesell Gulch fault.

Joint patterns appear to follow foliation patterns in the Precambrian rocks and bedding patterns in the sedimentary units. Joints sets are commonly developed both parallel and perpendicular to foliation and/or bedding orientation.
CHAPTER III

STRUCTURAL ANALYSIS

Historical Background

Schmidt and others (1993) determined that most Rocky Mountain basement-cored folds could be placed into two end-member styles of folding, termed mode 1 and mode 2. Mode 1 structures have basement deformation confined to a narrow zone of cataclasis adjacent to a single fault, cover rocks that thin on the forelimb, and an interlimb angle of 60 degrees or less (the angle between the cover rocks on the backlimb and the overturned forelimb). Mode 2 structures, by contrast, have basement deformation spread over a broad zone between the principal fault and the anticlinal hinge surface, cover rocks that do not thin appreciably, interlimb angles greater than 90 degrees, and common backthrusts.

The style of deformation was found to depend partly on the pre-existing basement fabric (Schmidt and others, 1993). Well-foliated basement rocks that have foliation either subparallel to sedimentary bedding or subparallel to the axial surfaces tend to produce mode 2 folds (also Chase and others, 1993). Basement foliation, however, was found to have little affect on cover-rock geometries when the foliation is at an angle that is ≥ 15 degrees to the principal fault or to the axial surfaces (Chase and others, 1993). Both Chase and others (1993) and Schmidt and others (1993) recognized that the influence of a basement foliation can be reduced if pre-existing faults have been reactivated during development of the foreland uplift. In effect, the
reactivated faults can become the dominant fabric if oriented subparallel to the axial surfaces and/or principal fault.

Structural Analysis of the Royal Gorge Arch

Introduction and Philosophy

During the 1993 field season, 565 separate rock exposures were examined in detail over the 60 km² area of the Royal Gorge arch. Outcrops were studied in both the Precambrian, crystalline core of the uplift, and in the overlying Phanerozoic, sedimentary cover. Various indicators of brittle deformation were recorded, including 2959 joint orientations, 429 basement foliations, 94 sedimentary bedding orientations, and 215 slickenside measurements. At each location, the attitude of the bedding or foliation surface was determined, joint orientations measured, joint spacing (both horizontally and vertically) approximated, and slickenside surfaces and slickenlines recorded (if present).

The following discussion details the methodology and results of the domain analysis, joint analysis, and slickenside analysis of the Royal Gorge arch.

Domain Analysis

Foliations and bedding orientations were subdivided into homogeneous, cylindrical domains as described by Ramsay and Huber (1987). All orientation data were plotted as poles on π-diagrams and Kamb contoured using equal area stereographic projections to plot the perpendicular (poles) to foliations and bedding (Allmendinger, 1992). The domains were subdivided by the degree of homogeneity of the plots as described below. Domain boundaries were determined by first plotting
poles to foliations (or bedding) from a specific and small area of the arch, defining either a great circle pattern (if cylindrical) or a point concentration. Poles to foliations (or bedding) were then plotted from a greater and greater area until the great circle distribution or clustering of poles in the pattern was observed to change or deviate. At this point, a tentative domain boundary was drawn on the map and the "deviant" points replotted on a new stereographic projection. This process of trial-and-error continued until all domains were defined.

Results of the stereographic analysis of the fabric data show 10 homogeneous domains, consisting of three sedimentary domains and seven basement domains (Figures 4-6). The areal distribution of the domain boundaries is shown in Figure 7, and includes reduced, stereographic projections of the poles to foliation or poles to bedding for each domain. Nearly all of the domains, except Domains IX and X, appear to be point concentrations, making the determination of great circles through the data unnecessary.

The sedimentary domains are subdivided by the axial trace of the Royal Gorge arch. Domain I (Figure 4) shows fairly flat-lying sedimentary units down the arch axis, both at the northern and southern ends (Figure 7). Domains II and III (Figure 4) represent the west and east flanks of the Royal Gorge arch, respectively.

A synoptic diagram (Figure 8) was constructed by first plotting the point concentration of each sedimentary domain (I, II, and III) and then calculating the pole to the best fit great circle. The synoptic diagram of the three sedimentary domains defines a broad fold with axis about 5° S 7 E (173 Azimuth), indicating that the Royal Gorge arch has a slight plunge to the south-southeast.

The basement (Figures 5 and 6) was subdivided into seven homogeneous domains, reflecting a response to both lithologic and fabric differences across the
Figure 4. Equal Area Stereographic Projections of Poles to Sedimentary Bedding for Domains I, II, and III, Kamb Contoured.
Figure 5. Equal Area Stereographic Projections of Poles to Basement Foliation for Domains IV, V, and VI, Kamb Contoured.
Figure 6. Equal Area Stereographic Projections of Poles to Basement Foliation for Domains VII, VIII, IX, and X, Kamb Contoured.
Royal Gorge arch. The lithologic contact between granodiorite to the north and quartzofeldspathic gneiss to the south closely coincides with the boundaries between Domains VI and V, and Domains VI and IV (Figure 7). This observation indicates
that lithology differences directly influence the deformation of basement rocks in the Royal Gorge arch. More subtle variations in lithology may extend to Domain VI, as this domain also parallels the contact between quartzofeldspathic gneiss and granodiorite.

**Joint Analysis**

Joint orientation and approximate horizontal and vertical joint spacings were recorded at each field station (see Figure 2). Jointing was common throughout the study area, but appeared to vary by lithology. Joints tended to be aligned both parallel and perpendicular to foliation in the quartzofeldspathic gneiss and in the foliated areas of the granodiorite. Granitic pegmatites were more highly jointed and randomly fractured due to the brittle nature of the large silicate minerals. The sedimentary units commonly exhibited jointing that was both parallel and perpendicular to bedding surfaces, similar to the aforementioned relationship to foliation. Joint orientations and spacings were found to be quite consistent over the 360 m of vertical relief exposed in the canyon of the Royal Gorge.

Poles to joints for each of the aforementioned structural domains were plotted on $\pi$-diagrams using equal area stereographic projections. Figures 9-13 show the Kamb contoured results in combination with rose diagrams for each domain. Reduced, stereographic projections of the joint data are also shown in Figure 7.

Joint sets appear to be strongly influenced by foliation and bedding orientation (fabric) from domain to domain. Nearly all domains (Figures 4-6 and 9-13) show joint concentrations that are coincident with the respective fabric $\pi$-diagrams. Joint sets and fabric orientations in respective domains also commonly exhibit the same direction and magnitude of orientation variation with respect to other domains. For example,
Figure 9. Kamb-contoured Stereographic Projections of Poles to Joints and Rose Diagrams for Joints in Domains I and II.
Figure 10. Kamb-contoured Stereographic Projections of Poles to Joints and Rose Diagrams for Joints in Domains III and IV.
Figure 11. Kamb-contoured Stereographic Projections of Poles to Joints and Rose Diagrams for Joints in Domains V and VI.
Figure 12. Kamb-contoured Stereographic Projections of Poles to Joints and Rose Diagrams for Joints in Domains VII and VIII.
Figure 13. Kamb-contoured Stereographic Projections of Poles to Joints and Rose Diagrams for Joints in Domains IX and X.
Domain IV and Domain V show about a 20-degree counterclockwise difference in both foliation and joint patterns going from Domain IV to Domain V (Figures 5, 10, and 11). Joint pattern variations between domains not only reflect the influence of fabric, but independently reinforce the choice of domain boundaries based on fabric by confirming the presence of joint differences from domain to domain.

Fracture densities across the Royal Gorge arch were determined by multiplying together the average, measured, horizontal and vertical joint spacings at each field station. The resultant single values (in square feet), and contoured in Figure 14, are inversely proportional to the fracture density. In other words, the smaller the multiplied value, the greater the fracture density. For example, a fracture spacing of 2 ft (0.6 m) between horizontal joints and a fracture spacing of 3 ft (0.9 m) between vertical joints yields a value of 6 ft\(^2\) (0.56 m\(^2\)). Whereas, a fracture spacing of 0.5 ft (0.15 m) between horizontal joints and a fracture spacing of 1.0 ft (0.3 m) between vertical joints yields a value of 0.5 ft\(^2\) (0.045 m\(^2\)), with the latter value having a higher fracture density. The values in Figure 14 are plotted in square feet, as opposed to square meters, to avoid small decimal values.

The most noticeable difference in fracture density values across the Royal Gorge arch coincides with the major lithologic boundary within the basement rocks (Figure 2). Densities within the granodiorite are consistently greater than 1.0 ft\(^2\), whereas densities within the quartzofeldspathic gneiss to the south show variations between 0.1 ft\(^2\) and greater than 1.0 ft\(^2\) (Figure 14).

Secondary trends in the fracture density map include common increases in density around fault zones (the Mikesell Gulch fault), and a general tendency of the density trends to follow foliation patterns (compare Figures 2 and 14). This tendency
Increasing Fracture Density

- Greater than 1.0 ft sq
- 0.999 to 0.5 ft sq
- 0.499 to 0.1 ft sq
- Less than 0.1 ft sq

Legend: Fracture Density Values were determined by multiplying together the average, measured, horizontal and vertical joint spacings.

Figure 14. Fracture Density Map of the Surface-exposed, Basement Rocks in the Royal Gorge Arch. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
implies that fracture spacing is also strongly influenced by basement fabric orientation and not just lithology differences.

**Slickenside Analysis**

Slickenside surfaces are commonly marked by lineations on the surface, called slickenlines (Fleuty, 1975). These lineations are termed kinematic indicators because they indicate the relative sense of movement of the opposing, rock surfaces (Marshak and Mitra, 1988).

Most of the observed slickenlines in the study area were what Marshak and Mitra (1988) call groove lineations. These features form simply by opposing surfaces scratching against one another during movement. The resulting lineations are usually ambiguous as indicators of the sense of offset, but they do indicate the direction of shear. The basement rocks in the Royal Gorge arch commonly exhibited some type of groove lineation.

The second most common type of slickenline found at the Royal Gorge arch are fiber lineations (Marshak and Mitra, 1988). These lineations form by incremental vein-mineral precipitation along the fault surface, resulting in long thin mineral fibers elongated in the direction of motion. Imbrication of the fibrous minerals indicates the sense of movement of the fault. Within the study area, fiber lineations were most commonly observed in the sedimentary units, and were usually composed of calcite.

Several methods to analyze slickensides and/or faults have been described in the literature. Anderson (1951) described a method using conjugate fault populations in isotropic rocks to determine the principal stress directions at the time of faulting. However, because the basement rocks within the Royal Gorge arch are not isotropic, and many of the observed slickenlines do not indicate the sense of movement,
Anderson's method was not used in the present study. Anderson's method also assumes that the same tectonic event induces both formation of faults and movement along fault surfaces. This assumption could not be made for the present study as several deformation events have been documented, likely causing multiple movements of early-formed fractures.

There have been two other graphical methods developed to analyze slickenside arrays. Both methods assume that: (1) any planar discontinuity in a rock may be activated or reactivated by tectonic stress, regardless of its origin; and (2) all faults move independently but consistently within the same stress field.

The first method, described by Angelier (1979), requires knowledge of the movement sense on each slickenside surface. His method attempts to determine the mean stress tensor, knowing the orientations and senses of slip on numerous faults. Angelier noted that families of conjugate faults should have slickenlines perpendicular to the intersection of fault planes, and the senses of motion should be symmetrically distributed so that the acute angle between faults contains the direction of shortening. By plotting fault planes, slickenlines and senses of motion on a stereographic projection, the compression (P) and tension (T) axes can be determined, and respective principal stress axes inferred.

The second method, first described by Arthaud (1969) and modified by Aleksandrowski (1985), is based on the concept of "movement planes," and only requires knowledge of the sense of movement on a few slickenside surfaces to be employed successfully. Because most of the slickenside surfaces at the Royal Gorge arch did not indicate the sense of movement, the method of Aleksandrowski (1985) was utilized for the present analysis.
The first step in the Aleksandrowski method is to plot the poles to the slickenside surfaces on an equal area stereographic projection. Next, the slickenlines are plotted, and the corresponding poles and slickenlines are connected with a great circle, thus defining the movement plane of the fault (M-plane). All available movement directions are then plotted on the M-planes with arrows indicative of the hanging wall direction of movement at the slickenline location. The final step is to plot the poles (πM-planes) to the movement planes on a separate stereographic projection. Because the method is essentially a statistical analysis, the higher the number of slickenside measurements, the higher the confidence of the interpretation.

According to the method, intersection points of the M-planes should define the axes of the direction of maximum shortening and maximum extension of the rock mass. Arrows indicative of the sense of movement should point away from the Sigma 1 direction, and point toward the Sigma 3 direction. Concentrations of πM-planes should cluster about Sigma 2, the intermediate principal stress direction. Therefore, successful employment of the method on a slickenside population allows the determination of all three principal stress directions simultaneously.

The results of the use of the Aleksandrowski method on populations of slickensides across the Royal Gorge arch are shown in Figures 15-27. In each figure, the "1," "2," and "3" represent the Sigma 1, Sigma 2, and Sigma 3 principal stress directions, respectively. Reduced, stereographic projections of the M-planes and the M-plane intersections are also shown in Figure 7. Whenever determinations of principal stress directions were ambiguous (for example, Domain X, Figure 27), a question mark was attached to the interpreted stress number or numbers. Slickenside populations where plotted on a domain-by-domain basis, assuming the stresses acting on each homogeneous domain were approximately uniform. The assumption of
Figure 15. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain I, Following Aleksandrowski's Method.
Figure 16. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain II (North), Following Aleksandrowski's Method.
Figure 17. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain II (South), Following Aleksandrowski's Method.
Figure 18. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain III (North), Following Aleksandrowski's Method.
Figure 19. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain III (South), Following Aleksandrowski’s Method.
Figure 20. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain IV, Following Aleksandrowski’s Method.
Figure 21. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain V, Following Aleksandrowski's Method.
Movement Planes  
M-Plane Intersections Kamb

Poles to M-Planes Kamb
Domain VI - Slickensides

Figure 22. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain VI, Following Aleksandrowski's Method.
Figure 23. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain VII (North), Following Aleksandrowski's Method.
Figure 24. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain VII (South), Following Aleksandrowski's Method.
Figure 25. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain VIII, Following Aleksandrowski's Method.
Figure 26. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain IX, Following Aleksandrowski’s Method.
Figure 27. Slickenside M-Plane Diagrams Used in the Interpretation of Principal Stress Axes for Domain X, Following Aleksandrowski's Method.
stresses being uniform across a particular domain appears to have been valid, as the stress patterns are fairly well defined in most cases, and the directions of principal stresses change from domain to domain. In the three situations where stress patterns seemed to shift within a domain, the domain was arbitrarily subdivided (Domains II, III, and VII, Figures 16-19, 23, and 24) and stress directions determined separately.

Stress determinations were ambiguous in several instances, especially when arrows indicative of the sense of movement were pointing in multiple directions (for example, Domains VII, IX and X). Domain X was the most ambiguous as it is segmented by numerous faults (Figure 7), and may represent more than one stress system. Overall, however, use of the method of Aleksandrowski appeared to work fairly well for the slickenside populations at the Royal Gorge arch.

Slickensides populations were consistent across large parts of the sedimentary domains even though many measurements were taken on several different formations. The majority of the slickensides were observed in the Cretaceous Dakota, Jurassic Morrison, Ordovician Fremont, and Ordovician Harding Formations. The consistency of these results implies a single, unique stress system was in operation across most of the Royal Gorge arch during the Laramide orogeny. This stress system was responsible for the final offset of the rocks, creating the observed slickenside surfaces in the sedimentary units.

In order to cross-check the results of the stress field interpretations, a comparison was made between the method of Aleksandrowski (1985) and the method of Angelier (1979) where sufficient senses of movement were known. Figures 28 and 29 show the results of these comparisons for Domains I and IV, respectively. In both cases, the two methods show very similar principal stress directions, independently confirming that the Aleksandrowski method provided an adequate paleostress analysis.
Figure 28. Comparison of Aleksandrowski's and Angelier's Methods for Domain I. Showing Directions of Sigma 1, Sigma 2, and Sigma 3 for Each Method.
Figure 29. Comparison of Aleksandrowski's and Angelier's Methods for Domain IV showing directions of Sigma 1, Sigma 2, and Sigma 3 for each method.
The interpreted stress directions for each domain (and subdomain) were projected to horizontal and vertical, eliminating any inclined stresses, and the horizontal stress directions plotted as a "stress axes map" (Figure 30). The Arabic numbers on Figure 30 refer to projections of Sigma 1, Sigma 2, and Sigma 3, respectively. Where the stress directions shifted within a domain, both directions were shown. In some cases, the ambiguous nature of the interpretation prevented plotting of horizontal stress directions (Domain X).

A stress trajectories map (Figure 31) was then constructed by connecting the maximum and minimum horizontal stress directions from the stress axes map. The maximum stress trajectories are represented by solid lines and the minimum stress trajectories are represented by dashed lines on Figure 31.

The maximum horizontal stress direction within the basement rocks in the northern part of the Royal Gorge arch appears to trend N 60-70 E (Figure 31). This direction compares quite favorably with the N 70 E direction of maximum compression found by Jurista and others (1995) for the Webster Park graben, immediately west of the Royal Gorge arch (Figure 2). Jurista and others (1995) also found this same stress field to have been distributed uniformly across much of the Rocky Mountain front in Colorado.
Figure 30. Horizontal Stress Axes Map for the Basement Rocks (Stippled) of the Royal Gorge Arch Showing Sigma 1, Sigma 2, and Sigma 3 for Each Domain. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Figure 31. Horizontal Stress Trajectory Map for the Basement Rocks (Outlined) of the Royal Gorge Arch Showing Maximum and Minimum Horizontal Stress Directions. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
CHAPTER IV

STRUCTURAL INTERPRETATIONS

Introduction

This chapter includes interpretations of stress data, stress trajectories, and cross section constructions across the Royal Gorge arch. Comparisons are made between stress directions and lithology differences, and relationships are examined between domain boundaries and fold hinge surfaces. A fault-propagation fold model is interpreted for the uplift, using a blind principal fault. The chapter concludes with a structural summary of the development of the Royal Gorge arch as interpreted from the dynamic analysis.

Paleostress Trajectories

Because the domains defined by kinematic/dynamic analysis are approximately the same as the foliation domains (compare Figures 4-6 with Figures 15-27), stresses are interpreted to have changed orientation through the basement core of the Royal Gorge arch in response to fabric and lithologic differences. These orientation changes are traceable from domain to domain (Figure 31). Stress trajectories also appear to have been more intensely warped in the quartzofeldspathic gneiss, where paleostress patterns were strongly influenced by pre-existing fabric orientation, as compared to the weakly-foliated granodiorite and sedimentary units. The horizontal stresses apparently travelled through the quartzofeldspathic gneiss in orientations both parallel and
perpendicular to the curved foliation direction (see representative foliation strikes in Figure 31). The basement rocks to the north, in the granodiorite, did not respond in this fashion however, as basement fabric was not a dominant control. The relatively uniform stress direction through the granodiorite (N 60-70 E) is interpreted to be simply due to a more homogeneous lithology.

The changes in stress orientations do not appear to have been caused by rotation during the Laramide as the rotational angles and senses of rotation of the domains defined by foliations and the kinematic/dynamic analysis do not always match (Figure 32 and Figure 7).

Slickenside surfaces are interpreted to have shown their latest movement during the Laramide event, and not during the Ancestral Rocky Mountain event (nor any earlier deformation), because the slickensides in the Ordovician through Cretaceous-age rocks are consistent with a single stress system. Any pre-Laramide movement records may have been simply overprinted during the Laramide event. This final movement could only have influenced rocks after their deposition, and therefore, the latest movement must be younger than the Cretaceous sedimentary rocks. The interpreted stress field also matches the generally assumed, regional, Laramide stress pattern for the Front Range, supporting the interpretation that the slickensides developed during the Laramide (Jurista and others, 1995). For this reason, and because there were no stereonet patterns that suggest two different movement arrays, it is assumed that Neogene deformation had only a minor affect on the Royal Gorge arch. The sedimentary rocks on the extreme northern and southern sections of the arch (Domain I) do appear to more closely match a Neogene, extensional, stress pattern. Local reactivation of these areas (adjacent to the Mikesell Gulch fault and to the Wet Mountains) during the Neogene appears likely.
Figure 32. Contoured Poles to Basement Foliation and Slickenside M-plane Diagrams for Domains VII (South) and VIII.
The formation of basement domains was apparently strongly influenced by pre-existing fabric and lithologic differences, as shown by domain boundaries coinciding with lithologic boundaries (Figure 7). This implies that the basement domains were established during an earlier, possibly Precambrian, folding event, and were subsequently modified by the Laramide event. The development of three, separate, domains in the sedimentary rocks is interpreted as a differential response to these heterogeneities in the underlying basement rocks, and probably not due to a second Laramide folding event.

Joint patterns in the crystalline basement and adjacent sedimentary units do not generally match (Figures 9-13 and Figure 7). These differences may also be explained by heterogeneous folding of the sedimentary units in response to fabric differences in the underlying basement rocks. Differences were also expected because of the inherent fabric change between the sedimentary rocks and basement rocks.

Cross Section Constructions

Two cross sections (Figures 33 and 34) were constructed across the Royal Gorge arch. They were drawn approximately east-west, parallel to the Gorge fault (Figure 2). Cross section A-A' is drawn north of the Gorge fault, and section B-B' is drawn to the south.

Both cross sections are interpreted as modifications of fault-propagation folds (Suppe and Medwedeff, 1984, and 1990) and were constructed assuming that fault angles and fault depths are predictable from surface dip domain data. The tips of the blind propagation thrusts were determined by projecting hinge surfaces down from the surface. The fault-propagation fold models can be easily modified to fault-bend folds by extending the tips of the blind propagation thrusts horizontally (to the east). Deep,
Figure 33. Cross Section A-A’ (No Vertical Exaggeration) Across the Northern Part of the Royal Gorge Fault-propagation Fold Interpretation. The Exact Location is Shown on Figure 7 Domains Roman Numerals.
Across the Northern Part of the Royal Gorge Arch Showing a Detachment -6.8 Mi (-11 Km). The Exact Location is Shown on Figure 7. Domains are Identified by

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Figure 34. Cross Section B-B' (No Vertical Exaggeration) Across the Southern Part of the Royal Gorge Fault-propagation Fold Interpretation. The Exact Location is Shown on Figure 7. Domain Roman Numerals.
Detachment -6.2 Mi (-10 Km)
near-horizontal detachment surfaces were then interpreted using arbitrary basement horizons (dashed on the cross sections), and maintaining a constant thickness to these horizons. The depth to this deep detachment varies from -10 to -11 km below mean sea level, and falls within the 10-15 km range estimated by C. F. Kluth (personal communication) for other areas of the Colorado Front Range.

The dip of the basement fault ramp changes from 20 degrees north of the Gorge fault to 13 degrees south of the Gorge fault. The fault angles were determined by the surface dip domain on the backlimb of the uplift.

Overall fold development of the Royal Gorge arch seems to approximate the model of Narr and Suppe (1989) and the mode 2 structures of Schmidt and others (1993), where basement faulting does not propagate into the overlying sedimentary cover. These models both interpret a cataclastically deformed basement forelimb instead of faulting. The deformation is confined to wedge-shaped zones of cataclasis and/or slip on basement fabric surfaces. Both cross sections (Figures 33 and 34) show fold hinge surfaces in the basement that stop at the sedimentary cover, with only local, minor offsets, similar to the interpretation by Schmidt and others (1993) for Kaufman Ridge anticline, Colorado. The Royal Gorge arch also differs from these published models by maintaining a deeper fault, which stays well within the basement on both sections. The tip of the fault is approximately the same depth below sea level on both sections.

Major differences in the two sections are the width of the anticline, ramp angle, and a slight difference in depth to the major detachment surface. These differences in geometry, as defined by the surface dip data, support the interpretation that the Gorge fault serves as a compartmental fault and effectively separates the deformation of the arch from north to south. The Gorge fault, and the smaller, unnamed, east-west
trending fault one mile (1.6 km) to the south, both appear to follow the basement foliation (Figure 2). This well-developed fabric in the quartzofeldspathic gneiss probably served as the easiest route for faulting even though the maximum principal stress direction varied as much as 45-degrees from the strike of the foliation. Oblique slip on the Gorge fault is also indicated by non-horizontal slickenlines in the basement and sedimentary units adjacent to the fault. Chase and others (1993) saw a similar fabric-fault relationship in small Rocky Mountain foreland structures, but only when the foliation was less than 15-degrees to the faulting. In these instances however, the authors were not examining tear faults, but frontal anticline bounding faults (principal faults) and anticline hinge surfaces.

By comparing the cross sections to the domain boundary map (Figure 7), an often coincident relationship is observed between the structural domain boundaries and the fold hinge surfaces. Structural domains are listed across the top of each cross section in Roman numerals for reference. It is suggested that at least some domain boundaries formed as a consequence of fault propagation hinge development, and were further complicated by the inherent fabric and lithologic variations in the basement rocks prior to uplift.

**Structural Development of the Royal Gorge Arch**

The Royal Gorge arch was subdivided into 10, homogeneous, structural domains based on fabric data, consisting of three domains developed in Phanerozoic, sedimentary rocks and seven domains in basement. The domains in sedimentary rocks developed during Laramide uplift of the arch in response to heterogeneities in the underlying basement rocks. The seven domains in basement rocks appear to have
developed during one of the Precambrian folding events, and were subsequently modified during the Laramide event.

The method of Aleksandrowski (1985) was successfully used on slickenside arrays to determine the stress field which caused the Laramide uplift of the Royal Gorge arch. Stress trajectories were fairly uniform in the surrounding sedimentary rocks, with the maximum compressional stress trending N 45-55 E. This direction closely matches the conclusions of Jurista and others (1995) for the Laramide development of the Colorado Front Range. Stress trajectories in the basement rocks were markedly more non-uniform however, as paleostress patterns were commonly controlled by fabric orientation.

Overall development of the Royal Gorge arch can be modelled as a fault-propagation fold (or a fault-bend fold), similar to models proposed by Narr and Suppe (1989) and Schmidt and others (1993). A deep, near horizontal detachment surface is estimated at 10-11 km below mean sea level upon which a 13-20 degree ramp developed, creating the uplift. The principal basement fault remained well within the basement (blind), with only minor faulting expressed at the surface.

The Royal Gorge arch appears to have been effectively compartmentalized from north to south by the Gorge fault. Well developed foliation in the basement rocks near the fault apparently served as planes of slip, even though oriented 35-45 degrees from the direction of maximum compression in adjacent sedimentary rocks, and 15-20 degrees from the direction of maximum compression in adjacent foliated rocks.
CHAPTER V

HYDROGEOLOGIC ANALYSIS

Introduction

This chapter presents the hydrogeochemical and isotopic data sets gathered from water samples within the Royal Gorge crystalline aquifer. A general discussion of ground water chemical changes within fractured aquifers is included, along with the results of previous studies near the Royal Gorge arch. Ground water sampling strategy is outlined and precision techniques are detailed. The chapter concludes with an introduction to hydrogen and oxygen isotope methods.

Mixing in Fractured Media

Many of the concepts of mixing in porous media also apply to fractured media. On a macroscopic scale, transport in a fractured system is the result of both mechanical mixing and diffusion (Domenico and Schwartz, 1990). Although fractures are the primary flow paths for solute transport, matrix diffusion adjacent to the fractures must be included in overall solute movement. Diffusion in response to chemical gradients within the matrix appears to dominate concentration distributions in fractured media. This observation is in direct conflict with porous media, where diffusion is usually insignificant. Studies by Grisak and Pickens (1980) and Tang and others (1981) explain the dominance of diffusion in fractured rocks as a result of large concentration gradients due to localization of fluids in fractures. The net effect of
matrix diffusion is to provide solute "storage" in the rock mass, with the rate of change described by Fick's Law.

A porous medium is usually assumed to be isotropic with respect to dispersivity. Longitudinal dispersivity at any point has a single value regardless of the direction of the velocity vector (Freeze and Cherry, 1979). Likewise, transverse dispersivities have single values relative to the longitudinal dispersivity direction. Differences between longitudinal and transverse dispersivities are tied to the mechanism of dispersion, and not to directional properties of a porous medium. Because fractured aquifers are strongly anisotropic with respect to the orientation and density of fractures, dispersion cannot be simply described by the equations for homogeneous porous media (Freeze and Cherry, 1979).

Dispersion may also be influenced by partitioning of fluids at fracture intersections. Although partitioning does not reduce chemical concentration in itself, the mixing with other fluids in fractures will cause dilution (Domenico and Schwartz, 1990). Large-scale dispersion can occur if fractures exist on regional scales or cut across several rock units.

Chemical concentrations in fractures are also affected by fluid mixing at fracture intersections, variability of apertures, and any velocity changes associated with different scales of fracturing (Domenico and Schwartz, 1990). Variable fracture aperture and roughness can give rise to "channeling," where fluids move along a preferred network of irregularly spaced pathways, bypassing much of the fracture network. Neretnieks (1985) showed that channels may not intersect one another for significant distances, a phenomena termed "pure" channeling. If this situation develops, the dispersivity will continue to increase with travel distance, never reaching an asymptotic value.
Analysis of fluid mixing through fractures with diffusion into the matrix is considered important in many hydrologic subdisciplines. Examples include studies of aquifer recharge, ground water geochemistry, contaminant transport, ground water age-dating, and radioactive waste disposal (Grisak and Pickens, 1980).

Previous Studies Near the Royal Gorge Arch

In the last decade, the United States Geological Survey has published two water quality investigations near the present study area (Cain, 1987; Hearne and Litke, 1987). Cain looked primarily at the surface waters of the upper Arkansas River, and determined various relations between conductivity, streamflow, and water quality. Linear regression equations were developed to predict the total dissolved solid content and selected major ion concentrations from conductivity measurements.

Hearne and Litke, by contrast, examined ground waters from the sedimentary section to the east of the Royal Gorge arch, and immediately south of Canon City (the suburban community of Lincoln Park). The authors looked at the mixing effects of raffinate (liquid waste from uranium ore processing) on other adjacent ground water aquifers. They concluded that raffinate from nearby uranium processing operations was changing the chemical characteristics of the local ground water.

The report by Hearne and Litke (1987) also includes the only published stable and radioactive isotope data for the Royal Gorge area. Hearne and Litke concluded that uranium ore-tainted (raffinate) ground waters could be identified using a combination of $\delta D$, $\delta^{18}O$, $\delta^{34}S$, $\delta^{13}C$, and tritium. They noted a $\delta D$ to $\delta^{18}O$ ratio indicative of evaporation from uranium tailings ponds, and tritium values implying recharge since 1952. Although they examined aquifers outside the bounds of the
present study, the $\delta^D$, $\delta^{18}\text{O}$, and tritium results are included below in Table 2 for comparison. All sample locations are shown in Figure 2.

Table 2
Selected Isotope Ratios From Hearne and Litke (1987)

<table>
<thead>
<tr>
<th>Station # (See Figure 2)</th>
<th>$\delta^D$ (per mil)</th>
<th>$\delta^{18}\text{O}$ (per mil)</th>
<th>Tritium (TU)</th>
</tr>
</thead>
<tbody>
<tr>
<td>48 (well)</td>
<td>-78.5</td>
<td>-10.5</td>
<td>1±4</td>
</tr>
<tr>
<td>49 (well)</td>
<td>-73.5</td>
<td>-10.0</td>
<td>-2±4</td>
</tr>
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<td>-73.5</td>
<td>-9.9</td>
<td>NA</td>
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<tr>
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<td>-9.9</td>
<td>-2±4</td>
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<td>-9.9</td>
<td>3±4</td>
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<td>0±5</td>
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<tr>
<td>55 (well)</td>
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<td>-10.0</td>
<td>46±5</td>
</tr>
<tr>
<td>56 (well)</td>
<td>-84.0</td>
<td>-11.4</td>
<td>5±4</td>
</tr>
<tr>
<td>57 (well)</td>
<td>-86.5</td>
<td>-11.4</td>
<td>3±4</td>
</tr>
<tr>
<td>58 (well)</td>
<td>-81.5</td>
<td>-10.7</td>
<td>26±5</td>
</tr>
<tr>
<td>59 (well)</td>
<td>-79.5</td>
<td>-9.6</td>
<td>75±6</td>
</tr>
<tr>
<td>60 (well)</td>
<td>-83.0</td>
<td>-10.2</td>
<td>64±5</td>
</tr>
<tr>
<td>61 (well)</td>
<td>-95.5</td>
<td>-12.1</td>
<td>41±6</td>
</tr>
</tbody>
</table>
Hydrogeochemistry of Fractured Crystalline Aquifers

Ever since the work of Headden (1903), it has been widely held that ground waters from quartzofeldspathic igneous rocks are Na-Ca-HCO₃ dominated, and rich in silica. Garrels (1967) prepared the following comprehensive summary of water-rock interactions for igneous systems:

1. The contents of the ground waters are due to soil waters containing CO₂ in excess of equilibrium with the atmosphere.

2. Plagioclase feldspar and mafic minerals are weathered much more rapidly than K-feldspar or quartz.

3. The usual high percentage of plagioclase relative to mafic minerals leads to high amounts of Na⁺ and Ca²⁺ in the waters.

4. Dissolved silica is mostly from plagioclase and mafic minerals, and very little from quartz dissolution.

5. Potassium and magnesium ions probably come from biotite and/or hornblende.

6. Bicarbonate is the dominant anion in most waters. High concentrations of sulfate may be from dissolution of anhydrite, or more commonly, from oxidation of pyrite. Chloride ions are commonly "contaminants" from precipitation or from fluid inclusions.

7. The chief and first weathering product is usually kaolinite.

Nordstrom and others (1989) term ground waters derived from outside the rock mass as "allochthonous," and ground waters from inside the rock mass as "autochthonous." Allochthonous waters have chemically evolved from a source...
independent of the rock mass, and were emplaced into the rock at a later time. Autochthonous, by contrast, implies that the ground water geochemistry is derived from the rock mass, and is dependent on the mineralogy and water-rock interactions during residence in the rock. Very old ground waters are expected to show a convergence of the autochthonous and allochthonous categories as the ground water and rock framework equilibrate.

In recent years, three important events have focused renewed attention on crystalline rock aquifers: (1) the results of the Kola superdeep drillhole, (2) radioactive waste research in crystalline complexes, and (3) energy research on hot dry-rock systems (Nordstrom and others, 1989).

The 13 km drillhole in the Kola Peninsula in the Fennoscandian Shield has penetrated some of the oldest rocks in the Archean, encountering Ca-Na-Cl dominated brines at depths of 0.8 to greater than 9 km (Kozlovsky, 1984). Ground waters from fractured crystalline rocks of the Canadian Shield have also been surveyed as part of an assessment for nuclear waste disposal.

Gascoyne and Kamineni (1994) found that typical Precambrian Shield ground waters start out as dilute, slightly alkaline and Ca-Na-HCO₃ in composition, regardless of specific rock type. The waters tended to evolve to higher pH, Na-Ca-HCO₃ compositions due to reactions with plagioclase feldspar, calcite precipitation, and ion exchange on clays. At depths greater than 200 m, the chemical composition was found to be a function of the dissolution of salts from the rock matrix, and generally increased in salinity with depth. Studies at the Stripa mine in Sweden (Nordstrom and others, 1989) and in the Canadian Shield (Frape and Fritz, 1982) both indicate a pronounced enrichment in deuterium relative to the meteoric water line for samples.
with very high total dissolved solids. Frape and others (1984) interpret these ground waters to be extremely old, suggesting that they may represent ancient crustal fluids.

Stednick (1989) noted that stream water chemistry across the Colorado Front Range is quite variable. His findings for streams in two small alpine basins in northern Colorado varied from Ca-HCO$_3$ to Na-HCO$_3$ in composition. He also found that the rate of snowmelt and the route of subsequent runoff was affecting the total dissolved solids in the streams.

Wood and Low (1986) conclude that the eastern Snake River Plain aquifer in Idaho appears to be an "autochthonous" ground water, where chemical reactions with the rock framework dominate the system. Large quantities of sodium, chloride, and sulfate are being removed by solution from the aquifer rock mass, whereas, calcite and silica are being precipitated. Mineralogic and thermodynamic evidence suggests that olivine, pyroxene, plagioclase, and pyrite are being actively dissolved, with the olivine dissolving faster than the pyroxene. Sulfur isotopic analysis indicates that 65% of the sulfate in solution is derived from dissolution of anhydrite and 35% is from pyrite. The authors conclude that water-rock reactions within the eastern Snake River Plain aquifer act as both a source and a sink for selective solutes.

**Hydrogeochemical Sampling Strategy**

Forty-seven ground-water and surface-water samples were collected from all available water wells, springs, and streams in the Royal Gorge crystalline aquifer during August 1994. Three additional stable isotope samples were taken from the Arkansas River in May 1995. The location of each sample is shown in Figure 2. All water wells were purged prior to sampling, at least to the extent possible, without draining entire cisterns or holding tanks. Whenever possible, wells were sampled from
the wellhead, and not from a holding tank. Detailed notes were taken about the locations of the water wells and whether or not a holding tank was used and/or circumvented during sampling.

Unfiltered samples were collected in 1-liter, clean plastic bottles at each station. Nine measures of water quality were determined at each site, and listed in Appendices B and C. Water temperature, pH, conductivity, and iron content were measured immediately at the time of sample collection. The remaining six water quality measurements shown in Appendices B and C were determined within 12 hours of sample collection. Repeats of every measurement were run every fifth sample (samples 5, 10, 15 etc...) for quality control, documenting the consistency of the instrumentation within a small percentage of error. The values contoured on Figures 35-41 used only an average value of the repeats.

Hydrogeochemical Parameters and Ions

The following sections detail the specifics of each of the water quality parameters and ions listed in Appendices B and C. Individual instrumentation is discussed, and any necessary calibrations noted.

Alkalinity

Alkalinity is the equivalent sum of bases titratable with a strong acid needed to bring a water to a specified carbonate system end point (Lloyd and Heathcote, 1985). Total alkalinity is determined at the bicarbonate (HCO$_3^-$) end point, approximately pH 4-5. At this pH, the proton condition corresponding to a pure CO$_2$ solution is satisfied as shown below:
Figure 35. Map of Alkalinity (as Calcium Carbonate) Distribution in the Northern Part of the Royal Gorge Crystalline Aquifer. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Figure 36. Map of Conductivity Distribution in the Northern Part of the Royal Gorge Crystalline Aquifer. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Figure 37. Map of Hardness (as Calcium Carbonate) Distribution in the Northern Part of the Royal Gorge Crystalline Aquifer. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Figure 38. Map of pH Distribution in the Northern Part of the Royal Gorge Crystalline Aquifer. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Figure 39. Map of Chloride Distribution in the Northern Part of the Royal Gorge Crystalline Aquifer. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Figure 40. Map of Iron Distribution in the Northern Part of the Royal Gorge Crystalline Aquifer. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Figure 41. Map of Nitrate (as Nitrogen) Distribution in the Northern Part of the Royal Gorge Crystalline Aquifer. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
\[ [H^+] = [\text{HCO}_3^-] + 2[\text{CO}_3^{2-}] + [\text{OH}^-] \]

The total alkalinity end point is usually determined with the indicator methyl orange. This point represents the total amount of strong alkali that would have to be added to a \( \text{CO}_3^-\text{H}_2\text{O} \) solution to produce the original pre-titration solution. Total alkalinity is almost identical with the bicarbonate concentration at pH values less than 9, and also in hard waters. In hard waters, the dissolved magnesium and calcium content is high and the calcium carbonate content is relatively low as shown by the following equation:

\[ \text{Ca}^{2+} + 2\text{HCO}_3^- = \text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 \]

Alkalinity can be either expressed in milliequivalents per liter or as the equivalent amount of calcium carbonate (mg/l as CaCO₃).

Total alkalinity was measured in the field using the Hach Company Digital Titrator, using a Bromcresol Green-Methyl Red Indictor Powder Pillow. Sulfuric acid was added to the solution as the titrating acid. Upon reaching pH = 5.1, the solution changed from a green to a gray end point, giving the total alkalinity of the sample. For higher alkalinity solutions, the titration continued to a pH = 4.5, where the solution turned a light pink color. All samples were recorded in mg/l of equivalent CaCO₃.

The distribution of total alkalinity in the Royal Gorge crystalline aquifer is shown in Figure 35, and listed in Appendix B. Most alkalinites from ground water samples were in the 150-300 range of mg/l as CaCO₃, whereas the stream and river samples had considerably lower alkalinites, in the range of 58-186 mg/l as CaCO₃. One sample from a spring and one sample from a water well had anomalously high alkalinites, 779 mg/l and 620 mg/l, respectively.
Conductivity

Conductivity is a measure of the ability of a solution to conduct an imparted electrical current. The conductance reflects the concentration and charge of the ions, and the rate at which the ions move under the influence of a potential (Lloyd and Heathcote, 1985).

Electrical conductivities of ground water samples are recorded in micromhos per centimeter or microsiemens per centimeter (µS/cm), as the units are numerically the same. In this study, conductivities are reported in the latter, microsiemens per centimeter. Both ionic activity and electrical conductivity increase with increasing temperature. Therefore, temperatures must be recorded simultaneously with conductivity, and normalized to 25 degrees C for standard comparison. Lloyd and Heathcote (1985) comment that since natural waters contain a variety of ionic and uncharged particles, conductivity measurements cannot be used as a direct substitute for determinations of total dissolved solids (TDS). They do note however, that normally an approximate correlation exists between conductance and TDS, and the following equation is commonly used:

\[ \text{TDS} = k \times \text{Conductivity} \]

Where "k" is a constant correlation factor, normally between 0.55 and 0.80 for ground waters. TDS is expressed in mg/l and conductivity is in µS/cm.

Conductivities were measured using a Jenco Electronics, Ltd. Model 103, with ranges from 0-20,000 µS/cm. The instrument automatically converted the conductivity values to the standard temperature of 25 degrees C.
The conductivities measured on waters from the Royal Gorge crystalline aquifer are listed in Appendix B, and their areal distribution is shown in Figure 36. The samples from springs exhibited the highest conductivities with a mean value of 1857±1434 μS/cm. One spring sample had a conductivity of 5445 μS/cm (Sample #25), which was more than double any other water sample in the aquifer. Water wells exhibited a mean conductivity of 1301±409 μS/cm, while stream samples had a mean conductivity of approximately 300 μS/cm.

**Hardness**

Hardness is the property of water that causes difficulty in lathering of soap and scaling in boilers and kettles (Lloyd and Heathcote, 1985). Hardness is caused by the presence of divalent cations in water, especially Ca²⁺ and Mg²⁺, but also Fe²⁺ and Sr²⁺. Generally, hardness is reported in terms of mg/l of calcium carbonate, just as alkalinity was reported above. Because both alkalinity and hardness are tied to the bicarbonate ion, there is a natural relationship between the two. The difference between the total hardness and alkalinity is the permanent hardness (Lloyd and Heathcote, 1985).

Mineral reactions with acidic ground waters from the soil zone (see pH section below) can add considerable amounts of Ca²⁺ and Mg²⁺ to the ground water, as shown below:

\[
\begin{align*}
Na_0.5Ca_0.5Al_{1.5}Si_{2.5}O_8 + 3/2 H^+ + 11/4 H_2O &= \text{Andesine} \\
3/4 Al_2Si_2O_5(OH)_4 + 1/2 Na^+ + Ca^{2+} + Si(OH)_4 &= \text{Kaolinite} \\
KMg_3AlSi_3O_{10}(OH)_2 + 7H^+ + 1/2 H_2O &= \text{Phlogopite}
\end{align*}
\]
\[ \frac{1}{2} \text{Al}_2\text{Si}_2\text{O}_5(\text{OH})_4 + \text{K}^+ + 3\text{Mg}^{2+} + 2\text{Si(OH)}_4 \]
\[ \text{Kaolinite} \]

\[ \text{Ca}^{2+} + \text{SO}_{4}^{2-} = \text{CaSO}_4 \]
\[ \text{Gypsum} \]

The calcium and magnesium ions created through these reactions are balanced by the formation of calcium carbonate by the following reaction if saturated (which is also tied to the pH and alkalinity balance of the water):

\[ \text{Ca}^{2+} + 2\text{HCO}_3^- = \text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 \]

In addition, Domenico and Schwartz (1990) believe that the most important cation exchange reactions are the water-softening reactions where Ca\(^{2+}\) and Mg\(^{2+}\) exchange with Na\(^+\) as the water moves through clayey material. This process can also effectively reduce calcium and magnesium concentrations, while increasing sodium concentrations. Dissolved sodium in ground water can be produced in the andesine to kaolinite reaction above, and also by reactions with albite and montmorillonite (Freeze and Cherry, 1979). The balance between all of the above reactions (including pH and alkalinity reactions) determines the total hardness of a ground water system.

Total hardness was measured in the field with a Hach Company Digital Titrator, using an EDTA titration cartridge, Hardness 1 Buffer Solution, and a Man Ver 2 Hardness Indicator Powder Pillow. The proper hardness for each sample was determined when the solution changed from red to pure blue. The measured total hardness values for water samples from the crystalline aquifer are listed in Appendix B, and the distribution areally is shown in Figure 37.
pH

The pH of a solution is a measure of acid-base equilibria. Acids and bases are generally defined in terms of proton transfer. Water can be viewed as possessing properties of a weak acid and a weak base, as shown below:

\[ H_2O + H_2O = H_3O^+ + OH^- \]

Where \( H_3O^+ \) is the hydronium ion, and \( OH^- \) is the hydroxyl ion. This equation shows that pure water must have equal amounts of hydronium and hydroxyl ions, and is termed "neutral." If there are more hydronium ions than hydroxyl ions the solution is "acid," and if the opposite is true, the solution is "alkaline."

Because the concentration of hydrogen ions can vary by many orders of magnitude, pH has been defined by the following equation:

\[ pH = -\log (H^+) \]

Where by convention, \( (H^+) = (H_3O^+) \), and \( (H^+) \) is the hydronium activity and not concentration (Lloyd and Heathcote, 1985). The pH of a solution is unitless.

Water samples from the Royal Gorge crystalline aquifer were analyzed immediately in the field with a portable, Markson Model 6077, Microcomputer pH-Vision brand pH meter. Each day, prior to sample collection, the portable pH meter was calibrated with known solutions of \( pH = 7 \) and \( pH = 10 \), as the sampled waters were dominated by \( pH \) values greater than 7. Temperatures were also recorded by the instrument simultaneously.
Rainwater and melted snow in nonurban areas generally have pH values between 5 and 6, as the equilibrium pH for freshwater in contact with CO\textsubscript{2} at the earth's atmospheric pressure (10\textsuperscript{-3.5} bar) is 5.7 (Freeze and Cherry, 1979).

The dissolution of feldspars, micas, and other silicates in granitic terrains is strongly influenced by the chemically aggressive nature of water with elevated CO\textsubscript{2} concentrations. Dissolved CO\textsubscript{2} in the soil zone (from organic decay) quickly combines with water in the following reaction:

\[
H_2O + CO_2 = H_2CO_3 = H^+ + HCO_3^-
\]

The H\textsuperscript{+} ions then react with the common silicate minerals in the Royal Gorge crystalline aquifer, such as microcline and andesine (see Appendix A), to form kaolinite and other clays, as below:

\[
\begin{align*}
KAlSi_3O_8 + H^+ + 9/2 H_2O &= 1/2 Al_2Si_2O_5(OH)_4 + K^+ + 2Si(OH)_4 \\
\text{Microcline} &\rightarrow \text{Kaolinite}
\end{align*}
\]

\[
\begin{align*}
Na_0.5Ca_0.5Al_1.5Si_2.9O_8 + 3/2 H^+ + 11/4 H_2O &= \\
\text{Andesine} &\rightarrow 3/4 Al_2Si_2O_5(OH)_4 + 1/2 Na^+ + Ca^{2+} + Si(OH)_4 \\
\text{Kaolinite} &\rightarrow 
\end{align*}
\]

The resulting incongruent dissolution of feldspars, micas, and clays involves consumption of H\textsuperscript{+} ions, causing a progressive increase in pH as the reactions proceed. Ground waters travelling through such rocks can rapidly acquire pH values of 7-8 and concentrations of HCO\textsubscript{3}\textsuperscript{-} of tens to hundreds of milligrams per liter (Freeze and Cherry, 1979). Some bicarbonate ions and some Ca\textsuperscript{2+} ions are then lost by calcite precipitation in fractures, as documented in thin sections from the Royal Gorge.
crystalline aquifer (Appendix A). This reaction was also observed by Gascoyne and Kamineni (1994) in plutonic aquifers of the Canadian Shield, and shown below:

\[ \text{Ca}^{2+} + 2\text{HCO}_3^- = \text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 \]

All measured pH values and temperatures are listed in Appendix B and contoured in Figure 38.

**Temperature**

Temperature measurements were taken with the pH instrument in order to compare temperatures from spring to spring, and also to serve as a cross-check of ground water samples. The conductivity and pH instruments converted the actual temperature measurements to standard conditions for data comparison.

The springs consistently showed the highest temperatures as they were most affected by the summer daily temperature maximums (Appendix B). The water in the Arkansas River was also fairly warm, compared to most ground waters, as it too was exposed to the high summer heat during discharge.

**Chloride**

Chloride is generally considered to be a conservative ion, as it tends to stay in solution and not react appreciably during ground water flow (i.e. limited adsorption and/or ion exchange). For this reason, chloride transport is assumed to be dominated by advection, and an ion is usually believed to travel at the same velocity as water with no retardation.
Chloride concentration was measured with a Hach Company Digital Titrator using the Mercuric Nitrate Method, with the Diphenylcarbazone Powder Pillow as indicator. The chloride content was determined when the solution changed from pale yellow to light pink. To avoid chloride interference from the pH electrodes, the chloride content was determined using separate water samples from the pH measurements.

The results of the chloride measurements are given in Appendix C, and the distribution across the Royal Gorge crystalline aquifer is shown in Figure 39. The concentrations are again highest in the spring samples (mean 153±155 mg/l) and lowest in the Arkansas River (mean 7±3 mg/l), with the well samples and Grape Creek samples falling in between (means of 76±52 mg/l and 23±5 mg/l, respectively). These values follow the same trends as conductivity, with chloride content mimicking the overall ionic concentrations.

Iron

Iron concentration was measured using a portable Hach Company colorimeter. The iron content was detected using a FerroVer Iron Reagent Powder Pillow (1,10-Phenanthroline Method) with an orange color indicating the presence of iron in the sample. As noted above, the iron concentration was determined immediately upon sampling the various waters, as was pH, temperature, and conductivity.

The concentration of iron in the Royal Gorge crystalline aquifer was uniformly quite low (see Appendix C). Only samples #1, #20, and #34 had concentrations greater than 1.0 mg/l, with only sample #1 having a concentration above the measurable instrument maximum (>5.0 mg/l). The spotty nature of the iron
concentration is clearly visible in Figure 40, which shows four isolated areas above background levels.

**Nitrate**

Nitrate (NO$_3^-$) has been increasingly mentioned in recent popular literature as a health hazard (Stuart and others, 1995). While the nitrate ion is not the real danger, the conversion from nitrate to nitrite (NO$_2^-$) within the digestive tracks of human infants and various farm animals can create life-threatening situations (Tyson and Harrison, 1989).

The nitrate ion is easily introduced into ground water systems from septic tank effluent, feed lots, and livestock manure (Stuart and others, 1995). Nitrates can also be produced in small concentrations by electrical atmospheric discharges and within certain plants. Hazell and others (1992) concluded that some nitrate in Nigerian basement aquifers was from soil bacteria.

The conversion of organic nitrogen from septic tank effluent and manure to the ammonium ion (ammonification or mineralization), followed by oxidation to nitrate (nitrification), is commonly cited as a way to produce high concentrations of nitrate (Freeze and Cherry, 1979). Nitrate produced through this process is easily transported advectively due to its high solubility (Stuart and others, 1995). The EPA Maximum Contaminant Level (MCL) is given as 10 mg/l nitrate as nitrogen (NO$_3$-N) (U.S. Environmental Protection Agency, 1990).

Nitrate (as nitrogen) concentration was measured using a portable Hach Company Colorimeter, with a NitraVer 5 Nitrate Reagent Powder Pillow. An amber color developed when nitrate was detected with the reagent. The measurements taken
on water samples from the Royal Gorge crystalline aquifer are given in Appendix C, and the areal distribution is shown in Figure 41.

The mean NO$_3$-N concentration for water wells in the Royal Gorge crystalline aquifer was 4.3±3 mg/l, whereas spring and stream samples had mean NO$_3$-N concentrations of less than 1.5 mg/l. The difference in means between these sample subsets appears related to human activity. An arbitrary concentration cut-off of >5 mg/l NO$_3$-N was chosen in order to delineate impacted areas of the aquifer. This was the same concentration criteria chosen by Kehew and Brewer (1992) to define their human impacted ground water.

**Sulfate**

Freeze and Cherry (1979) report that sulfate (SO$_4^{2-}$) is normally not a significant constituent in silicate rocks, and there is no tendency toward development of a sulfate facies in ground waters through these rock types. However, there have been significant amounts of sulfate reported recently in ground waters from basement aquifers. Chilton and Foster (1995) reported sulfate concentrations greater than 2000 mg/l from basement aquifers in Africa, and Gascoyne and Kamineni (1994) reported sulfate concentrations of up to several 100 mg/l in crystalline aquifers in Canada.

Concentrations of sulfate in water samples from the Royal Gorge crystalline aquifer were measured with a portable Hach Company Colorimeter. The contents of a SulfaVer 4 Sulfate Reagent Powder Pillow were added to each sample, with a white turbidity developing if sulfate was detected.

Sulfate concentrations were found to be fairly high in all well and spring samples (ground waters), with values beyond the maximum measurable limit of the instrument. The immediate presence of a white turbidity was observed in every
These values were reported in Appendix C as being greater than 80 mg/l. Only in the stream and river samples were sulfate concentrations found to be consistently less than 80 mg/l, generally between 20 and 35 mg/l.

Isotopic Analysis

Hydrogeologic Importance of Isotopes

Applications of isotope hydrology are being used increasingly in aquifer characterizations where investigators need to determine the origin and/or the mixing of waters. Environmental isotopes, as direct tracers of the water itself, can provide information on the spatial and temporal patterns of a flow system. It is generally assumed that if the isotope content does not change along a flow path, it will reflect the source of the water. Stable isotopic tracers have an advantage over chemical tracers in that they do not become diluted and do not readily interact with the aquifer material at low temperatures. Recent research by Johnson and DePaolo (1994) used strontium ratios in an attempt to exploit the limited water-rock interactions, providing a model that related reaction rates of isotopic tracers to spatial variations in ground water velocity through fractures. They observed that more chemical reactions take place in very slow-moving ground waters, and that the resultant isotopic signature will reflect the increased residence time.

In fractured crystalline aquifers, it is generally accepted that ground water moves nearly exclusively through cracks and fractures in the impermeable rock. Because of the often unpredictable nature of these heterogeneities, Fontes (1980) has suggested two specific applications of isotope hydrology for fractured aquifers. First,
questions of mixing within the fractured system can be addressed by spatial sampling of ground waters, looking for possible differences across the aquifer. Second, questions of flow velocity and recharge can be addressed by testing for tritium occurrence. Different recharge and flow histories are indicated by either a lack of tritium content, the occurrence of a homogeneous tritium content, or the occurrence of a heterogeneous tritium content.

Stable isotopic concentrations in most ground waters reflect the local precipitation. Condensation, which involves isotope fractionation, depends on temperature. When the isotopic species removed are in thermodynamic equilibrium with those remaining in the system, the condition can be expressed by a Rayleigh-type equation (Hoefs, 1987):

\[(R_v/R_{vo}) = f(\alpha - 1),\]

where \(R_{vo}\) is the isotope ratio of the initial bulk composition, \(R_v\) is the instantaneous ratio of the remaining vapor, the fractionation factor is \(\alpha = R_l/R_v\) (\(l = \) liquid and \(v = \) vapor), and \(f\) is the fraction of the residual vapor. The instantaneous isotope ratio of the condensate leaving the vapor \((R_l)\) is given by the following equation (Hoefs, 1987):

\[(R_l/R_{vo}) = \alpha f(\alpha - 1).\]

In general, the more pronounced the cooling process, the more the water vapor becomes depleted in "heavy" isotopes \((^{18}\text{O} \text{ and D})\). The "heaviest" isotopes of water are preferentially the first to fall as precipitation. The resulting depletion of the D/H and \(^{18}\text{O}/^{16}\text{O}\) ratios in the residual cloud vapor and the instantaneous composition of
the raindrops from the cloud can be given as a function of the fraction of the vapor remaining in the cloud. Multiple cooling events will also give a progressively depleted condensed phase.

There are several isotopic effects which are directly or indirectly attributable to the condensation process:

1. The altitude effect causes precipitation to become progressively depleted with increasing elevation.

2. The continental effect causes precipitation to become progressively depleted with increasing distance from marine and coastal locations. The further the water vapor travels the "lighter" it becomes isotopically.

3. The seasonal effect shows winter precipitation to be more depleted than summer precipitation.

4. The amount effect shows a tendency for heavier rainfall events to be more depleted (Fontes, 1980), although the reverse situation can also occur (enrichment).

In contrast to condensation, the evaporation process tends to cause isotopic enrichment of the remaining liquid phase. On a $\delta$D vs. $\delta^{18}$O diagram, evaporated waters fall below the meteoric water line (Fontes, 1980).

Hydrogen isotopes are fractionated in proportion to oxygen isotopes in the condensation/evaporation processes because a corresponding difference in vapor pressure exists between H$_2$O and HDO, and between H$_2^{16}$O and H$_2^{18}$O (Hoefs, 1987). The average isotope values of fresh waters and precipitation were shown by Craig (1961a) to follow a linear relationship called the "Meteoric Water Line," where:

$$\delta D = 8\delta^{18}O + 10$$
Hoefs (1987) noted that neither the coefficient "8" nor the constant "10," also called the deuterium excess "d" if the slope is close to eight, are really constant. Both values have been found to vary according to local climatic differences (Harmon and Schwarcz, 1981; Jouzel and others, 1982; Leguy and others, 1983; Machavaram and Krishnamurthy, 1995).

Deviations from the Meteoric Water Line (MWL) usually tend toward low δD and/or high δ18O values (plotting to the right side of the line), and are typically attributed to evaporation effects. These deviations are most pronounced in areas where the relative humidity is consistently less than 100% (Hoefs, 1987).

Isotopic data should not be used as a "stand-alone" method to solve hydrogeologic problems. The application is best viewed as another tool to be utilized by the hydrogeologist whenever solutions to water body source, recharge, and mixing are necessary. Like many other methods, isotopic studies must be considered within the framework of the overall hydrogeological data.

Lloyd and Heathcote (1985) point out that isotopic studies can be both expensive and inconclusive. Widespread and continued use of environmental isotopes since the 1960's however, has repeatedly demonstrated their application and importance in hydrogeological research. When used in combination with other geochemical and hydraulic data, isotopic data will give less ambiguous and quite irrefutable results (Payne, 1981).

**Isotope Definitions**

By definition, an isotope of an element has the same number of protons, but a different numbers of neutrons, changing only the atomic weight of the element. Isotopes are termed stable if they do not naturally decay, and unstable or radioactive if...
they undergo radioactive decay. Stable isotopes are commonly used in hydrogeological studies when examining the sources of water and to understand the history of the water after entering the flow system. Applications involving radioactive isotopes are usually confined to determinations of the age of the waters.

Of the more than 1000 isotopes of naturally occurring elements, only about 36 are presently used in hydrogeologic studies, including hydrogen ($^1\text{H}$, $^2\text{H}$, $^3\text{H}$), carbon ($^{12}\text{C}$, $^{13}\text{C}$, $^{14}\text{C}$), nitrogen ($^{14}\text{N}$, $^{15}\text{N}$), oxygen ($^{16}\text{O}$, $^{18}\text{O}$), and sulfur ($^{32}\text{S}$, $^{34}\text{S}$). The hydrogen isotopes of masses 2 and 3 are termed deuterium (D) and tritium (T), respectively. These names and letter designations for the hydrogen isotopes will be used throughout the remainder of this report.

Isotope effects are defined by Hoefs (1987) as differences in chemical and physical properties arising from differences in atomic mass. Although isotopes will tend to follow similar chemical patterns, the addition or subtraction of neutrons changes the mass and alters the rate of chemical reactions (Hoefs, 1987). It is in the lightest elements that these mass differences are most pronounced, and it is these same elements that are utilized most frequently in hydrogeology (see above). Table 3 shows some of the physiochemical properties that differ from water molecule to water molecule.

Isotope reactions obey the Law of Mass Action and are described by temperature-dependent equilibrium constants, commonly called fractionation factors ($\alpha$). The fractionation factor is related to the isotopic equilibrium constant (K) by the equation: $\alpha = K^{1/n}$, where $n$ is the number of atoms exchanged in the reaction (Hoefs, 1987). The equilibrium constant of a chemical reaction is then directly related to an associated isotope exchange reaction. Mass balance calculations are therefore
Table 3

Physiochemical Properties of Selected Water Isotopes
Modified From Hoefs (1987, p. 4)

<table>
<thead>
<tr>
<th>Constants</th>
<th>$H_2^{16}O$</th>
<th>$D_2^{16}O$</th>
<th>$H_2^{18}O$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Melting Point (760 Torr, in Celsius)</td>
<td>0.00</td>
<td>3.81</td>
<td>0.28</td>
</tr>
<tr>
<td>Boiling Point (760 Torr, in Celsius)</td>
<td>100.00</td>
<td>101.42</td>
<td>100.14</td>
</tr>
<tr>
<td>Temperature of Greatest Density (Celsius)</td>
<td>3.98</td>
<td>11.24</td>
<td>4.30</td>
</tr>
</tbody>
</table>

Chemical equilibrium is a necessary condition for isotopic equilibrium. However, chemical equilibrium does not always proceed at the same reaction rate as isotopic equilibrium. Fritz and Fontes (1980) note that chemical equilibrium between carbon dioxide and water is more rapid than oxygen isotope equilibrium between the two.

Isotopic ratios ($R$) are expressed with the heaviest isotope as the numerator and the light isotope as the denominator ($^{18}O/^{16}O$). Isotopic ratios of coexisting species are expressed by fractionation factors:

$$\alpha_{A-B} = \frac{R_A}{R_B}$$

Where "A" and "B" are different chemical compounds or states.

Because equilibrium fractionation factors are close to unity, the enrichment factor $\varepsilon$ is often used and defined as:
\[ \varepsilon_{A-B} = (\alpha_{A-B} - 1) \times 1000. \]

This notation allows the expression of enrichment in permil (1/1000). Fritz and Fontes (1980) suggest that the enrichment factor would be better called the isotope separation factor, because it can be either positive or negative.

Because it is quite difficult to measure absolute isotope abundances, it has become customary to simultaneously determine relative isotope abundances using a standard. An agreed upon standard reference allows the relative measurements to serve as "absolute" abundances. The results are expressed in the \( \delta \) - notation. The relative differences are expressed in parts per thousand (permil).

\[ \delta \text{ (permil)} = \left( \frac{R_{\text{sample}}}{R_{\text{standard}}} - 1 \right) \times 1000 \]

An increasing value of \( \delta \) implies an increasing proportion of the heavier isotope as compared to the standard. Therefore, if one sample is "heavy" compared to another sample, it has a more positive \( \delta \) value.

Substituting the definition of \( \delta \) into the definition of the fractionation factor gives the following relationship between \( \delta \) and \( \alpha \):

\[ \alpha_{A-B} = \frac{1000 + \delta_{A}}{1000 + \delta_{B}}. \]

Isotope Standards

International isotope standards are necessary to make inter-laboratory comparisons and to interpret the \( \delta \) values. International standards must meet certain criteria: (a) they must remain stable for long time periods, (b) they must be homogeneous in composition, (c) they must be available in large quantities, and (d)
they must be easy to transport and store. The $\delta$ value of an international standard should also fall within the middle of the range of naturally occurring isotopic variation. The irregularities and calibration problems of international standards have been discussed in detail by Friedman and O’Neil (1977), Gonfiantini (1978), and Coplen and others (1983).

Because there is a finite amount of any international standard and expense involved in obtaining the standard, it is common practice to prepare a secondary or "lab" standard for routine measurements. These lab standards are frequently calibrated against the accepted international standard. The later conversion of all samples from the lab standard (LAB) to the international standard (STD) is given by:

$$\delta_{\text{sample}} - \delta_{\text{STD}} = \delta_1 + \delta_2 + \delta_1\delta_2 \times 10^{-3}$$

where, $\delta_1 = \delta_{\text{sample}} - \delta_{\text{LAB}}$, and $\delta_2 = \delta_{\text{LAB}} - \delta_{\text{STD}}$.

The international standard used for oxygen and hydrogen is termed Standard Mean Ocean Water (SMOW), and was first prepared by Craig (1961b). The value of SMOW corresponded to a hypothetical water with oxygen and hydrogen isotope ratios equal to the mean isotopic ratios of ocean water (Epstein and Mayeda, 1953; Horibe and Kobayabawa, 1960; Craig, 1961b).

The original SMOW standard has been exhausted, and has since been replaced V-SMOW (Vienna Standard Mean Ocean Water). V-SMOW was artificially prepared by Professor H. Craig of the University of California, La Jolla in 1967 at the request of the International Atomic Energy Agency (IAEA). The oxygen and hydrogen isotopic content of V-SMOW is nearly identical to SMOW, and has been defined with respect.
to an existing water standard of the National Bureau of Standards (NBS), Gaithersburg, MD, which is called NBS-1 where:

\[
\left( ^{18}O/^{16}O\right)_{\text{SMOW}} = 1.008 \left( ^{18}O/^{16}O\right)_{\text{NBS-1}}
\]

\[
\left( ^{D}/^{H}\right)_{\text{SMOW}} = 1.050 \left( ^{D}/^{H}\right)_{\text{NBS-1}}
\]

Craig (1961) evaluated the absolute isotope ratio of SMOW and found the following:

\[
\left( ^{18}O/^{16}O\right)_{\text{SMOW}} = (1993.4 \pm 2.5) \times 10^{-6}
\]

\[
\left( ^{D}/^{H}\right)_{\text{SMOW}} = (158 \pm 2) \times 10^{-6}
\]

The absolute values for V-SMOW have been measured by several authors and are listed as follows:

\[
\left( ^{18}O/^{16}O\right)_{V-\text{SMOW}} = (2005.2 \pm 0.45) \times 10^{-6}
\]

(Baertschi, 1976)

\[
\left( ^{D}/^{H}\right)_{V-\text{SMOW}} = (155.76 \pm 0.05; 155.95 \pm 0.08; 155.6 \pm 0.12) \times 10^{-6}
\]

(Hagemann and others, 1970; DeWit and others, 1980; Tse and others, 1980)

It has been generally agreed that the defined SMOW and V-SMOW are isotopically identical (Gonfiantini, 1978), and no further distinction will be made when reporting \( \delta \) values for either of the above standards.

Other water standards used by the IAEA and NBS for intercalibration of isotopic ratios in natural waters include SLAP (Standard Light Antarctic Precipitation)
and GISP (Greenland Ice Sheet Precipitation). SLAP was collected by melting ice and firn from Plateau Station, Antarctica in 1967. The $^{18}O$ and D isotopic content for SLAP is very low, approaching the lowermost range of natural water variation (Gonfiantini, 1981). GISP, by contrast, is a water with an $^{18}O$ and D isotopic content intermediate between V-SMOW and SLAP.

**Hydrogen Isotopes**

Hydrogen has two stable isotopes of mass one and mass two (deuterium) with atomic abundances of 99.984% and 0.015%, respectively (Fritz and Fontes, 1980). The radioactive isotope of hydrogen, tritium, occurs naturally and decays with a half-life of 12.3 years into $^{3}\text{He}$ (Wallick and others, 1984). Tritium is produced in the upper atmosphere by the collision of cosmic rays with oxygen and nitrogen atoms, producing thermal neutrons. These neutrons then collide with each other and with nitrogen to produce tritium (Wallick and others, 1984). Tritium has also been introduced "artificially" into the atmosphere via thermonuclear testing. Libby (1963) estimated that 12 times the natural level of tritium (about 360 kg) were in the waters of the northern hemisphere at the conclusion of the atmospheric nuclear testing cycle (1958-1963). A ratio of $T/H = 10^{-18}$ is defined as one tritium unit (TU).

**Oxygen Isotopes**

There are three stable isotopes of oxygen, $^{16}O$, $^{17}O$, $^{18}O$. Because $^{18}O$ has an approximate abundance of 0.1995% and $^{17}O$ has an abundance of 0.0375%, it has become convention to ignore $^{17}O$ measurement and concentrate on the ratio of $^{18}O/^{16}O$ to take advantage of the greater mass difference and abundance (Hoefs,
1987). Just like hydrogen isotopes, the main processes that affect oxygen isotopes are evaporation, condensation, and exchange reactions.

**Isotopic Sampling Strategy and Results**

Sampling of all available water wells, springs and streams for isotopic analysis took place in August 1994. All water wells were sufficiently purged prior to sampling. Stable isotope samples were transferred immediately to air-tight scintillation vials (20 ml) to prevent evaporation. Tritium samples from all wells were collected in 1-liter plastic bottles, and also immediately sealed to prevent contamination from atmospheric tritium. Every attempt was made to collect isotopic samples from the actual wells and not from a cistern or holding tank, which could contaminate the sample by putting water in contact with the atmosphere and/or allow evaporation. Detailed notes were taken as to the location of the water well, spring, or stream, when sampled, and to the configuration of the water well system in operation, noting particularly if a cistern was being used or not.

The results of stable isotope and tritium analyses of ground waters and surface waters in the Royal Gorge arch are listed in Table 4 below. The hydrogen ratios and oxygen ratios were determined using the facilities of the Institute for Water Sciences (WMU), following the procedures outlined in Appendices D and E. Enriched tritium analyses was performed by the University of Waterloo, Ontario, Canada on four water samples. The selection of the tritium samples was based on apparent differences in stable isotope and/or chemical concentrations of the sampled waters. All water sample locations are shown on Figure 2, with the exception of Station #64, which is located approximately 3 miles (5 km) up river to the west of Station #63.
<table>
<thead>
<tr>
<th>Station # (See Figure 2)</th>
<th>ΔD (±1 per mil)</th>
<th>Δ¹⁸O (±0.1 per mil)</th>
<th>Tritium (TU)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 (well)</td>
<td>-90</td>
<td>-12.1</td>
<td>—</td>
</tr>
<tr>
<td>2 (well)</td>
<td>-90</td>
<td>-12.4</td>
<td>—</td>
</tr>
<tr>
<td>3 (spring)</td>
<td>-84</td>
<td>-10.8</td>
<td>—</td>
</tr>
<tr>
<td>4 (spring)</td>
<td>+35</td>
<td>+9.8</td>
<td>—</td>
</tr>
<tr>
<td>5 (spring)</td>
<td>-96</td>
<td>-13.3</td>
<td>—</td>
</tr>
<tr>
<td>6 (spring)</td>
<td>-90</td>
<td>-12.3</td>
<td>—</td>
</tr>
<tr>
<td>7 (spring)</td>
<td>-88</td>
<td>-11.7</td>
<td>—</td>
</tr>
<tr>
<td>8 (spring)</td>
<td>-57</td>
<td>-6.0</td>
<td>—</td>
</tr>
<tr>
<td>9 (well*)</td>
<td>-97</td>
<td>-13.2</td>
<td>—</td>
</tr>
<tr>
<td>10 (well)</td>
<td>-98</td>
<td>-13.2</td>
<td>—</td>
</tr>
<tr>
<td>11 (well*)</td>
<td>-95</td>
<td>-12.9</td>
<td>—</td>
</tr>
<tr>
<td>12 (well)</td>
<td>-106</td>
<td>-14.2</td>
<td>—</td>
</tr>
<tr>
<td>13 (well)</td>
<td>-101</td>
<td>-13.4</td>
<td>—</td>
</tr>
<tr>
<td>14 (well*)</td>
<td>-100</td>
<td>-13.5</td>
<td>—</td>
</tr>
<tr>
<td>15 (well*)</td>
<td>-94</td>
<td>-12.6</td>
<td>—</td>
</tr>
<tr>
<td>16 (well)</td>
<td>-100</td>
<td>-13.4</td>
<td>—</td>
</tr>
<tr>
<td>17 (well)</td>
<td>-101</td>
<td>-13.5</td>
<td>17.2±1.3</td>
</tr>
<tr>
<td>18 (well)</td>
<td>-99</td>
<td>-13.3</td>
<td>—</td>
</tr>
<tr>
<td>19 (well)</td>
<td>-103</td>
<td>-13.2</td>
<td>—</td>
</tr>
<tr>
<td>20 (spring)</td>
<td>-105</td>
<td>-12.9</td>
<td>—</td>
</tr>
</tbody>
</table>
Table 4-Continued

<table>
<thead>
<tr>
<th>Station # (See Figure 2)</th>
<th>ΔD (±1 per mil)</th>
<th>Δ18O (±0.1 per mil)</th>
<th>Tritium (TU)</th>
</tr>
</thead>
<tbody>
<tr>
<td>21 (well)</td>
<td>-100</td>
<td>-11.6</td>
<td>13.4±1.0</td>
</tr>
<tr>
<td>22 (well)</td>
<td>-91</td>
<td>-12.4</td>
<td>1.7±0.4</td>
</tr>
<tr>
<td>23 (well*)</td>
<td>-98</td>
<td>-13.3</td>
<td>—</td>
</tr>
<tr>
<td>24 (well)</td>
<td>-100</td>
<td>-10.5</td>
<td>—</td>
</tr>
<tr>
<td>25 (spring)</td>
<td>-94</td>
<td>-9.2</td>
<td>—</td>
</tr>
<tr>
<td>26 (well*)</td>
<td>-98</td>
<td>-12.9</td>
<td>—</td>
</tr>
<tr>
<td>27 (well)</td>
<td>-108</td>
<td>-10.9</td>
<td>2.9±0.4 and 2.4±0.4</td>
</tr>
<tr>
<td>28 (well*)</td>
<td>-109</td>
<td>-10.3</td>
<td>—</td>
</tr>
<tr>
<td>29 (well)</td>
<td>-85</td>
<td>-11.8</td>
<td>—</td>
</tr>
<tr>
<td>30 (well)</td>
<td>-91</td>
<td>-12.6</td>
<td>—</td>
</tr>
<tr>
<td>31 (well*)</td>
<td>-92</td>
<td>-12.7</td>
<td>—</td>
</tr>
<tr>
<td>32 (well*)</td>
<td>-93</td>
<td>-12.9</td>
<td>—</td>
</tr>
<tr>
<td>33 (well*)</td>
<td>-102</td>
<td>-13.9</td>
<td>—</td>
</tr>
<tr>
<td>34 (stream)</td>
<td>-85</td>
<td>-11.9</td>
<td>—</td>
</tr>
<tr>
<td>35 (stream)</td>
<td>-84</td>
<td>-12.0</td>
<td>—</td>
</tr>
<tr>
<td>36 (stream)</td>
<td>-126</td>
<td>-16.7</td>
<td>—</td>
</tr>
<tr>
<td>37 (spring)</td>
<td>-29</td>
<td>-4.3</td>
<td>—</td>
</tr>
<tr>
<td>38 (stream)</td>
<td>-124</td>
<td>-16.7</td>
<td>—</td>
</tr>
<tr>
<td>39 (stream)</td>
<td>-124</td>
<td>-16.8</td>
<td>—</td>
</tr>
<tr>
<td>40 (stream)</td>
<td>-124</td>
<td>-16.7</td>
<td>—</td>
</tr>
<tr>
<td>41 (stream)</td>
<td>-115</td>
<td>-15.5</td>
<td>—</td>
</tr>
<tr>
<td>42 (stream)</td>
<td>-125</td>
<td>-16.7</td>
<td>—</td>
</tr>
</tbody>
</table>

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Table 4-Continued

<table>
<thead>
<tr>
<th>Station #</th>
<th>(\delta^D)</th>
<th>(\delta^{18}O)</th>
<th>Tritium</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(±1 per mil)</td>
<td>(±0.1 per mil)</td>
<td>(TU)</td>
</tr>
<tr>
<td>43 (stream)</td>
<td>-126</td>
<td>-16.7</td>
<td>—</td>
</tr>
<tr>
<td>44 (stream)</td>
<td>-125</td>
<td>-16.6</td>
<td>—</td>
</tr>
<tr>
<td>45 (stream)</td>
<td>-125</td>
<td>-16.7</td>
<td>—</td>
</tr>
<tr>
<td>46 (stream)</td>
<td>-68</td>
<td>-10.9</td>
<td>—</td>
</tr>
<tr>
<td>47 (stream)</td>
<td>-65</td>
<td>-10.9</td>
<td>—</td>
</tr>
<tr>
<td>62 (stream)</td>
<td>-123</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>63 (stream)</td>
<td>-123</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>64 (stream)</td>
<td>-124</td>
<td>—</td>
<td>—</td>
</tr>
</tbody>
</table>

* Water wells with cisterns or holding tanks.

Stations #1-47 were collected during the first two weeks of August 1994, whereas, Stations #62-64 were collected on May 11, 1995 by Frank Kipple of the Colorado Division of Water Resources, Pueblo, Colorado. Stations #48-61 are identified on Figure 2 as USGS isotope ratios taken from Hearne and Litke (1987).

The distribution of \(\delta^D\), \(\delta^{18}O\), and tritium values from water samples across the study area are shown in Figures 42, 43, and 44, respectively.
Figure 42. Distribution of δD Ratios in Waters at the Royal Gorge Arch. Inset Map of Royal Gorge Arch.
Waters at the Royal Gorge Arch. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Figure 43. Distribution of $\delta^{18}O$ Ratios in Waters at the Royal Gorge Arch. Inset Map Shows the Get Royal Gorge Arch.
Waters at the Royal Gorge Arch. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Figure 44. Distribution of Enriched Tritium Values From Selected Water Wells. Inset Map Shows the City of Canon and the Royal Gorge Arch.
Generalized Geology of Royal Gorge Arch

Granodiorite
Quartzofeldspathic Gneiss

Mesozoic Rocks
Paleozoic Rocks

Inset Map Shows the Generalized Geology of

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CHAPTER VI

HYDROGEOLOGIC INTERPRETATIONS

Introduction

This chapter contains interpretations of the ground water movement and ground water quality of the Royal Gorge crystalline aquifer. The chapter begins with a discussion of ground water flow directions, ground water sources, and possible ground water mixing prior to discharge into the Arkansas River. Interpretations of water quality and selected water quality statistics are then presented. The chapter concludes with a discussion of the use of slickensides as a way to determine the direction of maximum hydraulic conductivity of the aquifer.

Ground Water Movement

Ground Water Flow Directions

Ground water movement through the basement complex of the Royal Gorge arch is dominated by fracture flow. Wells drilled into the crystalline aquifer typically yield less than 10 gallons per minute, and often less than 2 gallons per minute (Appendix F). If sufficient, water-charged fractures are not intersected by a wellbore, the well is plugged and abandoned due to low yields. Because of this, the need to identify dominant directions of flow becomes critical when exploring for new sources of drinking or irrigation water in crystalline aquifers.
The ground water flow paths through the Royal Gorge crystalline aquifer are shown in Figure 45. The dominant direction of ground water flow varies between S 45-70 W (225-250 Azimuth), with recharge from the northeast near Twin Mountain. Another, smaller recharge area appears to be the Gorge Hills, north of Fremont Peak.

Concentration maps of several hydrogeochemical parameters and chemical ions measured in the ground water system (Figures 35-41) also indicate this same dominant direction of flow. Concentration "plumes" often follow flow paths that are identical to the interpreted flow lines, with concentrations generally increasing along flow paths. All of the plumes are consistent with interpretations of ground water flow to the southwest.

The dominant ground water flow direction also corresponds very closely with the interpreted direction of maximum compression and with the basement foliation direction. Although stress trajectories change orientation through the basement rocks in this part of the study area (Figure 31), they still indicate a direction of maximum compression trending N 70 E (and S 70 W). This matches the range of dominant ground water flow directions observed for the Royal Gorge crystalline aquifer.

The Mikesell Gulch fault trends about N 45 E (and S 45 W) across the northern part of the study area (Figure 2). Isotopic data indicate that ground water may be migrating along this fault zone and into the sedimentary units adjacent to the fault, mixing with the waters in the crystalline aquifer.

The above observations suggest that ground water movement through the Royal Gorge crystalline aquifer is controlled by the last stress system to affect the rocks. This stress system has apparently caused preferential aperture widening of fractures in a direction that is parallel to the maximum compressional stress and/or parallel to foliation.
Figure 45. Water Table Surface Map for the Northern Part of the Royal Gorge Crystalline Aquifer. Well Control Points are Listed in Appendix F. Inset Map Shows the Generalized Geology of the Royal Gorge Arch.
Sources of Ground Water

The $\delta D$ vs. $\delta^{18}O$ relationship is shown in Figure 46, including a comparison to the global meteoric water line (MWL) of Craig (1961a). The 47 water samples with $\delta D$ and $\delta^{18}O$ measurements fit the following linear regression equation:

$$\delta D = 6.1\delta^{18}O - 19.4$$

This equation exhibits a shift in the slope and in the constant compared to the MWL due to evaporation effects. Most of the measured isotope values do, however, fall very close to the MWL, except five spring water samples (#3, #4, #8, #20, and #25) and a cluster of four well samples (#21, #24, #27, and #28) which fall below the MWL. The remainder of the water samples (38 out of 47) fit the MWL quite closely (Figure 46), implying a geologically recent precipitation origin for these waters as their $\delta D$ vs. $\delta^{18}O$ slope is close to 8. These 38 water samples fit the following linear regression equation:

$$\delta D = 8.1\delta^{18}O + 10.3$$

The evaporation effects found in spring samples were expected, but the apparent evaporation effects observed in the four water wells were not predicted. All four of these wells are located in a small, isolated area of the crystalline aquifer, just south of the Mikesell Gulch fault and between the Arkansas River and outcrops of Cretaceous-age sedimentary units adjacent to the fault (Figure 2). The wells exhibiting apparent evaporation are also identified on Figures 42 and 43, the $\delta D$ and $\delta^{18}O$ maps, respectively.

It appears from Figure 45 that the four wells are partially separated from the major source of recharge to the northeast by a saddle in the water table surface. This
Figure 46. Graph of the $\delta D$-$\delta^{18}O$ Relationship of Water Samples Taken From the Royal Gorge Crystalline Aquifer vs. the Global Meteoric Water Line (MWL). "S" and "W" Refer to Spring and Well Samples, Respectively.
separation probably diverts ground water flow away from these wells. The source of recharge to these wells is probably direct precipitation in the immediate vicinity of the wells. Gat (1981) suggests that local variations in isotopic signature of ground waters may be due to evaporation processes before recharge, in migration through the unsaturated zone. He found that the more sluggish the movement due to limited recharge, the more chances of evaporation, interaction with the rock matrix, and/or mixing with stagnant water bodies. Gat (1981) further noted that in semi-arid zones (as in the study area) the water loss by evaporation assumes importance, shifting the resultant ground water away from the MWL along "evaporation lines." The near flat ground water gradient to the northeast of the four wells, and toward the regional recharge, is also indicative of slow moving, relatively stagnant ground waters.

The evaporation interpretation for the four wells is also supported by the reported depth to the water table in wells #28 and #24 (see Appendix F). These wells have two of the thickest unsaturated zones in the crystalline aquifer, 100 feet (30 m) and 70 feet (21 m), respectively. It can also be assumed that well #27 has a thick unsaturated zone, comparable to well #28, although no official depth was reported to the state of Colorado. The thin unsaturated zone in well #21 (25 feet or 7.5 m) however, may suggest a mixed source that is partially influenced from evaporated recharge in the vicinity of wells #27 and #28, as well #21 is directly down gradient (Figure 45). Figure 46 also shows well #21 falling between wells #27 and #28 and the MWL on the δD vs. δ18O graph, suggestive of a mixture between evaporative recharge and nonevaporative recharge.

Other possible sources for the so-called evaporated water are reused irrigation water (surface evaporation) and/or water influx associated with the Mikesell Gulch fault. The low chloride concentrations in wells #24, #27, and #28 do not support a
reused, irrigation source where enrichment in chloride would be expected from increased evaporation at the surface. The second scenario is also unlikely, as any potential ground waters flowing upward along the Mikesell Gulch fault would have to climb a substantial head gradient to reach the wells (Figure 45).

The enriched tritium measurement from well #27 has an average value of 2.7±0.4 TU (Table 4 and Figure 44). According to Fontes (1980), this value suggests limited mixing with post-1952 waters. This observation also seems to fit the aforementioned interpretation, that limited and partially evaporated recharge dominates wells #24, #27, and #28. This evaporation appears to be taking place in the unsaturated zone, prior to reaching the water table, and not from mixing with reused, surface waters (such as irrigation or pond waters).

**Mixing of Ground Waters**

Wells #11, #22, #29, #30, #31, and #32 all show a noticeable change in isotopic values, compared to the other nineteen wells in the study area. By examination of the geologic map (Figure 2), it can be seen that each of these six anomalous wells were drilled into or through Cretaceous-age sedimentary units at the surface, whereas, the remaining nineteen wells were spudded directly in the Precambrian-age crystalline aquifer. The shift in isotopic signature appears to be related to mixing from waters in the overlying sedimentary package. These six wells have a δD mean value of -91±3 per mil, whereas the other wells have a mean δD value of -100±5 per mil (Table 4 and Figure 42). The δ¹⁸O values also show a correspondingly comparable shift to more enriched values in the six wells (Table 4 and Figure 43).
The nine per mil difference in $\delta D$ value means between well groups was tested for statistical significance by a two-sample t-test with unpoled variances. The test was run with equal means for the null hypothesis and unequal means for the alternative hypothesis. The resultant p-value $= 0.0005$ suggests that there may be a significant difference in mean $\delta D$ value between these populations. The statistical significance of this test is tainted by the small number of samples, but it may be valid to assume that the water samples were in fact randomly collected from a larger population of potential wells.

According to available state well log records, three of the six wells were drilled through the Cretaceous section and completed in the crystalline aquifer, with the remaining wells completed in sandstone (wells #22, #29, and #30, Figure 2). Because chemical data from all six wells fall within regional variation (Appendices B and C), it can be assumed that the sedimentary aquifer(s) is in communication with the crystalline aquifer beneath. Only the isotopic signature shows a measureable difference between aquifers, suggesting a mixing of two different water sources is taking place.

The $\delta D$ vs. $\delta^{18}O$ diagram (Figure 46) shows both populations of waters falling on or very near the MWL, suggesting that the source of both the sedimentary aquifer(s) and the underlying crystalline aquifer is precipitation. The nine per mil difference probably reflects a difference in elevation or temperature between the two precipitation sources. According to the water table map (Figure 45), the regional recharge area for both the sedimentary aquifer(s) and the crystalline aquifer appears to be to the northeast, up-gradient. If this is the case, both aquifers should have the same isotopic signature. Because they do not however, it must be assumed that the sedimentary units are getting at least some of their water from another source.
Because the sedimentary units are everywhere in contact with the Mikesell Gulch fault (Figure 2), it is suggested that another source of water may be influx along the fault, feeding the sedimentary rocks. Although this water appears to be from precipitation, it is likely from a different source area (lower elevation) and from a different ground water flow system. The water moves into the sedimentary units as there is no hydraulic gradient preventing such migration as noted earlier for part of the crystalline aquifer adjacent to the fault (Figure 45). Wells drilled through the sedimentary units and into the crystalline aquifer then pick up a mixed isotopic signature from the two different sources.

A water sample from well #22, which was completed in the sedimentary aquifer(s), was sent to the University of Waterloo for enriched tritium analysis. The result was a measurement of 1.7±0.4 TU, suggesting limited mixing with post-1952 waters (Fontes, 1980). This tritium measurement is quite low compared to another determination at well #17, about two miles to the east of well #22. Well #17 has an enriched tritium measurement of 17.2±1.3 TU, even though both wells are about the same distance from the regional recharge area to the northeast (Figures 44 and 45), the only difference being that well #17 has no sedimentary units on top of the crystalline aquifer. Well #21 also had an enriched tritium measurement of 13.4±1.0 TU, again, considerably higher than well #22. According to Fontes (1980), these higher tritium measurements imply recharge since the 1960's.

It is suggested, therefore, that recent recharge is occurring to the northeast, near Twin Mountain, and is migrating to the southwest, down-gradient, progressively mixing high tritium waters with pre-1950's waters and giving tritium measurements in the teens (wells #17, and #21). The low tritium measurements in wells #22, and #27, by contrast, imply limited recharge since the 1950's, although for different reasons.
Water from well #22 is interpreted to be dominantly from a pre-1950's source associated with the Mikesell Gulch fault, whereas water from well #27 is interpreted to be partially isolated from regional recharge, and is dominated by pre-1950's water with only a slight mixing effect from local precipitation.

**Ground Water Interactions With the Arkansas River**

The isotopic signature for the Arkansas River is remarkably consistent across the study area (see Figures 42 and 43, and Table 4). River samples taken in August 1994 and in May 1995 also show nearly identical isotopic ratios. The water in the Arkansas River is interpreted to be dominated by discharge from outside the study area, and exhibits depleted ratios from both higher elevations (snow-melt) and corresponding cooler temperature precipitation, giving an average δD of -124 per mil and an average δ¹⁸O of -16.6 per mil.

The source waters for the Arkansas River originate about 80 miles (130 km) northwest of the study area, near Leadville, Colorado. Over this distance, the river elevation changes from about 6000 feet (1830 m) above sea level near Canon City to about 10,000 feet (3050 m) above sea level near Leadville, a difference of 4000 feet (1220 m) in altitude. According to Yurtsever and Gat (1981), a typical altitude effect gradient for δD is between 1.5-4.0 per mil/100 m. Using 2.5 per mil/100 m as a typical altitude effect value, it can be shown that a 1220 m change in elevation should yield about a 30.5 per mil depletion in δD compared to precipitation at Canon City. This calculation assumes that the continental effect over the 80 miles (130 km) from Canon City to Leadville is negligible.

The average δD value for ground water within the Royal Gorge crystalline aquifer was found to be -98±6 per mil. It can be assumed that this is an average value.
for local precipitation at 6000 feet (1830 m) above sea level. The altitude effect calculated above yields an average δD value of -129 per mil for precipitation falling at 10,000 feet (3050 m) above sea level (-98 - 31 = -129). Because the Arkansas River near Canon City has an average δD value of -124±3 per mil, it is suggested that discharge in the Arkansas River (at Canon City) is dominated by precipitation from close to 10,000 feet (3050 m) above sea level.

The isotope ratios from Grape Creek are also believed dominated by discharge from outside the study area (see water samples #34, #35, #46, and #47, Figure 2, and Table 4), although at a lower elevation than the source of the Arkansas River.

The consistent isotope ratios along the Arkansas River imply little influence of ground water baseflow into the river. Mixing of isotope ratios between ground water and river water has been examined by Mook (1970), who was able to estimate the seasonal fraction of snow-melt and baseflow into the Rhine River, knowing the two end-member components. This is not possible in the present study even though the Arkansas River and the baseflow end-member components can be estimated, as there is no measurable contribution to the river isotopic signature across the Royal Gorge arch. The average δD value for the Arkansas River is -124±3 per mil, while the average δD value for the ground water from wells in the crystalline aquifer is -98±6 per mil, a difference of over 25 per mil. This difference is quite significant considering the laboratory precision of δD values is ±1 per mil. The very consistent conductivity measurements (Appendix B) of the Arkansas River (mean 239±35 µS/cm) also support the conclusion that very little ground water is discharging into the river as baseflow, compared to the snow-melt discharge.

There is also no measurable isotopic change in the Arkansas River at the confluence with Grape Creek near Canon City (see Figure 2). This again implies that
the discharge of the Arkansas is substantial compared to Grape Creek, which has an average δD value of -76±11 per mil.

The water table surface map of the crystalline aquifer does indicate a substantial gradient toward the Arkansas River, implying baseflow inflow is occurring (Figure 45), but not in quantifiable amounts. This large drop in hydraulic head, adjacent to the river, also precludes significant influx of river water into the aquifer. It is concluded therefore, that the indeterminate percentage of baseflow into the Arkansas River is overshadowed by snow-melt discharge from higher elevations (10,000 feet or 3050 m), even during the seasonally low discharge period (late summer) when many samples were taken (August, 1994).

The above conclusions are supported by discharge data from the upstream gaging station at Parkdale, Colorado (near the intersection of the Mikesell Gulch fault with the river, Figure 2), and the downstream gaging station at Canon City (just past the confluence with Grape Creek, Figure 2). Table 5 shows recorded values of mean discharge at each station, on a monthly basis, for one annual cycle (October 1992-September 1993). The data show a substantial loss in discharge of the Arkansas River as it passes through the Royal Gorge crystalline aquifer. This loss is substantial even though Grape Creek discharges into the Arkansas River between the gaging stations. The Arkansas River however, is not interpreted to be a losing stream as it passes through the crystalline aquifer. Instead, it is apparently losing discharge because water is being diverted from the river between gaging stations by the city of Canon City and several ungaged private ditches. According to Frank Kipple of the Colorado Division of Water Resources, there is insufficient data available on the private ditches to calculate a meaningful water balance between gaging stations (personal communication). For this reason, a water balance was not attempted on this segment.
Table 5

Monthly Mean Discharge of the Arkansas River at Parkdale, Colorado and at Canon City, Colorado From October 1992 to September 1993

<table>
<thead>
<tr>
<th>Month</th>
<th>Mean Discharge at Parkdale in Cubic Feet per Second</th>
<th>Mean Discharge at Canon City in Cubic Feet per Second</th>
</tr>
</thead>
<tbody>
<tr>
<td>October</td>
<td>381</td>
<td>259</td>
</tr>
<tr>
<td>November</td>
<td>471</td>
<td>373</td>
</tr>
<tr>
<td>December</td>
<td>508</td>
<td>449</td>
</tr>
<tr>
<td>January</td>
<td>503</td>
<td>461</td>
</tr>
<tr>
<td>February</td>
<td>561</td>
<td>506</td>
</tr>
<tr>
<td>March</td>
<td>736</td>
<td>678</td>
</tr>
<tr>
<td>April</td>
<td>429</td>
<td>373</td>
</tr>
<tr>
<td>May</td>
<td>1606</td>
<td>1449</td>
</tr>
<tr>
<td>June</td>
<td>2979</td>
<td>2742</td>
</tr>
<tr>
<td>July</td>
<td>2026</td>
<td>1833</td>
</tr>
<tr>
<td>August</td>
<td>761</td>
<td>628</td>
</tr>
<tr>
<td>September</td>
<td>580</td>
<td>487</td>
</tr>
</tbody>
</table>

of the Arkansas River. Based on the high ground water gradient into the river (Figure 45), and the static isotopic data, it can be concluded that water is discharging into the river from the crystalline aquifer. However, because the discharge in the river overwhelms the ground water discharge, no meaningful, quantifiable value can be determined from the available data.
Ground Water Quality

Ground Water Quality Interpretations

Figures 35-41 (Chapter V) show the distributions of various water quality measurements across the Royal Gorge crystalline aquifer. The following section details specific conclusions of each parameter as listed previously in Appendices B and C (Chapter V).

The alkalinity of waters within the Royal Gorge crystalline aquifer is interpreted to be primarily the result of bicarbonate concentration, as the pH within the aquifer is consistently less than 9. The dominant reactions which determine alkalinity begin with the carbonate equation below:

\[ H_2O + CO_2 = H_2CO_3 = H^+ + HCO_3^- \]

where \( CO_2 \) input is predominantly from the soil zone on top of the aquifer (see the pH discussion). Bicarbonate can be also produced during the dissolution of calcite according to the following reaction:

\[ CaCO_3 + H_2O + CO_2 = Ca^{2+} + 2HCO_3^- \]

Because calcite was observed within fractures during thin-section analysis, it seems likely that both of these reactions have maintained the alkalinity at a fairly constant level within the aquifer (150-300 mg/l as CaCO_3). The ground waters have a higher alkalinity than river waters as they are able to pick up dissolved carbon dioxide from within the soil zone, whereas river waters equilibrate with lower, atmospheric levels of carbon dioxide and alkalinity remains slightly lower (58-186 mg/l as CaCO_3).
The two water samples with higher than average alkalinites also have fairly high hardness levels (Stations #1 and #25, Appendix B), indicating high levels of dissolved calcium and/or magnesium. These two samples are probably more strongly influenced by the second reaction, producing bicarbonate and dissolved calcium in tandem.

Conductivity measurements indicate that the highest concentration of dissolved ions are in the springs, followed by the wells and streams, respectively. These results were expected as several of the springs had experienced evaporation effects (see following isotopic evidence), which tended to concentrate dissolved solids. In addition, much of the water in the springs and wells probably travelled substantial distances through the aquifer, thereby increasing the mineral reaction potential and increasing the ionic concentration. The lower conductivities of the stream and river samples can also be explained using the same rationale, where reactions with rock matrices are minimal compared to runoff volume.

The mineralogies present in the Royal Gorge crystalline aquifer (see Appendix A) are all consistent with the hardness reactions given above. Micas, feldspars, gypsum, and clays are all present in sufficient amounts to allow the reactions to proceed in the subsurface. The balance between the creation of calcium and magnesium ions, and their destruction through calcium carbonate precipitation and cation exchange determines the total hardness of the waters in the aquifer.

By comparing the distribution of total hardness (Figure 37) to the water table map (Figure 45), it appears that the lowest hardness values are near zones of recharge. The measured hardness values also seem to follow linear trends which are parallel to the direction of ground water movement. These zones of higher hardness are probably related to either local increases in the production of calcium and magnesium ions, or to
local increases in the cation exchange rate. Both can be directly related to minor variations in aquifer mineralogy. The measured hardness values for the river and stream samples are considerably lower than the well and spring samples. The wells have a mean hardness of $498 \pm 227$ mg/l as CaCO$_3$, and the springs a mean hardness of $688 \pm 537$ mg/l as CaCO$_3$. Whereas, the stream samples had mean hardnesses of less than 150 mg/l as CaCO$_3$. This relationship is expected as the stream samples are dominated by snow-melt (see the following isotopic evidence) and runoff, with limited opportunity for rock interactions and associated mineral reactions. The ground water samples, by contrast, have higher hardnesses as they react with the minerals in the aquifer during migration away from the recharge areas.

The fairly high conductivities, coupled with the high sulfate concentrations in the ground water, suggest that dissolution of gypsum is a major factor in determining the observed hardnesses. This reaction produces only calcium and sulfate ions, and not bicarbonate, and may account for the fact that hardness is generally greater than alkalinity in ground waters.

The measured pH values within the Royal Gorge crystalline aquifer range from 6.92 to 8.93 (Appendix B), and are consistent with the ranges of pH from shallow granitic terrains reported by Freeze and Cherry (1979), and Gascoyne and Kamineni (1994). The areal distribution of pH values for the crystalline aquifer is shown in Figure 38. The dominance of feldspars and other silicate minerals appears to be controlling the pH of waters within the aquifer through the dissolution reactions reported above.

Figure 39 shows chloride concentration roughly paralleling the direction of ground-water movement, following approximately linear flow paths from recharge zone to discharge zone. The lowest ground-water chloride concentrations are again
nearest the recharge zones, with a slight increase in concentration along the flow paths.

Iron content of the water samples is probably a function of more reducing conditions locally within the aquifer system, and secondly, a function of pyrite, biotite, and magnetite concentrations. The following is an example of an iron oxide half-reaction which increases the dissolved iron content of ground waters and lowers the redox potential (Kehew and Brewer, 1992):

\[
\text{Fe}_2\text{O}_3 + 6\text{H}^+ + 2e^- = 2\text{Fe}^{2+} + 3\text{H}_2\text{O}
\]

This half-reaction requires an electron donor such as organic carbon from the soil zone and contributes to reducing conditions. Isolated concentrations of iron-rich minerals also likely contribute to an increase in iron concentration beyond background (less than 1.0 mg/l).

Most of the nitrate is clustered within three plumes (Figure 41), with the outer contours equal to the 5 mg/l concentration cut-off level. On the scale of the overall crystalline aquifer, these contaminants approximate point sources of nitrates that probably migrate and disperse to the southwest under advective flow. Although the two smaller plumes show only a single value exceeding 5 mg/l, their shapes are constrained by lower concentrations, allowing separation of the plumes.

Only two well samples equal or exceed the EPA Maximum Contaminant Level for nitrate as nitrogen (Samples #10 and #12). Three other well samples approach but do not exceed the MCL (10 mg/l), at 9.5 mg/l, 9.0 mg/l, and 9.0 mg/l nitrate as nitrogen (Samples #14, #11 and #31, respectively). All of these high concentrations fit into the aforementioned plumes on Figure 41. The sources of the plumes are interpreted as originating from either horse stables (manure) along Highway 50, or else...
from septic systems due to closely-spaced housing developments (or a combination of both). The often thin to non-existent soil zone (direct rock outcrops) on top of the aquifer makes efficient septic construction especially difficult.

Denitrification, defined as the process which breaks nitrate down into nitrogen oxide or nitrogen gas (Stuart and others, 1995), is enhanced by acidic soils, moist soils, and highly organic soils (Broadbent and Clark, 1967). In the case of the Royal Gorge crystalline aquifer, the soil conditions do not generally fit these criteria as they are not particularly moist and not very organic rich. Therefore, denitrification cannot be expected to be a very efficient process in the crystalline aquifer. The elongated plumes of nitrate that are observed indicate that only limited denitrification is taking place during transport. Denitrification may be expected however, in areas of high dissolved iron concentration, where conditions may be slightly more acidic, and dissolved iron and nitrate should not co-exist. Figures 40 and 41 show low concentrations of nitrate in areas of high concentrations of dissolved iron.

Chilton and Foster (1995) interpreted high concentrations of sulfate in their study as the result of oxidation of pyrite and pyrrotite within basement gneisses. Moran and others (1978) noted that pyrite oxidation is one of the most important acid-producing reactions, contributing H\(^+\) ions to the ground water system. Fontes and others (1989) also interpreted some of the sulfate in ground waters within the Stripa mine as being derived from reactions with pyrite. Gascoyne and Kamineni (1994) interpreted their observed sulfate concentrations as being from oxidation of sulfide minerals, but added that reactions with gypsum-infilling minerals in fractures was also important. Examples of these reactions are shown below (Freeze and Cherry, 1979):

\[
\text{FeS}_2 + \frac{15}{4} \text{O}_2 + \frac{7}{2} \text{H}_2\text{O} = \text{Fe(OH)}_3 + 4\text{H}^+ + 2\text{SO}_4^{2-}
\]

Pyrite \hspace{1cm} Sulfate
Small amounts of pyrite are interpreted to react with ground water flowing through the crystalline aquifers to create the sulfate ion, as shown in the first reaction above. The dissolution of andesine and other silicate minerals releases Ca\(^{2+}\), which then combines with sulfate and precipitates gypsum on the fracture surfaces, as shown in the second reaction. Freeze and Cherry (1979) report that the combined effects of infiltration, pyrite oxidation, calcite dissolution, and subsequent evapotranspiration cause gypsum precipitation at shallow depths.

Domenico and Schwartz (1990) state that within the Great Plains region of the United States annual potential evaporation exceeds annual precipitation by a considerable amount. The Royal Gorge arch sits on the edge of this physiographic region and therefore, likely has a similar water budget. Normal rain and snow-melt events probably experience high evaporation, depositing gypsum prior to reaching the water table. Exceptional recharge events flowing through the fractures then redissolve some of the deposited gypsum, reversing the reaction, and create high concentrations of sulfate in the shallow subsurface. Concentrations of over 5000 mg/l sulfate have been documented in recharge waters in arid climates (Hendry and others, 1986). Freeze and Cherry (1979) report that for sulfate to occur as a dominant ion, only very small amounts of gypsum need to be dissolved.

The above reactions are interpreted to be the cause of elevated sulfate concentrations within the Royal Gorge crystalline aquifer. Although only small amounts of pyrite are present for sulfide dissolution (Appendix A), the adjacent, Jurassic Ralston Creek Formation contains sufficient gypsum for precipitation and redissolution reactions. The semi-arid climate probably contributes to the precipitation.
of gypsum, and redissolution during the limited rainfall events. River and stream waters, by contrast, do not pass through sufficient rock matrix to produce high sulfate concentrations, and remain lower in sulfate.

**Ground Water Quality Statistical Interpretations**

Because chloride is often cited as a conservative ion, linear regression analysis was run on each of the hydrogeochemical parameters (except sulfate), including $\delta$D, using chloride as the independent variable. The basic hypothesis is to determine if chloride concentration had statistical use in predicting any of the other hydrogeochemical parameters. If the assumption is made that increased evaporation leads to an increase in chloride content, there may exist a linear relationship between selected geochemical parameters and chloride concentration that is indicative of evaporation. For example, as $\delta$D becomes more positive (enriched), a corresponding increase in chloride concentration may imply partial evaporation has occurred.

Plots of each of the aforementioned geochemical parameters (except temperature, sulfate, and $\delta^{18}$O) versus chloride concentration are shown in Figures 47-53. The statistical results and plots were generated using a statistical package by Minitab (1992). All data from Appendices B and C were used in the analyses. The number of samples in each test was 47, except in the cases of nitrate and iron which had 45 and 44 samples, respectively.

The results indicate that there is statistical significance ($\alpha = 0.05$) to predict alkalinity, conductivity, hardness, nitrate, and pH, knowing chloride concentration. There is not a statistically valid relationship however, between chloride and iron, and chloride and $\delta$D. Although statistical significance is identified for five of the parameters, examination of the plots (Figures 47-53) suggests that some of these
Figure 48. Plot of Alkalinity (as CaCO₃) Values vs. Chloride Concentration for Water Samples Taken From the Royal Gorge Crystalline Aquifer.
Figure 49. Plot of Conductivity Values vs. Chloride Concentration for Water Samples Taken From the Royal Gorge Crystalline Aquifer.
Figure 50. Plot of Hardness (as CaCO₃) Values vs. Chloride Concentration for Water Samples Taken From the Royal Gorge Crystalline Aquifer.
Figure 51. Plot of pH Values vs. Chloride Concentration for Water Samples Taken From the Royal Gorge Crystalline Aquifer.
Figure 52. Plot of Iron Concentration vs. Chloride Concentration for Water Samples Taken From the Royal Gorge Crystalline Aquifer.
Figure 53. Plot of Nitrate (as Nitrogen) Concentration vs. Chloride Concentration for Water Samples Taken From the Royal Gorge Crystalline Aquifer.
relationships are less strong than others. Conductivity vs. chloride and hardness vs. chloride show the least amount of scatter on the plots, implying stronger linear relationships.

The "R-squared" term is an approximate measure of how much variation within the dependent variable is explained by knowing the independent variable (chloride). The R-squared terms for alkalinity, conductivity, hardness, pH, and nitrate (vs. chloride) are 35.9%, 81.9%, 79.6%, 14.6%, and 11.5%, respectively. These values support the above observation that there is variability in the strength of the linear relationships. An R-squared term >50% can be arbitrarily chosen as indicative of a stronger linear relationship. The high R-squared terms for conductivity and hardness imply that these geochemical parameters are best predicted knowing chloride concentration. A good correlation between conductivity and chloride was also well documented by McDonald and Partners (1965) for Lower Indus ground waters, and Bugg and Lloyd (1976) for ground waters from the Cayman Islands.

The lack of a significant relationship between chloride concentration and δD implies that chloride concentration is not simply related to evaporation. Because there are some evaporative effects observed in the isotope data which are attributed to evaporation in springs (or evaporation prior to reaching the water table), a two-sample t-test was also run on chloride concentration vs. δD values for just the spring samples (9 samples). The results of this analysis, again, showed no significant relationship between chloride concentration and δD values (α = 0.05, p-value = 0.534). This suggests that there is no predictable chloride increase with isotopic enrichment, even in the springs.

The lack of a correlation between iron and chloride supports the earlier conclusion that iron concentration is primarily a redox phenomenon. Iron is probably
influenced by local reducing conditions, as opposed to following systematic relationships with other ions.

Statistically significant correlations ($\alpha = 0.05$) between chloride and alkalinity, conductivity, hardness, pH, and nitrate imply that relationships exist between these parameters. As was noted above, alkalinity, pH, and hardness are all tied either directly or indirectly to the carbonate system. Consistent relationships between these parameters is, therefore, not surprising. Generally, chloride, alkalinity, and hardness show increases in concentration away from recharge areas and along flow lines. The resulting increase in TDS along flow lines also leads to a systematic increase in conductivity away from recharge areas. The pH is so strongly tied to alkalinity through chemical reactions that it too shows a positive correlation to chloride content (even though the R-squared term is lower).

The statistically significant correlation between nitrate and chloride is also not unexpected, although for a different reason. Both nitrate and chloride concentrations generally show increases in septic tank effluent. Because the nitrate concentrations were interpreted to be due to septic systems and/or horse manure, it follows that there is a correlation between increasing chloride and increasing nitrate.

Use of Slickensides to Determine Maximum Hydraulic Conductivity Direction

One of the more significant findings of this study is the suggestion of a new method of determining the direction of maximum hydraulic conductivity in fractured aquifers using slickenside arrays. The recognition of a preferred direction of maximum aperture in a fractured aquifer remains an important hydrogeologic consideration because flow through a fracture is proportional to the cube of the aperture width. It is
generally assumed, in a brittle deformational environment, that the fracture direction with the widest aperture parallels the direction of maximum stress.

Previous studies, including many recent publications (Boeckh, 1992; Greenbaum, 1992; Mabee and others, 1994; Sidle and Lee, 1995), have relied upon fracture analysis from air or space imagery to determine stress directions affecting fractured aquifers. This methodology has met with varying degrees of success as many of the studied crystalline aquifers have experienced up to four tectonic events (Greenbaum, 1992), complicating the interpretation of the last active stress system.

The technique presented in the current study, of using slickenside arrays to identify the last operative stress field, appears to alleviate much of the fracture analysis "averaging" from multiple tectonic events. Joints and fractures from earlier tectonism may become reactivated in a subsequent event, but fracture apertures are assumed to open or close differentially according to the last stress system. For this reason, it is imperative in any hydrogeologic study to first identify the most recent stress system that influenced the fractures, prior to predicting avenues of higher fracture flow. The present stress field as reported by Zoback (1992) should also be examined. Previously open fractures may be affected if the present system has a direction of maximum compression that is perpendicular to the fracture orientation.

In the Royal Gorge arch, the direction of Laramide maximum compressional stress was first determined from slickenside arrays using the method of Aleksandrowski (1985), on a domain by domain basis, and then applied to the entire aquifer through the construction of stress trajectories. The slickensides measured in the field were assumed to represent the last penetrative "jostling" of the rocks, regardless of when the fractures initially developed. All field data (water table measurements, isotopic, and chemical data) are consistent with the interpretation that
ground water in the Royal Gorge crystalline aquifer is preferentially migrating along fractures parallel to the Laramide direction of maximum compression, as determined from the structural analysis. The present stress system, although not well constrained in the study area (Zoback, 1992), does not appear to deviate significantly from the Laramide system. Results indicate that the direction of maximum hydraulic conductivity of a fractured aquifer is definable with this technique.
CHAPTER VII

CONCLUSIONS

1. The Royal Gorge arch is subdivided into 10 homogeneous structural domains based on fabric analysis. The three sedimentary domains were positioned during Laramide uplift of the arch in response to heterogeneities in the underlying basement. The seven basement domains appear to have been strongly influenced by pre-existing fabric and lithologic differences, and were later modified during Laramide fold development.

2. Joint patterns in the crystalline aquifer also appear to be strongly controlled by fabric orientations and lithology, and exhibit identifiable differences from domain to domain. Jointing in the granitic gneiss often follows foliations and is higher in density as compared to joint spacing in the granodiorite. Jointing in the granodiorite also follows foliations, but only in areas where the foliations are well developed.

3. Slickenside arrays were used to determine the last operative stress field which affected the uplift. Stress trajectories were fairly uniform in the surrounding sedimentary rocks, but apparently refracted considerably through the basement domains under the strong influence of fabric and lithology variations.

4. Overall development of the Royal Gorge arch can be approximated by a basement fault-propagation fold. The principal fault ramps up between 13-20 degrees off of a deep detachment surface, but remains well within the basement.

5. Ground water reactions within the fractured rock matrix of the Royal Gorge crystalline aquifer appear to dominate the hydrogeochemistry. The chemical
reactions are driven by the particular mineralogy that is present locally. Weathering of feldspars and other silicates to clays, in combination with the CO₂-carbonate system, determines the range of pH, alkalinity, and hardness for waters in the aquifer.

6. High levels of sulfate in the ground water are attributed to gypsum precipitation and redissolution during recharge events. Nitrate plumes appear to be near-point sources emanating from closely spaced septic systems and/or horse manure concentrations at riding stables.

7. Hydrogeochemical parameters appear to follow ground water flow lines as interpreted from the water table surface map. Areas of ground water recharge often have the lowest concentrations of chemical parameters, whereas, the highest concentrations are in areas of discharge or down gradient from recharge areas.

8. Ground waters in the Royal Gorge crystalline aquifer appear to be dominated by precipitation as most of the stable isotopic data fall on or near the meteoric water line (MWL).

9. Isotopic data suggest that baseflow from the Royal Gorge crystalline aquifer is negligible compared to run-off from outside the study area, with the source from close to 10,000 feet (3050 m) above sea level (snow-melt).

10. Isolated areas of the crystalline aquifer have experienced higher than normal evaporation effects and may be partially separated from the dominant ground water flow. Other areas adjacent to the Mikesell Gulch fault have experienced mixing with other ground water sources.

11. Limited tritium data indicate recharge has occurred in most of the crystalline aquifer since the 1960's.

12. Ground water flow lines generally trend S 45-70 W, with recharge from the northeast near Twin Mountain. This range in flow path direction is nearly identical
to the interpreted direction of maximum horizontal compression for the crystalline aquifer. This direction also parallels the elevation gradient of the arch, which largely resulted from the Laramide stresses. Ground water is apparently following fractures, that may be preferentially enhanced by the last penetrative stress system, down the elevation gradient.

13. A new method of determining the direction of maximum hydraulic conductivity in fractured aquifers using slickenside arrays is suggested. The last, maximum, compressional stress direction was defined by the field collection of slickenside data, followed by stress interpretation using the method of Aleksandrowski (1985), and finally, construction of stress trajectories across the aquifer. Analysis of the Royal Gorge crystalline aquifer demonstrates that a preferred direction of maximum hydraulic conductivity is definable with this technique.

14. Detailed analysis of the structural framework is a prerequisite for hydrogeologic study of fractured aquifers. Researchers can no longer approach the study of fractured media from a purist viewpoint. The complexities of fluid flow through fractures demand an integration of structural and geochemical data with hydrogeological data. This study is an example of the integration of several geologic subdisciplines that are focused to better understand ground water movement through a large-scale, fractured aquifer.
Appendix A

Microscope Thin Section and Point-Count Data
The following data were generated by petrographic analysis of 31 thin sections collected from the study area. All locations are shown on Figure 2, and identified by the appropriate map location number. The mineral percentages were visually estimated on each thin section. Five of the thin sections were point-counted as a check of the estimates (specimen numbers: 23-93, 54-93, 56-93, 193-93, 503-93). The differences between the estimated values and the point-counted values were consistently less than 5%. Any differences were resolved by adjusting the estimated values up or down to match the point-counted values.

**Map Location #1:**
Specimen Number: 18-93
Medium-brown, fine-grained, hypidiomorphic, diabasic, augite-gabbro-norite, with color index=24.
Plagioclase feldspar (An$_{53}$) 75%
Augite 12%
Magnetite 7%
Biotite 5%
Chlorite Trace
Hematite Trace

**Map Location #2: Point-counted.**
Specimen Number: 23-93
Red-brown, medium-grained, hypidiomorphic, granular, biotite-granodiorite, with color index=34.
Biotite 29%
Quartz 29%
Plagioclase feldspar (An$_{24}$) 28%
Microcline 8%
Sphene 2%
Sericite 2%
Apatite Trace
Chlorite Trace
Hornblende Trace
Muscovite Trace
Zircon Trace
Map Location #3:
Specimen Number: 26-93
Light-pink, medium-grained, allotriomorphic, granular, biotite-granite, with color index=5.
Quartz \(40\%\)
Plagioclase feldspar (An\(_{21}\)) \(30\%\)
Microcline \(25\%\)
Biotite \(5\%\)
Hematite Trace
Magnetite Trace
Sericite Trace
Zircon Trace

Map Location #4:
Specimen Number: 46-93
Dark-brown, fine-grained, hypidiomorphic, diabasic, augite-gabbro-norite, with color index=45.
Plagioclase feldspar (An ?) \(55\%\)
Augite \(43\%\)
Magnetite \(2\%\)
Calcite (in fractures) Trace
Hematite Trace
Muscovite Trace
Sericite Trace

Map Location #5: Point-counted.
Specimen Number: 54-93
Dark-red-brown, plagioclase feldspar-biotite-hornblende-quartz-gneiss, with 1 mm magnetite porphyroblasts in a fine-grained matrix.
Quartz \(26\%\)
Hornblende \(25\%\)
Biotite \(24\%\)
Plagioclase feldspar (An\(_{38}\)) \(22\%\)
Magnetite \(2\%\)
Garnet \(1\%\)
Apatite Trace
Calcite (in fractures) Trace
Clay alteration (in fractures) Trace
Hematite Trace
Microcline Trace
Sericite Trace
Zircon Trace

Map Location #6:
Specimen Number: 55-93
Dark-green, quartz-plagioclase feldspar-hornblende-amphibolite, with 3 mm porphyroblasts of hornblende.
Hornblende 60%
Plagioclase feldspar (An$_{70}$) 25%
Quartz 10%
Biotite 3%
Apatite Trace
Magnetite Trace
Pyrite Trace
Sericite Trace
Sphene Trace

Map Location #7: Point-counted.
Specimen Number: 56-93
Light-red-brown, muscovite-biotite-sillimanite-quartz-schist, with wavy bands of biotite, sillimanite and quartz around porphyroblasts of muscovite.
Quartz 36%
Sillimanite 28%
Biotite 20%
Muscovite 12%
Magnetite 3%
Clay alteration Trace
Hematite Trace
Microcline Trace
Orthoclase Trace
Sericite Trace
Zircon Trace

Map Location #8:
Specimen Number: 129-93
Light-gray, biotite-plagioclase feldspar-microcline-quartz-gneiss, with 3-5 mm quartzofeldspathic-rich layers and 2 mm porphyroblasts of magnetite.
Quartz 40%
Microcline 25%
Plagioclase feldspar (An$_{31}$) 20%
Biotite 13%
Magnetite 2%
Hematite Trace
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<th>Specimen Number 144-93</th>
<th>Specimen Number 163-93</th>
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<td>Zircon</td>
<td>Trace</td>
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<td>Trace</td>
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**Map Location #9:**
Specimen Number: 143-93
Dark-green, medium-grained, hypidiomorphic, porphyritic, biotite-granodiorite, with color index=46.

Plagioclase feldspar (An\textsubscript{45}) 27%
Biotite 25%
Hornblende 20%
Quartz 18%
Orthoclase 7%
Magnete Trace
Sphene 1%
Apatite Trace
Sericite Trace
Zircon Trace

**Map Location #10:**
Specimen Number: 144-93
Pink, fine-grained, hypidiomorphic, granular, granite, with color index=1.

Quartz 45%
Plagioclase feldspar (An\textsubscript{24}) 30%
Orthoclase 23%
Biotite 1%
Hematite Trace
Magnete Trace
Muscovite Trace
Sericite Trace
Zircon Trace

**Map Location #11:**
Specimen Number: 163-93
Pink, medium-grained, hypidiomorphic, granular, biotite-granite, with color index=3.

Quartz 40%
Orthoclase 27%
Plagioclase feldspar (An\textsubscript{24}) 25%
Biotite 3%
Microcline 3%
Muscovite 1%
Hematite Trace
Magnete Trace
Sericite Trace
Sphene Trace
Zircon Trace

Map Location #12:
Specimen Number: 177-93
Pink, medium-grained, hypidiomorphic, porphyritic, granite, with color index=1.
Microcline 35%
Quartz 35%
Plagioclase feldspar (An$_{24}$) 30%
Biotite Trace
Hematite Trace
Sericite Trace

Map Location #13:
Specimen Number: 183-93
Gray-tan, quartz-plagioclase feldspar-microcline (granitic) cataclasite, with 10-25 mm microcline porphyroclasts and some quartz porphyroclasts.
Microcline 40%
Plagioclase feldspar (An$_{25}$) 30%
Quartz 25%
Calcite (in veins) 3%
Chlorite 2%
Biotite Trace
Hematite Trace
Sericite Trace

Map Location #14:
Specimen Number: 192-93
Pink, medium-grained, hypidiomorphic, granular, biotite-granite, with color index=10.
Quartz 40%
Microcline 20%
Orthoclase 15%
Plagioclase feldspar (An$_{23}$) 15%
Biotite 9%
Sericite Trace
Zircon Trace
Map Location #15: **Point-counted.**
Specimen Number: 193-93
Pink, fine-grained, hypidiomorphic, granular, biotite-alkali feldspar granite, with color index=15.
Microcline 46%
Quartz 33%
Biotite 15%
Plagioclase feldspar (An\textsubscript{15}) 5%
Hematite Trace
Muscovite Trace
Zircon Trace

Map Location #16:
Specimen Number: 212-93
Light-gray-pink, plagioclase feldspar-quartz-microcline-mylonitic granite, with 8-10 mm porphyroclasts of microcline and quartz, and well-developed mylonitic texture.
Microcline 51%
Quartz 40%
Plagioclase feldspar (An ?) 8%
Biotite Trace
Chalcopyrite Trace
Chlorite Trace
Hematite Trace
Magnetite Trace
Muscovite Trace
Sericite Trace
Zircon Trace

Map Location #17:
Specimen Number: 284-93
Tan, plagioclase feldspar-quartz-gneiss, with 5-15 mm thick porphyroblastic-rich zones of quartz, plagioclase and microcline.
Quartz 72%
Plagioclase feldspar (An\textsubscript{21}) 20%
Biotite 3%
Microcline 3%
Muscovite 1%
Chlorite Trace
Hematite Trace
Magnetite Trace
Sericite Trace
Zircon Trace

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Map Location #18:
Specimen Number: 300-93
Medium-gray, epidote-plagioclase feldspar-biotite-quartz-gneiss, with some flaser texture.
Quartz 60%
Biotite 15%
Plagioclase feldspar (An$_{31}$) 15%
Epidote 5%
Magnetite 1%
Chlorite Trace
Microcline Trace
Zircon Trace

Map Location #19:
Specimen Number: 335-93
Green-gray, medium-grained, hypidiomorphic, porphyritic, orthoclase-plagioclase-bearing-hornblendeitite, with color index=60.
Hornblende 60%
Orthoclase 35%
Plagioclase feldspar (An$_{52}$) 5%
Magnetite Trace
Muscovite Trace
Sericite Trace
Zircon Trace

Map Location #20:
Specimen Number: 352-93
Red-brown, muscovite-biotite-plagioclase feldspar-quartz-gneiss.
Quartz 65%
Plagioclase feldspar (An$_{40}$) 20%
Biotite 10%
Muscovite 3%
Magnetite 1%
Apatite Trace
Hematite Trace
Microcline Trace
Zircon Trace
Map Location #21:
Specimen Number: 354-93
Dark-green, plagioclase feldspar-hornblende-biotite-quartz-gneiss, with 3 mm poikiloblasts of hornblende.

Quartz 45%
Biotite 35%
Hornblende 10%
Plagioclase feldspar (An25) 8%
Sphene 1%
Apatite Trace
Magnetite Trace
Zircon Trace

Map Location #22:
Specimen Number: 356-93
Olive-green, medium-grained, hypidiomorphic, porphyritic, chlorite-tonalite, with color index=22.

Plagioclase feldspar (An36) 40%
Quartz 35%
Chlorite 20%
Sericite 3%
Magnetite 2%
Apatite Trace
Hematite Trace
Zircon Trace

Map Location #23:
Specimen Number: 358-93
Dark-gray, quartz-plagioclase feldspar-biotite-schist, with up to 5 mm porphyroblasts of plagioclase.

Biotite 50%
Plagioclase feldspar (An39) 33%
Quartz 15%
Sphene 1%
Apatite Trace
Chlorite Trace
Magnetite Trace
Muscovite Trace
Sericite Trace
Zircon Trace
Map Location #24:
Specimen Number: 377-93
Red-green, plagioclase feldspar-quartz-microcline (granitic) cataclasite, with 10-15 mm porphyroclasts of microcline with quartz inclusions.
Microcline 40%
Quartz 30%
Plagioclase feldspar (An26) 27%
Chlorite 2%
Calcite Trace
Hematite Trace
Magnetite Trace
Muscovite Trace
Sericite Trace

Map Location #25:
Specimen Number: 412-93
Pink, medium-grained, allotriomorphic, porphyritic, biotite-granite, with color index=5.
Microcline 40%
Quartz 35%
Plagioclase feldspar (An23) 18%
Biotite 5%
Muscovite 1%
Hematite Trace
Sericite Trace
Sphene Trace

Map Location #26:
Specimen Number: 448-93
Blue-gray, muscovite-quartz-schist, with medium-grained muscovite, and up to 5 mm thick "lenses" of quartz grains.
Quartz 65%
Muscovite 30%
Biotite 4%
Magnetite 1%
Hematite Trace

Map Location #27:
Specimen Number: 471-93
Pink-green, sphene-hornblende-chlorite-quartzofeldspathic-mylonitic-gneiss, with some preserved gneissic texture, and a 10 mm thick quartz vein.
Feldspar (indeterminate varieties) 40%
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<tr>
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<td>Trace</td>
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**Map Location #28:**

**Specimen Number: 487-93**

Pinkish-brown, medium-grained, hypidiomorphic, granular, biotite-granodiorite, with color index=30.

- Plagioclase feldspar (An\textsubscript{33}) 30%
- Biotite 30%
- Quartz 25%
- Microcline 10%
- Magnetite 2%
- Apatite Trace
- Hematite Trace
- Muscovite Trace
- Sericite Trace
- Zircon Trace

**Map Location #29:** Point-counted.

**Specimen Number: 503-93**

Dark-pink, medium-grained, hypidiomorphic, porphyritic, chlorite-granite, with color index=23.

- Microcline 41%
- Quartz 24%
- Chlorite 13%
- Plagioclase feldspar (An\textsubscript{25}) 12%
- Epidote 7%
- Magnetite 3%
- Muscovite Trace
- Sphene Trace
- Zircon Trace
Map Location #30:
Specimen Number: 553-93
Pink, fine-grained, allotriomorphic, granular, porphyritic-granite, with color index=1.
Quartz 48%
Microcline 30%
Plagioclase feldspar (An ?) 20%
Hematite 1%
Chlorite Trace
Magnetite Trace
Muscovite Trace
Sericite Trace

Map Location #31:
Specimen Number: 560-93
Pink, medium-grained, hypidiomorphic, porphyritic, clinzoisite-granite, with color index=12.
Microcline 35%
Quartz 30%
Plagioclase feldspar (An_{32}) 20%
Chlorite 5%
Clinzoisite 5%
Muscovite 2%
Hematite 1%
Limonite Trace
Sericite Trace
Zircon Trace
Appendix B

Geochemical Parameters of Water Samples Taken From
the Royal Gorge Crystalline Aquifer
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<th>Station #</th>
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<th>Conductivity</th>
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<th>Temperature</th>
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<td>Well, Spring, or Stream</td>
<td>mg/l as CaCO&lt;sub&gt;3&lt;/sub&gt;</td>
<td>μS/cm</td>
<td>mg/l as CaCO&lt;sub&gt;3&lt;/sub&gt;</td>
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<td>8.39</td>
<td>21.6</td>
</tr>
<tr>
<td>46 (stream)</td>
<td>72</td>
<td>230</td>
<td>95</td>
<td>8.44</td>
<td>21.2</td>
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<tr>
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<td>95</td>
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<tr>
<td>47 (stream)</td>
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<td>220</td>
<td>90</td>
<td>8.29</td>
<td>18.3</td>
</tr>
</tbody>
</table>

* Water wells with cisterns or holding tanks.
Appendix C

Selected Major Ion Concentrations of Water Samples
Taken From the Royal Gorge Crystalline Aquifer
<table>
<thead>
<tr>
<th>Station #</th>
<th>Chloride</th>
<th>Iron</th>
<th>Nitrate as Nitrogen</th>
<th>Sulfate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Well, Spring, or Stream</td>
<td>mg/l</td>
<td>mg/l</td>
<td>mg/l</td>
<td>mg/l</td>
</tr>
<tr>
<td>1 (well)</td>
<td>96</td>
<td>&gt;5</td>
<td>0.3</td>
<td>&gt;80</td>
</tr>
<tr>
<td>2 (well)</td>
<td>10</td>
<td>0.17</td>
<td>2.0</td>
<td>&gt;80</td>
</tr>
<tr>
<td>3 (spring)</td>
<td>277</td>
<td>0.35</td>
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<td>&gt;80</td>
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<tr>
<td>4 (spring)</td>
<td>50</td>
<td>0.1</td>
<td>1.0</td>
<td>&gt;80</td>
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<tr>
<td>5 (spring)</td>
<td>42</td>
<td>0.06</td>
<td>1.4</td>
<td>&gt;80</td>
</tr>
<tr>
<td>6 (spring)</td>
<td>41</td>
<td>0.10</td>
<td>1.5</td>
<td>&gt;80</td>
</tr>
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<td>7 (spring)</td>
<td>82</td>
<td>0.32</td>
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<td>&gt;80</td>
</tr>
<tr>
<td>8 (spring)</td>
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<td>&gt;80</td>
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<tr>
<td>9 (well*)</td>
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<td>1.7</td>
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<tr>
<td>10 (well)</td>
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<td>0.25</td>
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<tr>
<td>11 (well*)</td>
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<td>12 (well)</td>
<td>114</td>
<td>0.05</td>
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<td>17 (well)</td>
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<td>&gt;80</td>
</tr>
<tr>
<td>18 (well)</td>
<td>37</td>
<td>0.06</td>
<td>5.5</td>
<td>&gt;80</td>
</tr>
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</table>

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<table>
<thead>
<tr>
<th>Station #</th>
<th>Chloride</th>
<th>Iron</th>
<th>Nitrate as Nitrogen</th>
<th>Sulfate</th>
</tr>
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<tbody>
<tr>
<td>Well, Spring, or Stream</td>
<td>mg/l</td>
<td>mg/l</td>
<td>mg/l</td>
<td>mg/l</td>
</tr>
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</tr>
<tr>
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<td>&gt;80</td>
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<td>Nitrate as Nitrogen</td>
<td>Sulfate</td>
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<td>-----------</td>
<td>----------</td>
<td>------</td>
<td>---------------------</td>
<td>---------</td>
</tr>
<tr>
<td>Well, Spring, or Stream</td>
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<td>&gt;80</td>
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<td>0.8</td>
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<td>0.5</td>
<td>34</td>
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<td>0.05</td>
<td>0.5</td>
<td>27</td>
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<td>0.18</td>
<td>0.8</td>
<td>25</td>
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<td>45 (stream)</td>
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<td>0.9</td>
<td>27</td>
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<td></td>
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<td>0.1</td>
<td>1.0</td>
<td>27</td>
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<td>46 (stream)</td>
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<td>NA</td>
<td>NA</td>
</tr>
<tr>
<td>47 (stream)</td>
<td>20</td>
<td>NA</td>
<td>NA</td>
<td>NA</td>
</tr>
</tbody>
</table>

* Water wells with cisterns or holding tanks.
Appendix D

Isotope Analysis Techniques and Precision Statistics
Sample Preparation Technique for Deuterium

For deuterium analysis, water must first be reduced to $H_2$ gas. This is accomplished by passing the water vapor over heated uranium (Bigeleisen and others, 1952) or, less frequently, zinc (Freidman, 1953; Coleman and others, 1982). Both reactions take place under vacuum and at temperatures over 400 degrees C. The metal used in this study was uranium, which follows the reaction below:

$$2H_2O + U \rightarrow 2H_2 + \text{UO}_2$$

The design of the vacuum line used in the formation of hydrogen is shown in Figure 54 (taken from Machavaram, 1993). Each water sample was first sealed within a 1-mm diameter capillary tube, and then introduced to the line at point 16. Following evacuation of the outer glass container, the capillary tube was broken using an internal metal weight and an external magnet. The water vapor was immediately transferred to a liquid nitrogen U-trap between points 14 and 16, with the assistance of a blow dryer. The U-trap was pumped to remove all non-condensable gases introduced through the capillary tube. Once an adequate vacuum was attained, by monitoring a pirani gauge (Edwards Model #502), the U-tube was sealed off at points 14 and 15 and the water allowed to thaw.

The water vapor is then put through the 750 degree C furnace by opening points 14, then 11 and 10, while keeping point 8 and 15 sealed. It should be noted that the valve at point 10 must be opened slightly at first to control the flow of water to the furnace and assure complete reduction of the water. After closing off points 3, and 5, the Toepler pump transfers the hydrogen gas to an evacuated collection reservoir at point 6. The hydrogen in the Toepler extension and in the collection

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Figure 54. Vacuum Extraction Line for the Preparation of Hydrogen Gas. The Numbers Refer to Specific Locations Mentioned in the Text.
reservoir is allowed to equilibrate for a few minutes, and then sealed and transferred to the mass spectrometer for analysis of hydrogen ratios.

The Toepler pump operates by the repetitive vertical movement of mercury governed by a controller box. During the downward stroke of mercury, the newly created hydrogen gas is let into the top chamber of the pump through a small vent. The upward stroke of mercury pushes the gas into a calibrated extension of the pump. A metal ball between the Toepler pump top chamber and the extension acts as a check-valve, preventing the backflow of gas. After several strokes of the pump, 99.9% of the hydrogen gas is pushed into the extension and collected. The exact number of Toepler strokes is determined on a sample-by-sample basis by monitoring the pressure reduction between points 7 and 8 (Figure 54).

Prior to running any field samples, the Toepler pump was calibrated with a laboratory standard to assure that the water was being completely reduced by the furnace. The process of calibration is carried out by running several weighted, water samples through the Toepler system. By knowing that each milligram of water should yield 55 μmoles of hydrogen gas, one is able to determine the theoretical yield of any known volume (or weight) of water. The yield is measured by raising the mercury to a marked level in the Toepler extension, with point 2 closed, and recording the pressure difference. A linear relationship is established between the amounts of gas and the pressure differences recorded with the Toepler. The calibration equation can then be used to determine the yield of any unknown gas sample.

Reproducibility of Hydrogen Measurements

Five samples of LAB-1 were run according to the hydrogen procedure described above and listed in Table 6. The mean and standard deviation of the δD
values were found to be -64.9, and 0.95, respectively. These values fall within the expected range of variability for \( \delta D \) analysis, and correspond to the expected values for \( \delta_{\text{LAB-1 - SMOW}} \).

Table 6

Reproducibility Measurements of Hydrogen Isotopic Ratios

<table>
<thead>
<tr>
<th>Number</th>
<th>( \delta D_{\text{SMOW}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-65.8</td>
</tr>
<tr>
<td>2</td>
<td>-64.8</td>
</tr>
<tr>
<td>3</td>
<td>-66.0</td>
</tr>
<tr>
<td>4</td>
<td>-64.0</td>
</tr>
<tr>
<td>5</td>
<td>-64.0</td>
</tr>
<tr>
<td>Mean</td>
<td>-64.9</td>
</tr>
<tr>
<td>St. Dev.</td>
<td>0.95</td>
</tr>
</tbody>
</table>

Sample Preparation Technique for 18-Oxygen

The most common method used for determining the oxygen isotopic ratio of a water sample is equilibration with carbon dioxide gas. The method was first proposed by Cohn and Urey (1938), and was then utilized by Epstein and Mayeda (1953) when they reported the first accurate, oxygen isotope values for natural waters. There are several additional methods, summarized by Machavaram (1993), that are used only
when water samples are limited in quantity. In this study, the carbon dioxide equilibration technique was used in the analysis of all samples.

The equilibration method is based on the principle that isotopic exchange between carbon dioxide and water, under equilibrium conditions, occurs with a fractionation factor that is constant for a given temperature. Equilibration of carbon dioxide takes place in the following steps:

\[
\text{CO}_2(g) = \text{CO}_2(aq) \\
\text{H}_2\text{O} + \text{CO}_2(aq) = \text{HCO}_3^- + \text{H}^+
\]

The isotopic exchange reaction is then as follows:

\[
\text{C}^{16}\text{O}_2 + \text{H}_2^{18}\text{O} = \text{C}^{16}\text{O}^{18}\text{O} + \text{H}_2^{16}\text{O}
\]

After a certain amount of time, this reaction attains equilibrium and the C\(^{16}\)O\(^{18}\)O is extracted and used for mass spectrometric analysis. The equilibration process is described in more detail by Roether (1970) and Tayler (1973).

The equilibrated carbon dioxide has an \(^{18}\)O/\(^{16}\)O ratio which is then related to the \(^{18}\)O/\(^{16}\)O ratio of water through the fractionation factor \(\alpha\), whereby:

\[
(\frac{^{18}\text{O}/^{16}\text{O}}{^{16}\text{O}/^{16}\text{O}})_{\text{CO}_2} = \alpha (\frac{^{18}\text{O}/^{16}\text{O}}{^{16}\text{O}/^{16}\text{O}})_{\text{H}_2\text{O}}
\]

The values reported in the literature for this fractionation factor range from 1.0407 to 1.0424 (Staschewski, 1964; Bottinga and Craig, 1969; Blattner, 1973; O'Neil and others, 1975).
Water samples were prepared in batches of nine and processed on Western Michigan University's mass spectrometer. A laboratory water standard was run with every batch initially, and then every second batch, after repeated confirmation of the reproducibility of the technique.

The technique involves the use of 10cc septum tubes, commonly known as "blood" vials. The septum tubes come with a rubber stopper emplaced on a glass tube that is capable of holding a vacuum. Initially, the tubes were under a slight vacuum, but a better vacuum was attained by slipping the tubes onto needles and pumping along points 4-8 of the carbon dioxide equilibration line (Figure 55). The needles were fitted into Cajon couplings and attached to the line, with the needles pointing down at points 4-8. After piercing the rubber stopper, the septum tubes were pumped to an acceptable vacuum.

Carbon dioxide was loaded into the line through valves 1 and 3, after purging the carbon dioxide inlet line of air at point 2. A liquid nitrogen trap between points 15 and 16 (Figure 55) allowed the simultaneous accumulation of enough pure carbon dioxide for loading up to two batches of samples (18 tubes). Additional pumping on the frozen carbon dioxide in the trap further purified the condensate by removing all non-condensable gases. By opening point 15, closing points 3, 9-13, 14 and 16, and removing the trap, pure carbon dioxide expanded and filled the attached septum tubes. A pressure of about 360 mmHg (about half an atmosphere) was always attained on the digital pressure readout gauge, prior to closing valve 15 and immediate replacement of the liquid nitrogen trap. The pressure was then allowed to equilibrate between the line and septum tubes for about 3-4 minutes. This step was to assure that equal amounts of carbon dioxide were taken into each septum tube via the needles. The attached septum tubes were then removed from the needles after closing valves 4-8. Once nine
Figure 55. Vacuum Extraction Line for the Preparation of CO$_2$-H$_2$O. The Numbers Refer to Specific Locations Mentioned in the Text.
septum tubes (one batch) were loaded by repeating these steps twice, the trap was removed and excess carbon dioxide pumped out through valve 16. If a second batch was to be prepared, the liquid nitrogen trap would be replaced and additional carbon dioxide would be loaded prior to pumping down the line.

The removed septum tubes, loaded with carbon dioxide, were then injected via syringes with 2cc of water. The water samples were differentiated by numbering each septum tube with a permanent marker. This stage of the process concluded with the insertion of each sample into a 25-degree C (± 0.01-degree) circulation bath, recording both the time and date.

After waiting approximately 30 hours, one septum tube at a time was pulled out of the water bath and immediately attached via needle to point 17 on the extraction line (Figure 54). The needle was not pushed through the rubber stopper at this stage, however. The next step involved pumping that portion of the line between the valve near point 17 and the needle sealed into the stopper. During this evacuation process, monitored by the pirani gauge between points 14 and 17, the lower half of the septum tube was placed in liquid nitrogen to freeze the water and carbon dioxide. Once a sufficient vacuum was attained and the water completely frozen, the valve at point 17 was shut and the needle was pushed through the stopper. A liquid nitrogen trap (-186 degrees C) was then established in the U-tube between points 14 and 16, and a dry ice/M-17 slush (-80 degrees C) replaced the liquid nitrogen around the septum tube. This allowed the carbon dioxide to be released in the septum tube, while holding the water bound as ice. The valve at point 17 was opened after sealing the valves at points 13, 14, and 15 (to prevent loss of the sample gas). All of the carbon dioxide in the septum tube was then cryogenically transferred to the U-tube trap. By slowly pumping on the trap through valve 14, the non-condensable gases were also
evacuated. After a sufficient vacuum was again reached by monitoring the pirani, the valves at 14, 15, 17, and 18 were closed and a new liquid nitrogen trap was established at the "cold finger" between the digital pressure readout and point 18. The next step involved replacing the U-tube liquid nitrogen trap with a dry ice/M-17 slush to release the carbon dioxide and retain any residual water. After opening valve 18, all the carbon dioxide gas was cryogenically transferred to the trap at the "cold finger." Closing valve 18 then sealed off the carbon dioxide gas.

The carbon dioxide in the "cold finger" was allowed to expand by removing the liquid nitrogen trap, and the pressure recorded after about five minutes (or until the pressure stabilized) by reading the digital pressure readout. This value was usually around 350 mmHg. The volume of gas in the "cold finger" was then determined by using the calibration equation for the "cold finger," which was established earlier by placing various, known amounts of gas into the "cold finger" (to the right of valve 18 in Figure 54), and regressing pressure against μmoles of gas. This equation allows the yield of any gas sample to be determined and monitored for incomplete extraction.

While the carbon dioxide was sealed in the "cold finger," the rest of the extraction line was pumped down by opening valves 14 and 15. This process was assisted by blow-drying the U-tube, which enhances the removal of residual water left in the trap. A sample tube was then hooked to the line at point 17 and opened to the pump for evacuation. The final step involved placing a dewar of liquid nitrogen around the sample tube, closing valves 14 and 15, and opening valve 18, to cryogenically move the carbon dioxide from the "cold finger" to the sample tube. The sample tube was subsequently closed and connected to the mass spectrometer for analysis.
To establish the equilibration time between carbon dioxide and water, a series of LAB-5 standards were run at a constant temperature (25 ± 0.01 degrees C). The resultant plot is shown in Figure 56, showing the determined δ¹⁸O values vs. time. A time of 30 hours (± 5 hours) was determined to be sufficient for complete equilibration of water and carbon dioxide.

**Corrections Involved in Equilibration Method**

A correction to the carbon dioxide equilibration method is necessary because the amount of oxygen introduced by carbon dioxide is not negligible compared to that of water (Craig, 1957). The overall equilibration reaction is as follows:

\[ C^{16}O_2 + H_2^{18}O = C^{16}O^{18}O + H_2^{16}O \]

Writing mass balance for both sides gives:

\[ \delta(CO_2)_{in}CO_2 + \delta(H_2O)_{in}H_2O = \delta(CO_2)_{fn}CO_2 + \delta(H_2O)_{fn}H_2O \tag{1} \]

The letters i and f denote the initial and final conditions, respectively, and n represents the mole fraction of the species of interest. Of the two unknowns in equation (1) (\(\delta(H_2O)_i\) and \(\delta(H_2O)_f\)), the value for \(\delta(H_2O)_i\) may be evaluated using the fractionation factor \(\alpha\) because:

\[ \delta(H_2O)_f = (1000 + \delta(CO_2)_f / \alpha) - 1000 \tag{2} \]

By rearranging equation (1) above the following can be written:
Figure 56. Plot of $\delta^{18}O$ vs. Time for the Septum Tube Method of Oxygen-Carbon Dioxide Gas Equilibration.
\[
\delta(H_2O)_{i}nH_2O = nCO_2\{\delta(CO_2)_f - \delta(CO_2)_i\} + \delta(H_2O)_{i}nH_2O
\]

Dividing through by \(nH_2O\) and inserting the value of \(\delta(H_2O)_{i}\) from equation (2) above yields the relationship below:

\[
\delta(H_2O)_{i} = [(1000 + \delta(CO_2)_f) / \alpha - 1000] + \delta(CO_2)_f - \delta(CO_2)_i(nCO_2/nH_2O)
\]

This equation is solved for the initial \(\delta\) value of the water before equilibration. The correction factor, represented by the ratio \(nCO_2/nH_2O\), cannot ordinarily be ignored. It is however, possible to circumvent this correction by keeping the ratio small, and by analyzing a standard often and under identical conditions. The vacuum extraction of oxygen described above avoided this correction by following both of these techniques. The mole fraction of \(CO_2\) was kept about three orders of magnitude smaller than the mole fraction of \(H_2O\). Also, laboratory standards were run with every batch, or every second batch, as described above.

**Reproducibility of Oxygen Measurements**

Four samples of LAB-5 were equilibrated according to the above procedure. All samples were equilibrated for approximately 25 hours. The mean of the \(\delta\) values was -10.56, and the standard deviation was 0.1. Both of these values are within the expected range for error and deviation, comparing LAB-5 to SMOW. Table 7 shows the actual reproducibility measurements for the oxygen procedure.
Table 7
Reproducibility Measurements of Oxygen Isotopic Ratios

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<th>Number</th>
<th>$\delta^{18}O_{SMOW}$</th>
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<td>-10.56</td>
</tr>
<tr>
<td>St. Dev.</td>
<td>0.1</td>
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</tbody>
</table>

Correction for Dissolved Salts

Carbon dioxide equilibrated with water having a high total dissolved solids content (TDS) may have a different $^{18}O/^{16}O$ ratio compared to a water with few dissolved solids (Taube, 1954). Oxygen fractionations occur between water bound in the hydration sphere of cations and free water (Gonfiantini, 1981). The system is represented by:

$$CO_2(gas) = H_2O (free) = H_2O (bound),$$

where the first pair (carbon dioxide and free water) is controlled by the equilibration fractionation factor $\alpha = 1.0412$ at 25 degrees C, and the second pair (free water and bound water) is controlled by the fractionation factor $\alpha'$, where:
\[
(1000 + \delta_S)/(1000 + \delta_{FW}) = \alpha
\]
\[
(1000 + \delta_{BW})/(1000 + \delta_{FW}) = \alpha'
\]

The subscripts FW and BW stand for "free" and "bound." \(\delta_S\) is the measured isotopic value of carbon dioxide equilibrated with salt water, and the desired value is \(\delta_G\), or the value obtained if no salt were dissolved in the water:

\[
(1000 + \delta_G)/(1000 + \delta_{W}) = \alpha
\]

If \(x\) is the fraction of bound water, Gonfiantini (1981) suggests the following isotopic mass balance equation:

\[
\delta_{W} = (1-x)\delta_{FW} + x\delta_{BW}
\]

By combining this equation with the above definition of \(\alpha\), Gonfiantini (1981) obtained:

\[
\delta_G = \delta_S[1 + x(\alpha' - 1)] + 1000x(\alpha' - 1)
\]

Because the fractionation factor \(\alpha'\) and the fraction of bound water \(x\) have different values for different cations, the equation above was modified to reflect when \(n\) different cations are dissolved in the water:

\[
\delta_G = \delta_S\left[1 + \sum_{i=1}^{n} x_i (\alpha'_i - 1)\right] + 1000 \sum_{i=1}^{n} x_i (\alpha'_i - 1)
\]
Where the fraction $x_j$ of water bound by the cation $i$ is equal to $n_j M / 55.56$, and $n_j$ is the hydration number of the $i$-th cation and $M_j$ its molality. The number 55.56 being the number of water moles per kilogram of water (Gonfiantini, 1981).

Sofer and Gat (1972) studied the isotopic effects of the carbon dioxide equilibration method using the most common cations of high concentration in natural waters. They suggested the following correction equation:

$$\left( \delta_G - \delta_{SG} \right) \left( 1000 + \delta_{SG} \right) = \left( 1.11M_{Mg} + 0.47M_{Ca} - 0.16M_K \right) \times 10^{-3}$$

Where $M$ is the molality of the cation in subscripts.

Sodium is not considered in the above equation as the fractionation factor is equal to one at 25 degrees C. Therefore, oxygen isotope values obtained via the carbon dioxide equilibration method for waters containing only sodium do not need to be corrected (Taube, 1954).

Sofer and Gat's equation suggests a correction beyond the normal analytical error (0.1 per mil) is only necessary when the molality of magnesium is greater than 0.09 (2.2 g/l), or when the molality of calcium is above 0.21 (8.4 g/l), or when the molality of potassium is above 0.62 (24.2 g/l).

Truesdell (1974) reported very similar correction factors to that of Sofer and Gat, but did not notice any fractionation for potassium at 25 degrees C. He also documented that both anions and cations play a role in the isotope fractionation between free water and bound water. Gotz and Heinzinger (1973) showed that the effects of dissolved anions can be neglected in natural waters as the fractionation factors change almost negligibly.
Studies by J. C. Fontes in 1975, and reported by Gonfiantini (1981), indicate that the carbon dioxide and water equilibration rate is considerably slowed by certain dissolved ions. The equilibration rate of a 4M solution of MgCl$_2$ is seven times slower than pure water, and that of Ca(NO$_3$)$_2$ is three times slower. Fontes found that the effects from NaCl, KCl, KNO$_3$, and HCl were much less serious and could be neglected.

Because the measured concentrations of cations and anions in the waters from the wells, springs, and streams in and around Royal Gorge arch were quite low, the Sofer and Gat correction equation was not utilized in the present study.
The isotopic analysis for this study was done with a VG Optima isotope ratio mass spectrometer manufactured by Fisons Instruments, England (formerly VG Isotech). The following section reviews the principle and mode of operation of mass spectrometry and the corrections necessary for D/H and $^{18}$O/$^{16}$O isotopic ratios.

If an ion of mass $m$ and charge $Z$ is accelerated in a potential $V$ and placed into a uniform magnetic field $B$, then the ion experiences a force and moves in a circular orbit of radius $R$, whose motion is defined by the mass-spectrometric equation, given as:

$$\frac{m}{Z} = \frac{B^2 R^2}{2V}$$

For ions of a single charge, the radius is determined by the dimension of the magnetic field and the electric field. Simultaneous adjustment of the two fields permits the selection of ions of particular masses, in effect, serving as a mass filter.

Analysis of a gas sample by mass spectrometry involves five steps: (1) ionization of gas molecules, (2) formation of the ionized molecules into a beam, (3) acceleration of the beam by an electric field, (4) deflection by a magnetic field, and (5) detection of the filtered beam. All five of these processes take place in the analyzer section of the mass spectrometer under high vacuum (Figure 57).

The analyzer section can be subdivided into three sections: (1) the source, (2) the flight tube, and (3) the collector. Gas ionization, beam formation and ion acceleration all occur in the source. Magnetic deflection takes place in the flight tube, and detection takes place in the collector.
Figure 57. Analyzer Diagram of the Mass Spectrometer.
Gases are ionized by passing a beam of electrons through the sample, which causes the gas to form both positively charged ions (losing an electron), and negatively charged ions (gaining an electron). The positively charged ions are accelerated and formed into a beam by raising the ionization chamber to a positive potential and accelerating the beam of ions out through a slit toward a second defining slit at ground potential. The two slits are called the source slit and the alpha slit, respectively.

The flight tube passes between the poles of the magnet and forms a semi-circular arc. Upon passing through the flight tube, the ionized beam is separated into subbeams of differing radii, corresponding to different masses. A particular radius, and hence mass, is selected by either the alpha slit in the source or by the resolving slit in the collector. The ions that pass through the resolving slit are detected by a Faraday cup. Current from the Faraday cup is proportional to the number of incident ions and therefore, proportional to the partial pressure of the corresponding isotopic species in the sample. Several Faraday cups are generally used simultaneously to detect different masses. One cup measures the major beam, due to the most abundant isotopic species of a molecule, and other cups measure minor beams from less abundant isotopic species.

Most commonly, a double inlet system for gas samples is used to accurately measure relative isotopic compositions of different samples with respect to a reference sample. The reference gas and sample gas are alternately introduced into the mass spectrometer via two holding reservoirs. Pressure in the reservoirs are independently controlled by metallic bellows, which open and close to achieve similar pressures for both the sample and reference gases. Thus, producing the same ion current across the major collector.
Both reservoirs are independently connected to the ion source through viscous leaks and a system of valves. The viscous leaks consist of metallic capillaries approximately 50 cm in length, with an inside diameter of 0.1 mm, connecting the reservoirs to the ion source. Near the ion source, the capillaries are constricted to allow only laminar flow. This change in gas flow, from viscous to laminar, causes some isotopic fractionation of the gas. However, the pressure gradient is kept large enough to prevent any back diffusion to the reservoir, thereby preserving the isotopic composition of the gas with time. Steady flow is rapidly established through the capillaries, and gas entering the ion source has the same composition as the gas in the reservoir (Nier, 1947; Halsted and Nier, 1950). A system of magnetic valves designed by McKinney and others (1950) is used to rapidly switch from the sample gas to the reference gas and vice versa. This design prevents alteration of the gas flow through the capillaries, even when the gases are not being admitted to the ion source.

Corrections Specific to Hydrogen Gas

This mass spectrometric correction, also called the $H_3^+$ correction, is necessary to avoid confusion between $H_3^+$ and HD$^+$ ions as both have mass 3. The $H_3^+$ ion is produced in the ion source by the following reaction:

$$H_2^+ + H_2 = H_3^+ + H,$$

where the $H_3^+$ ion is proportional to the square of the pressure of $H_2$.

The correction is carried out externally by operating the sample and reference gases at different pressures and extrapolation to zero $H_2^+$ current. An internal control
in the VG-Optima mass spectrometer permits an automatic correction by subtracting an amount proportional to the square of the major current from the minor signal.

**Corrections Specific to Carbon Dioxide-Oxygen**

Corrections involved in determining $\delta^{18}O$ and $\delta^{13}C$ from mass spectrometric analysis of CO$_2$ have been discussed in detail by Craig (1957), Dansgaard (1969), and Mook and Grootes (1973). Since the corrections were first introduced by Craig, they have henceforth been referred to as "Craig corrections."

There are two reasons for the corrections: (1) because of the contribution of $^{17}O$ to the signal of mass 45, and (2) because 0.2% of mass 46 is derived from species containing $^{13}C$ and $^{17}O$. The $\delta$-values obtained from the mass spectrometer are simply $\delta^{45}$ and $\delta^{46}$, representing the total content of all isotopic species with masses 45 and 46. Both of these values must be determined because the Craig corrections are mutually dependent on one another.
Appendix F

Water Well Data Used in the Construction of the Water Table Surface Map
<table>
<thead>
<tr>
<th>Permit #</th>
<th>Township/Range/Section</th>
<th>Total Depth</th>
<th>Water Level</th>
<th>Yield</th>
</tr>
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<tbody>
<tr>
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<td>(Water Sample #30)</td>
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<td>60501</td>
<td>NE SE 8 T18S R71W</td>
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<td>85</td>
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<td>(in feet)</td>
<td>(gal/min)</td>
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(Water Sample #29)
BIBLIOGRAPHY


Blattner, P., 1973, Oxygen from liquids for isotopic analysis, and a new determination of $\alpha_{\text{CO}_2\text{H}_2\text{O}}$ at 250 degrees C: Geochimica et Cosmochimica Acta, v. 37, p. 2691.


Friedman, I., 1953, Deuterium content of natural waters and other substances: Geochimica et Cosmochimica Acta, v. 4, p. 89.


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Black and white photographic prints (17" x 23") are available for an additional charge.

UMI
OJTT— naming anasrone nvimuie muuvienmr-

Om

Or

Orndorfer (Lower Ordovician)

Ordovician Sedimentary Rocks Undivided

Granodiorite (Precambrian)

with scattered Cambrian-age Gabbro Dikes

Quartzofeldspathic Gneiss (Precambrian)

with scattered Cambrian-age Gabbro Dikes

---

Geologic Contact

Fault, U upthrown side, D downthrown side

Anticlinal Axis

Synclinal Axis

Strike and Dip of Bedding

Strike and Dip of Overturned Bedding

Foliation

Vertical Foliation

Schistose Layers

Sampling Locations for Thin Sections

Locations of Water Sampling Points
(Numbers 48-61 Represent USGS Isotope Sampling Points; Hearne and Litke, 1987)

Scale 1:24,000

0

1 Mile

0

1 Kilometer

Topographic Contour Interval 40 Feet

Map Location

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Figure 1
Geology modified from Topography from the Mountain, and Royal City

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Figure 2

Geologic Map of Royal Gorge arch, South-Central Colorado

Timothy L. Clarey

1996

Geology modified from Mann (1957), Scott (1977), Scott and others (1978), and Taylor (1975). Topography from the Canon City, Cooper Mountain, Gribble Mountain, McIntyre Hills, Rice Mountain, and Royal Gorge USGS 7.5 Minute Series Quadrangles

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### Legend

**Map Units**

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<td>Raton Formation (Paleocene and Upper Cretaceous)</td>
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<td>Vermejo Formation and Trinidad Sandstone Undivided (Upper Cretaceous)</td>
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<td>Dakota Sandstone and Purgatoire Formation Undivided (Lower Cretaceous)</td>
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<td>Fountain Formation (Permian and Pennsylvanian)</td>
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<td>Harding Sandstone (Middle Ordovician)</td>
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<td>Manitou Limestone (Lower Ordovician)</td>
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<td>Ordovician Sedimentary Rocks Undivided</td>
</tr>
<tr>
<td>xgd</td>
<td>Granodiorite (Precambrian) *with scattered Cambrian-age Gabbro Dikes</td>
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<tr>
<td>xen</td>
<td>Quartzofeldspathic Gneiss (Precambrian)</td>
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Morrison and Ralston Creek Formation Undivided (Upper Jurassic)

Fountain Formation (Permian and Pennsylvanian)

Fremont Dolomite (Upper and Middle Ordovician)

Harding Sandstone (Middle Ordovician)

Manitou Limestone (Lower Ordovician)

Ordovician Sedimentary Rocks Undivided

Granodiorite (Precambrian)
with scattered Cambrian-age Gabbro Dikes

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(Numbers 48-61 Represent USGS Isotope Sampling Points; Hearne and Litke, 1987)

Scale 1:24,000

0 1 Mile

0 1 Kilometer

Topographic Contour Interval 40 Feet

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Figure 1: Geology model for Mountain, with Topography and Slickensides.

Domain VII - Joints

Domain VIII - Joints

Domain III (South) - Slickensides

Domain V (South) - Slickensides

Domain V (North) - Slickensides

Domain VIII - Joints

Domain VII - Joints

Movement Planes

N-Planes Intersections Kamb

Poles to Foliation

Domain VII

Domain VII (North) - Slickensides

Domain VII (South) - Slickensides

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Figure 7
Geologic Map of Royal Gorge arch, South-Central Colorado with Structural Domain Outlines
Timothy L. Clarey
1996
Geology modified from Mann (1957), Scott (1977), Scott and others (1978), and Taylor (1975). Topography from the Canon City, Cooper Mountain, Gribble Mountain, McIntyre Hills, Rice Mountain, and Royal Gorge USGS 7.5 Minute Series Quadrangles
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