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Investigation of Hydrogeologic Mapping to Delineate Protection Zones Around Springs

Report of Two Case Studies

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INVESTIGATION OF HYDROGEOLOGIC MAPPING TO DELINEATE PROTECTION ZONES AROUND SPRINGS

REPORT OF TWO CASE STUDIES

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NOTICE

The U.S. Environmental Protection Agency through its Office of Research and Development partially funded and collaborated in the research described here under Cooperative Agreement CR-817607 to the Utah Department of Environmental Quality. It has been subjected to the Agency's peer and administrative review and has been approved for publication as an EPA document. The document contains copyrighted material on pages 30 and 49. Mention of trade names or commercial products does not constitute endorsement or recommendation for use.

All research projects funded by the U.S. Environmental Protection Agency that make conclusions or recommendations based on environmentally related measurements are required to participate in the Agency Quality Assurance Program. This project was conducted under an approved Quality Assurance Project Plan and the procedures therein specified were used with the following exceptions. Geophysical logging of boreholes specified in the plan was not performed; following initial investigation at the field sites, it was determined that other techniques for acquiring the necessary information were more appropriate. In addition, the final springs chosen for study were different from those originally proposed and instrumentation for measuring water temperature in the field was changed. Information on the plan and documentation of the quality assurance activities and results are available from the Principal Investigator.

FOREWORD

The U.S. Environmental Protection Agency is charged by Congress with protecting the Nation's land, air, and water resources. Under a mandate of national environmental laws, the Agency strives to formulate and implement actions leading to a compatible balance between human activities and the ability of natural systems to support and nurture life. To meet these mandates, EPA's research program is providing data and technical support for solving environmental problems today and building a science knowledge base necessary to manage our ecological resources wisely, understand how pollutants affect our health, and prevent or reduce environmental risks in the future.

The National Risk Management Research Laboratory is the Agency's center for investigation of technological and management approaches for reducing risks from threats to human health and the environment. The focus of the Laboratory's research program is on methods for the prevention and control of pollution to air, land, water, and subsurface resources; protection of water quality in public water systems; remediation of contaminated sites and ground water; and prevention and control of indoor air pollution. The goal of this research effort is to catalyze development and implementation of innovative, cost-effective environmental technologies; develop scientific and engineering information needed by EPA to support regulatory and policy decisions; and provide technical support and information transfer to ensure effective implementation of environmental regulations and strategies.

This report presents a discussion of hydrogeologic characterization tools that can be used to evaluate potential zones of contribution to springs used in public water supplies and potential protection zones around the springs. Techniques described in this document provide a cost-effective means for obtaining basic information regarding hydrogeology and are applicable at many sites. It is published and made available by EPA's Office of Research and Development to assist the user community.

Clinton W. Hall

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ABSTRACT

Methods commonly used to delineate protection zones for water-supply wells are often not directly applicable for springs. This investigation focuses on the use of hydrogeologic mapping methods to identify physical and hydrologic features that control ground-water flow to springs to aid in delineating *springhead* protection zones. Two public-supply springs were selected as study sites to represent diverse geologic settings. One spring discharges from fractured dolomite and one from fractured siltstone, sandstone, and shale.

Hydrogeologic mapping techniques, as applied in this study, are methods for mapping geologic or hydrologic features or geophysical/geochemical signatures of subsurface features. These data are often supplemented with information from subsurface investigations to aid in extrapolating surface results to aquifer depths. In this investigation, geologic mapping, fracture-trace analysis, topographic analysis, catchment area estimation, geochemical characterization, elemental isotope studies, and a tracer study were used to locate and describe potential ground-water flow boundaries and pathways and develop conceptual models for site hydrogeology. Results were integrated to estimate the zones of contribution to each spring and evaluated for use in the delineation of potential protection zones. Data from borings and hydraulic tests supported this characterization by providing direct and indirect information regarding subsurface lithology and hydraulic parameters.

In these case studies, results of hydrogeologic mapping allowed development of the conceptual model for site hydrology and evaluation of potential ground-water flow controls. Although definitive ground-water flow boundaries suitable for delineation of practicable protection zones could not be identified at either site, these techniques provided sufficient information to support an initial evaluation of potential protection zones. These results would also serve as a strong basis for additional investigations if more detailed or reliable delineations were warranted.

This report was submitted in fulfillment of CR-817607 by the Utah Department of Environmental Quality under the partial sponsorship of the U.S. Environmental Protection Agency. This report covers a period from November 1990 to May 1993, and work was completed as of May 1993.

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Chapter 1
INTRODUCTION

BACKGROUND

Throughout the course of human history, more interest has been generated in springs than in any other ground-water feature (Davis and De Wiest, 1966). Historically, springs have determined the location of human settlements in arid regions. Many cities and towns in the western United States are named for springs. For centuries, hot springs and mineral springs were thought to be of medicinal or therapeutic value. Bottled "spring" water is still considered by many people to be of higher quality than ordinary tap water. This belief is at least partially responsible for the current high volume sales of bottled water. Unfortunately, this association of spring water with purity is not based on fact. Spring water can contain higher concentrations of dissolved solids than local public water supplies and many types of springs are more easily contaminated than other sources of drinking water.

Over 3400 public water-supply systems in the United States are known to obtain part or all of their drinking water from springs (Table 1). These systems provide drinking water for more than seven million people. However, the average number of people served by a single public-supply system utilizing springs as a water source is small. Public water supplies that utilize springs are more numerous in the western United States. More than 200 public systems use springs in each of the states of Colorado, Idaho, Oregon, Utah, and Washington.

Springs are an important component of the water supply in many states. There are several states in which more than 100,000 people are served by public systems that include springs. In the state of California, more than four million people (14.2% of 1990 population) receive their drinking water from systems with springs. In both Tennessee and Pennsylvania, more than 500,000 people receive their drinking water from such systems. In each of the states of California, Hawaii, Tennessee, and Vermont, more than 10% of the population is served by systems that include springs.

The Wellhead Protection Program was authorized by the 1986 Amendments to the Safe Drinking Water Act. This program is designed to assist states in protecting areas around wells (or springs) within their jurisdiction from contaminants that may have adverse effects on human health (Safe Drinking Water Act, section 1428[a]). One of the six critical elements of a wellhead protection program is delineation of a scientifically valid protection area around public ground-water sources.

The U.S. Environmental Protection Agency (U.S. EPA) is charged with providing technical guidance to state and local personnel on delineation of wellhead protection areas (WHPAs). As part of this effort, the U.S. EPA has published several guidance documents including "Guidelines for Delineation of Wellhead Protection Areas" (U.S. EPA, 1987), "Surface Geophysical Techniques for Aquifer and Wellhead Protection Area Delineation" (Violette, 1987), "Wellhead Protection Strategies for Confined-Aquifer Settings" (Kreitler and Senger, 1991), and "Model Assessment for Delineating Wellhead Protection Areas" (van der Heijde and Beljin, 1988). These publications focus primarily on delineation of WHPAs around wells in granular porous-media aquifers. The U.S. EPA has also published "Delineation of Wellhead Protection Areas in Fractured Rocks" (Wisconsin

Table 1. *Approximate Number of Springs Used for Public Water Supply in the United States (Various State Personnel, Personal Communication to M. Wireman, 1992)*

State	Approximate No. of Systems	Approximate Pop. Served	Total Pop. 1990 Census	% of Total Pop.
Alaska	15	4,000	550,000	0.7
Arkansas	13	42,000	2,351,000	1.8
California	106	4,238,688	29,760,000	14.2
Colorado	226	74,284	3,294,000	2.3
Connecticut	11	5,813	3,287,000	0.2
Georgia	31	239,210	6,478,000	3.7
Hawaii	67	124,000	1,108,000	11.2
Idaho	329	58,382	1,007,000	5.8
Illinois	4	15,625	11,431,000	0.1
Kansas	18	20,000	2,478,000	0.8
Maine	137	32,756	1,228,000	2.7
Maryland	48	44,333	4,781,000	0.9
Massachusetts	15	5,318	6,016,000	<0.1
Michigan	1	1,000	9,295,000	<0.1
Montana	200	15,000	799,000	1.9
Nebraska	2	644	1,578,000	<0.1
Nevada	49	21,171	1,202,000	1.8
New Hampshire	25	1,000	1,109,000	<0.1
New Jersey	10	500	7,730,000	<0.1
New Mexico	61	49,172	1,515,000	3.2
New York	100	25,000	17,990,000	0.1
North Carolina	5	250	6,629,000	<0.1
North Dakota	2	100	639,000	<0.1
Ohio	2	48,300	10,847,000	0.4
Oklahoma	10	38,870	3,146,000	1.2
Oregon	600	45,000	2,842,000	1.6
Pennsylvania	302	518,443	11,882,000	4.4
Rhode Island	3	495	1,003,000	<0.1
South Dakota	13	62,910	696,000	9.0
Tennessee	75	614,470	4,877,000	12.6
Texas	20	50,000	16,987,000	0.3
Utah	325	86,954	1,723,000	5.0
Vermont	189	63,581	563,000	11.3
Virginia	123	233,566	6,187,000	3.8
Washington	262	250,000	4,867,000	5.1
West Virginia	40	83,703	1,793,000	4.7
Wisconsin	20	500	4,892,000	<0.1
TOTALS:	**3,459**	**7,115,038**	**194,560,000**	**3.7%**

Note: States that are not listed in this table either made no use of springs in public water supplies or had no compiled statistics regarding springs at the time of data collection. Data were obtained from personnel in each state and are considered to be estimates of actual usage.

Geological and Natural History Survey, 1991), which focuses on delineation of protection areas for public wells developed in fractured-rock aquifers.

Hydrogeologic methods commonly used to delineate protection zones around water-supply wells are often not directly applicable to springs. Aquifer tests generally cannot be conducted using springs alone. However, such tests have been reported (e.g., Clarke, 1989). Tests to estimate hydraulic parameters of the aquifer supplying water to a spring usually require the use of nearby wells. Many public-supply springs are in remote areas where there are no monitoring or production wells nearby, and very little aquifer data are available. Hydrogeologic mapping techniques (e.g., geologic mapping, fracture-trace analysis, geochemical characterization, etc.) offer relatively low-cost methods for obtaining aquifer characterization data.

HYDROGEOLOGY OF SPRINGS

A spring is best defined as a concentrated discharge of ground water appearing at the ground surface as a current of flowing water (Todd, 1980). Springs occur in many forms. Bryan (1919) divided all springs into (1) those resulting from gravitational forces and (2) those resulting from non-gravitational forces. Gravity springs result from ground water flowing under hydrostatic pressure. Non-gravitational springs include those associated with volcanism and fractures that extend to great depths in the earth's crust. Most non-gravitational springs are thermal springs. The vast majority of springs are gravity springs.

Springs can be classified in a number of ways. Characteristics used to classify springs include discharge rate, type of aquifer (geology and structure), water quality, variability, and type of openings through which the water issues. The most common classification of springs, based on discharge rate, was developed by Meinzer (1923). This classification (Table 2) divides springs into eight orders of magnitude based on average discharge rate.

Table 2. *Classification of Springs Based on Average Discharge Rate (Meinzer, 1923)*

Magnitude	*Average Discharge*
First	*>170,000 l/min (> 100.0 ft^3/s)*
Second	*17,000 l/min - 170,000 l/min (10.0 ft^3/s - 100.0 ft^3/s)*
Third	*1700 l/min - 17,000 l/min (1.0 ft^3/s - 10.0 ft^3/s)*
Fourth	*380 l/min - 1700 l/min (100 gal/min - 1.0 ft^3/s)*
Fifth	*38 l/min - 380 l/min (10 gal/min - 100 gal/min)*
Sixth	*3.8 l/min - 38 l/min (1 gal/min - 10 gal/min)*
Seventh	*0.5 l/min - 3.8 l/min (1 pint/min - 1 gal/min)*
Eighth	*< 0.5 l/min (< 1 pint/min)*

The discharge rate of a spring is a function of three main variables - the hydraulic conductivity of the aquifer, the area contributing recharge to the aquifer, and the quantity of recharge (Davis and De Wiest, 1966). Based on the above classification there are only a few hundred springs of first magnitude in the world and about fifty in the United States. Most large magnitude springs discharge from volcanic rocks, limestones, or boulder/gravel aquifers.

Springs are also commonly classified based on geologic characteristics. The Water Resources Division of the United States Geological Survey recognizes eight principal types of springs based on hydrogeologic characteristics (Figure 1):

1. Artesian spring - release of pressurized water from a confined aquifer at the aquifer outcrop or through an opening in the confining unit.
2. Contact spring - occurs where a permeable water-bearing unit overlies a less permeable unit that intersects the ground surface.
3. Depression spring - forms where the ground surface intersects a water table.
4. Fracture spring - occurs where ground water flows predominantly through fractures that intercept the ground surface.
5. Geyser spring - periodic thermal spring resulting from expansive force of superheated steam within constricted subsurface channels (Todd, 1980).
6. Perched spring - occurs where infiltrating water discharges above the regional water table from a permeable geologic unit that overlies a less permeable geologic unit.
7. Seep spring - discharges from numerous small openings in permeable material. These springs typically have a very low discharge rate.
8. Tubular spring - discharges from rounded channels (karst solution openings, lava tubes).

The above classification is used by the U.S. Geological Survey for entering spring data into the National Water Data Storage and Retrieval System (Baker and Foulk, 1975). Other spring data in this national water data base include spring name, location, permanence, sphere of discharge, development or improvement data, and flow variability. This type of hydrogeologic spring classification is important as an aid to hydrogeologic mapping for purposes of delineating ground-water protection zones around springs. These classifications form the basis for a conceptual model of site hydrogeology. Each classification implies the potential existence of certain physical and hydrologic controls on ground-water flow. Knowledge of potential controls allows identification of appropriate site characterization techniques. The investigation may then be designed to test the assumptions of the conceptual model.

OBJECTIVES AND METHODOLOGY

The objectives of this cooperative research were investigation of hydrogeologic mapping techniques to determine the zone of contribution to springs and evaluation of this information for delineation of protection zones. The project was designed as a field study conducted at two sites in different hydrogeologic settings. Hydrogeologic mapping was chosen as the primary focus for this project because of its potential applicability to many hydrogeologic settings that include strongly anisotropic aquifers, such as fractured bedrock (U.S. EPA, 1987). Additional considerations included the relatively low technological requirements and implementation costs for many of these methods. Based on these studies, a general methodology for applying such techniques is proposed. The report is designed to aid investigators involved in planning characterization studies leading to the establishment of protection zones around springs. The project was limited in scope. Field studies were conducted at only two sites in the arid southwest region of the United States. Although general conclusions regard-

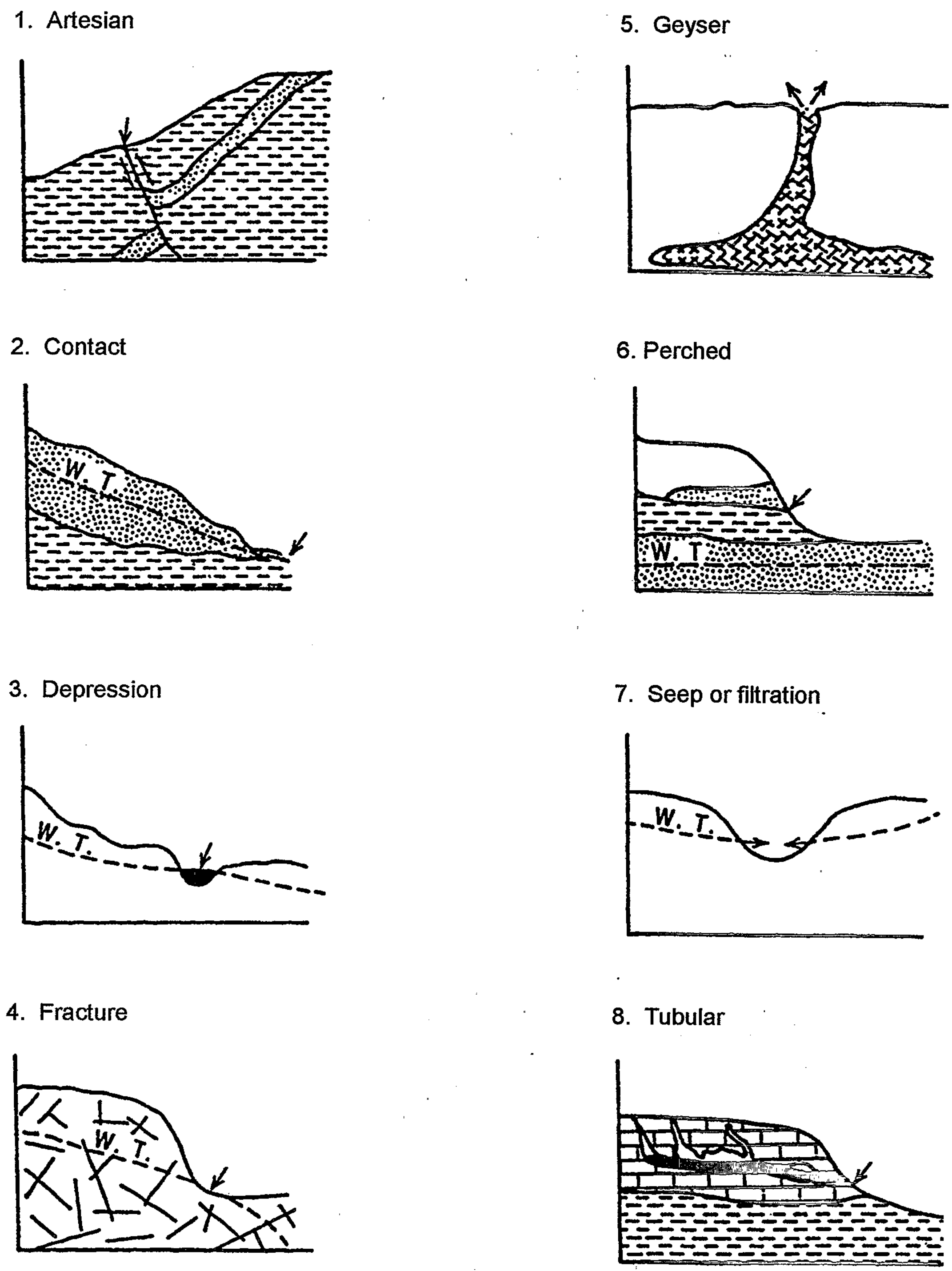


Figure 1. *Principal types of springs based on hydrogeologic characteristics (Baker and Foulk, 1975). Numbers are keyed to text descriptions. Arrows point to spring locations.*

ing utility of basic hydrogeologic characterization methods are applicable to delineation of protection zones in most settings, the utility of specific methods will be highly dependent on site characteristics.

Hydrogeologic mapping, as applied in this project, refers predominantly to geologic, hydrologic, geochemical, and geophysical techniques for characterizing subsurface features using the surface expression and geophysical or geochemical signatures of such features. These techniques include:

- fracture-trace analysis,
- analysis of land surface topography,
- geologic mapping,
- potentiometric surface mapping,
- geophysical surveys,
- tracer studies,
- stable and radioactive isotope studies, and
- geochemical characterization.

Depending on site conditions, ground-water flow boundaries and pathways may be identified and mapped in detail. A discussion of these methods and potential applicability is provided in Chapter 2.

The methodology applied in this study began with a review of site-specific literature to develop an initial conceptual model for site hydrogeology. Based on this model, potential ground-water flow controls were identified. Hydrogeologic mapping methods were chosen to test assumptions of the conceptual model and, potentially, delineate physical and hydrologic flow controls. Results were interpreted to refine the hydrogeologic conceptual model and estimate the zone of contribution to the springs. These results were then evaluated for use in delineating potential protection zones. Monitoring wells were installed at each site to provide subsurface information and allow preliminary hydraulic testing. Results of subsurface investigations were compared with indirect information inferred from the hydrogeologic mapping.

WELLHEAD PROTECTION CRITERIA AND THRESHOLDS

The delineation criteria and thresholds specified in the Drinking Water Source Protection (DWSP) Program for the State of Utah were used in this investigation as an example of state requirements in drinking water source protection programs. Approximately 785 springs are used in Utah as public water-supply sources. All sources of ground water for public supplies, including springs and tunnel discharges, are subject to the provisions of the wellhead protection program in Utah.

The Utah DWSP Program specifies three protection zones, with higher levels of management in the zones closest to the ground-water source. Zone 1 (closest to each source) is based on arbitrary distance. Zones 2 and 3 are based on ground-water time-of-travel criteria or flow boundaries. The three drinking-water source protection zones defined in Utah are:

Zone 1 - the area within a 100-foot (30 m) fixed-radius zone around the well or spring;

Zone 2 - the capture zone area within the 250-day ground-water time-of-travel to the well or spring, the boundary of the aquifer(s) that supplies water to the ground-water source, or the ground-water divide, whichever is closer; and

Zone 3 - the capture zone area within the 15-year ground-water time-of-travel to the well or spring, the boundary of the aquifer(s) that supplies water to the ground-water source, or the ground-water divide, whichever is closer.

The applicability of the information inferred from mapping to practicable protection zone delineation was evaluated at each study site using these management criteria.

Chapter 2
HYDROGEOLOGIC MAPPING METHODS FOR PROTECTION ZONE DELINEATION

The following is a general discussion of hydrogeologic mapping techniques and is not limited to those methods chosen for use at the study sites. These methods may be used to delineate ground-water flow boundaries, identify geologic features that affect ground-water velocity in many settings, and assess vulnerability of an area to ground-water contamination from surface sources. Potential physical and hydrologic controls on ground-water flow to springs include:

1. geologic units that act as flow boundaries due to low hydraulic conductivity,
2. physical limits on geologic formations that serve as aquifers,
3. physical properties of the aquifer materials (e.g., hydraulic conductivity),
4. fractures or karst features that act as preferential flow paths,
5. faults that may act as flow paths or flow boundaries,
6. ground-water divides, and
7. recharge and discharge locations/characteristics, including well pumping and irrigation.

Techniques that are useful in characterizing the hydrogeologic setting and the zone of contribution to springs are described below.

SURFACE MAPPING OF HYDROGEOLOGIC FEATURES

Geologic Mapping

Geologic mapping can be used to identify surface and subsurface lithologic units and their stratigraphic relationship, map geologic structure (faults, folds, and fractures), and map the spatial orientation of these features. Geologic mapping techniques are explained in more detail in Compton (1962) and Barnes (1981). In hydrogeologic studies, geologic mapping may often be used to:

1) infer the extent of lithologic units that provide water to springs,
2) identify low-permeability lithologic units that may form ground-water flow boundaries,
3) identify faults that may act as either barriers or preferential paths for ground-water flow,
4) characterize fracture orientations or karst features that may control ground-water flow, and
5) identify potential recharge areas.

Geologic mapping may yield sufficient information to indicate the zone of contribution to a spring through identification and delineation of formations that behave as aquifers and aquitards. The potential utility of such mapping depends, in part, on the degree to which the lithologic units of interest are exposed and the hydrogeologic setting.

Fracture-Trace Analysis

Fracture traces and lineaments are natural linear features expressed on the surface of the earth. These features are commonly caused by steeply dipping faults or joint systems (Billings, 1972). The terms fracture trace and lineament have been differentiated on the basis of length (Fetter, 1988; Lattman, 1958). Fracture traces are between 300 m and approximately 1300 m (1000 ft to 4300 ft) in length and lineaments are longer than 1300 m (4300 ft). Fracture traces may be identified on aerial photographs or topographic maps by straight stream segments, aligned erosional features, low spots in ridges, vegetation patterns, or other visually linear features (Lattman, 1958). Major fracture systems or faults, expressed as fracture traces and lineaments, may control ground-water movement, acting as preferential flow paths or flow boundaries. Fracture-trace analysis may be used to locate features for more detailed study during geologic mapping and other investigations. Such analysis generally is conducted as a preliminary step in characterization of fractured or faulted terrains.

Topographic and Geomorphic Analysis

Analyses of topographic maps may be used to locate the boundaries of surface drainage basins surrounding springs. In shallow, water-table aquifers, ground-water divides often coincide with drainage-basin divides (Fetter, 1988). Therefore, delineation of topographic divides surrounding springs is useful in developing the conceptual model for potential zones of contribution. Geomorphic analysis uses techniques of both topographic analysis and geologic mapping. Geomorphic principles (Bloom, 1978) may be applied to infer subsurface structure from present landforms. For example, the potential presence of subsurface karst features may be inferred from landforms such as depressions indicative of sinkholes.

CATCHMENT AREA ESTIMATION

In most hydrogeologic settings, the rate of discharge of a spring is related to the area contributing recharge to the aquifer and recharge rate. Using estimates of recharge rate and the discharge rate from the spring, the area of the topographic catchment basin required to provide the spring discharge can be estimated. This information may aid in determining if the zone of contribution to a spring is potentially larger than the topographic basin surrounding the spring. It should be recognized that detailed recharge information usually is difficult to acquire and the recharge estimate will generally incorporate a significant degree of uncertainty. Thus, estimates of required catchment areas may also include significant uncertainty. However, such techniques have been successfully used in many settings, including hydrologically ungauged catchments (Bonacci and Magdalenic, 1993).

TRACER STUDIES

Tracer studies are often used to identify specific ground-water flow paths and estimate ground-water velocity to discharge. Such investigations usually are performed by introducing a tracer into an aquifer through a well or sinkhole. Many inorganic and organic compounds have been used as tracers in previous studies. Applicable compounds depend on many site-specific conditions such as ambient ground-water quality, study objectives, and local regulations. Selected wells and/or springs are monitored to determine the discharge point(s), time of travel, and recoverable quantity of the tracer. In karst settings and other areas where ground-water flow occurs predominantly through a few discrete pathways, tracer studies may be the only means for reliable estimation of the zone of contribution to springs. Additional discussion regarding tracer studies is provided in numerous texts and journals (e.g., Davis and others, 1985; and Mull and others, 1988).

GEOCHEMICAL CHARACTERIZATION

The chemical composition of ground water is influenced by many factors as the water is recharged and flows through an aquifer. These factors include water-rock interaction along the flow path, composition of the recharging precipitation, water-soil-air interaction in the unsaturated zone, and biologic activity (Hem, 1985).

Analyses that may be useful indicators of aquifer processes include concentrations of major ions. Classification using such systems as the hydrochemical facies of Back (1966) and the calculation of pertinent mineral saturation indices (e.g., calcite, dolomite, and aragonite) may be valuable tools for inferring aquifer lithology and relative ground-water residence times. Temporal fluctuations in these parameters may indicate variations in ground-water flow paths or flow rates and residence time (Shuster and White, 1971; Quinlan and others, 1991). Temporal changes in physical/chemical parameters (e.g., temperature, pH, specific conductance, turbidity, and discharge rate) measured in the field may also indicate changes in ground-water flow path or residence time and may aid in defining the zone of contribution by indicating connections between ground water and surface water (e.g., Quinlan and others, 1991). Daily or weekly fluctuations in water temperature or other parameters may indicate a shallow aquifer, rapid recharge, recharge in proximity to the spring, or conduit flow conditions. Fluctuations in pH, specific conductance, turbidity, or flow may also be indicative of changes in recharge volume, area, or rates.

ISOTOPE STUDIES

Elemental isotopes may be used to estimate ground-water residence time, characterize recharge distribution, and trace ground-water flow from recharge to discharge areas. Several radioactive and stable isotopes have been used in such investigations (e.g., Hendry, 1988; Mazor and Nativ, 1992; Solomon and Sudicky, 1991; Taylor and others, 1992). Two common radioactive isotopes used to date ground water are tritium (^{3}H) and carbon-14 (^{14}C).

Tritium occurs naturally in the atmosphere, has a half-life of approximately 12.4 years, and is carried into ground water by recharging precipitation. Tritium was introduced into the atmosphere, and, thus, into ground water, at increased levels by atmospheric testing of nuclear weapons between 1952 and 1969 (Drever, 1988). As the timing of this significant increase in atmospheric tritium is known, tritium can be used in a qualitative way to estimate the average residence time of ground water.

Before atmospheric testing of nuclear weapons began in 1952, the natural tritium activity in rainwater was approximately 10 tritium units (TU). One TU equals one tritium atom per 10^{18} hydrogen atoms. Ground water containing less than about 2 TU to 4 TU entered the ground-water system prior to 1953. If the tritium content is significantly greater than 10 TU to 20 TU, the ground water has been recharged or in contact with the atmosphere since 1953 (e.g., Drever, 1988; Fetter, 1988). A more detailed analysis of average ground-water age based on tritium concentrations is also presented by Hendry (1988). However, site-specific data on tritium activity in precipitation since nuclear testing began is necessary for a more quantitative age assessment. Recent studies (e.g., Solomon and others, 1992) have used analyses of tritium and tritiogenic helium-3 (^{3}He) to overcome this limitation in certain situations and accurately date ground water with ages between 0 and 50 years. Ground water of similar age may also be dated using concentrations of anthropogenic chlorofluorocarbon compounds (Busenberg and Plummer, 1992; Szabo and others, 1996).

Carbon-14 also occurs naturally in the atmosphere and was introduced in elevated concentrations by atmospheric nuclear testing. Carbon-14 has a much longer half-life (about 5730 years) than tritium, so it can be used to date waters ranging in age from several thousand years to a few tens of thousands of years (Drever, 1988; Hendry, 1988). Difficulties in determining the initial carbon-14 concentration and separating the influence of carbon contributed from aquifer material may limit reliability of this technique. Other isotopes that have been used to assess the residence time of ground water and trace flow paths from recharge to discharge include the isotopic ratios of oxygen ($^{18}O/^{16}O$), hydrogen ($^{2}H/^{1}H$), chlorine-36 (^{36}Cl), and helium-4 (^{4}He) (Hendry, 1988; Marty and others, 1993; Mazor and Nativ, 1992).

Information regarding relative ground-water residence time and flow paths derived from isotopic studies may be useful in evaluating potential protection area delineations (Wisconsin Geological and Natural History Survey, 1991). This information may also be used to support or refute relative ages determined from hydraulic calculations (Mazor and Nativ, 1992). However, such determinations yield the average age of dissolved species. Mixing of ground waters of different ages may result in an isotopic age that differs significantly from the average ground-water travel time (Solomon and others, 1992). The applicability of these methods depends on the hydrogeologic setting and the results may be qualitative or ambiguous (Wisconsin Geological and Natural History Survey, 1991).

POTENTIOMETRIC SURFACE MAPPING

Potentiometric surface mapping is one of the most powerful tools in aquifer studies. Information regarding the hydraulic gradient, potential ground-water flow directions, and ground-water flow boundaries may be obtained using this tool. Temporal fluctuations in these parameters may also be assessed. However, care must be exercised in construction of potentiometric maps to ensure data representative of the aquifer in question are used. Equal care must be exercised in interpretation of this information, particularly in fractured media (e.g., Wisconsin Geological and Natural History Survey, 1991). This technique could not be applied in this investigation due to a lack of monitoring points (e.g., wells, springs, or seeps).

GEOPHYSICAL TECHNIQUES

Many surface geophysical methods have been applied to identify signatures of subsurface features that may control ground-water flow (e.g., faults, changes in lithology). Potential uses include delineation of subsurface lithologic and stratigraphic boundaries, depth to ground water or bedrock, and fracture/solution channel distribution/orientations. These techniques often complement results of geologic mapping by providing controls on the positions of aquifer and aquitard formations at depth. Applications of such methods are discussed in numerous texts (e.g., Telford and others, 1981). Surface geophysical techniques and their application to protection zone delineation are explained in greater detail in Violette (1987). Techniques include seismic methods (refraction and reflection surveys), electrical methods, ground penetrating radar, and potential field surveys (e.g., gravity and magnetic methods).

Borehole geophysical techniques may be useful if appropriate boreholes or wells are available. Information regarding subsurface lithology, stratigraphy, fracture density and orientation, and other important parameters are obtainable using such tools (Keys, 1989). Information from these studies may be used to control subsurface projections interpreted from mapping of surface features.

METHODS USED TO SUPPORT HYDROGEOLOGIC MAPPING

Hydraulic tests (e.g., pumping and slug tests) were performed and ground-water time-of-travel was estimated to support the hydrogeologic mapping conducted in this investigation. Hydraulic testing provided an independent estimate of aquifer parameters for evaluating zones of contribution, potential protection zones, and parameters qualitatively characterized using hydrogeologic mapping. The objective of these tests was to provide preliminary estimates of hydraulic parameters. The tests were limited in scope and were not designed to define heterogeneity or anisotropy within the potential protection zones at these sites.

As with many investigations in fractured media (U.S. EPA, 1991), the hydraulic tests were analyzed using standard solutions and methods based on porous media assumptions. An underlying assumption governing the use of standard ground-water flow equations to describe fractured bedrock settings is that the aquifer behaves as a porous medium at the problem scale (Wisconsin Geological and Natural History Survey, 1991). A second implicit assumption is that aquifer properties can be characterized using methods developed with the porous-media assumption. Although the spring sites studied in this project may better satisfy the porous-media assumptions at the scale of the zone of contribution, it is unlikely that either site behaved as a porous medium at the scale of the aquifer tests. Thus, results from these tests are considered preliminary estimates and subject to a relatively high degree of uncertainty. However, these results appear to be sufficient for the qualitative comparisons made in this study.

Chapter 3
RESULTS OF HYDROGEOLOGIC MAPPING

Two public-supply springs in Utah were selected as study sites in different hydrogeologic settings: a fractured carbonate aquifer (Olsens Spring) and a fractured clastic-rock aquifer (Sheep Spring). Both settings are common in this region. In each case, site-specific literature, including investigations by state and federal agencies and universities, was first reviewed to develop an initial conceptual model for site hydrogeology. Based on the conceptual model, potential ground-water flow controls (e.g., dominant fracture systems and low permeability boundaries) were identified and hydrogeologic mapping methods were chosen for investigation.

OLSENS SPRING

Location and Description

Olsens Spring is located in the Wasatch Range of the Middle Rocky Mountains physiographic province (Stokes, 1986) of northern Utah (Figure 2 and Table 3). The spring is about 5 km (3 mi) east of the Wasatch Fault, the eastern boundary of the Basin and Range province. It is located along the northwest margin of Mantua Valley, at an elevation of approximately 1579 m (5180 ft). It is one of several springs located around the margin of this small valley. West Hallings Spring is located approximately 275 m (900 ft) southwest of Olsens Spring within the same surface drainage basin. This study focused on Olsens Spring. However, West Hallings Spring and other springs in Mantua Valley were considered during these investigations, as appropriate.

Brigham City, which has a population of approximately 17,000, uses Olsens Spring and other springs as sources of potable water. Olsens Spring was developed as a water supply approximately 70 years ago, and no

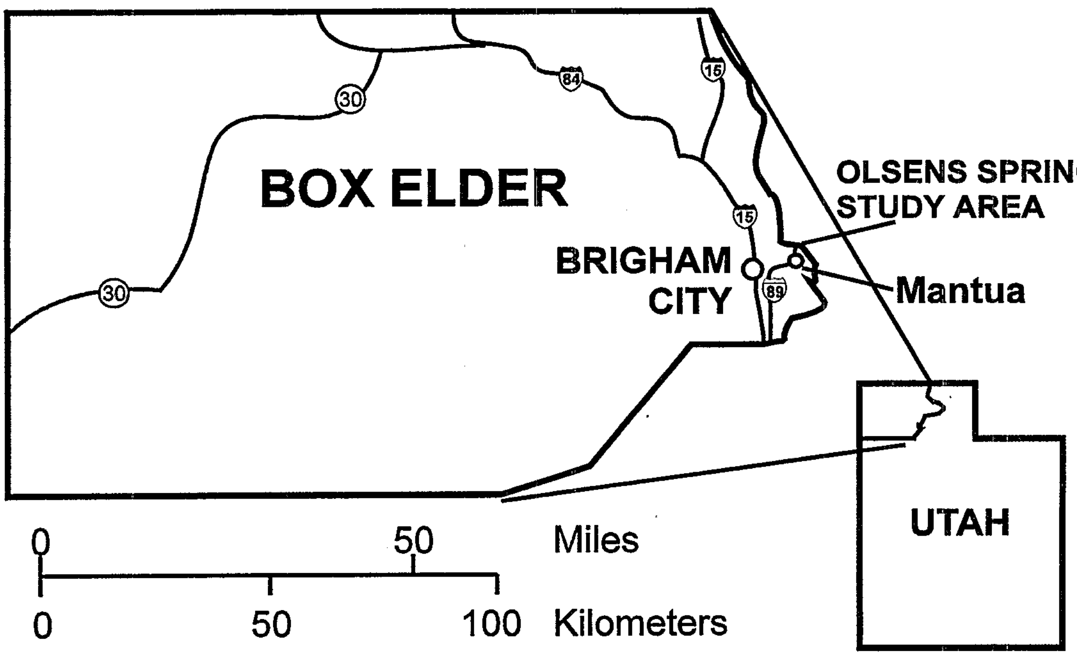


Figure 2. Location of Olsens Spring study area.

Table 3. *Location and Elevation of Springs and Associated Monitoring Wells Interpolated from 1:24,000 - Scale Topographic Maps*

	Latitude	Longitude	Elevation
Olsens Spring	41°30'52" N	111°56'30" W	1579 m (5180 ft)
West Hallings Spring	41°30'44" N	111°56'36" W	1579 m (5180 ft)
Mud Spring	41°31'08" N	111°56'05" W	1580 m (5185 ft)
Well 1 (Top of casing)	41°30'54" N	111°56'34" W	1609 m (5280 ft)
Well 2 (Top of casing)	41°30'55" N	111°56'32" W	1608 m (5275 ft)

plans are available to show how it was developed. The spring appears to have been developed by laying collection pipes in gravel. The pipes feed concrete spring boxes connected to the city water-supply system.

The collection area is in alluvial fan deposits, within several feet of the contact between the dolomite bedrock and the fan deposits. The recharge area is mostly hilly and mountainous topography, with several surface drainages that potentially contribute recharge. Estimated yield of this fourth magnitude spring (Table 2) is about 1700 l/min (1.0 ft^3/s). West Hallings Spring is a third magnitude spring that discharges about 6400 l/min (3.8 ft^3/s) (Brigham City Corporation, personal communication, 1987) from faulted limestone stratigraphically lower, but, in the same hill as Olsens Spring. Mud Spring is 790 m (2600 ft) northeast of Olsens Spring, and is a third magnitude spring discharging a maximum of about 5940 l/min (3.5 ft^3/s) (Bjorklund and McGreevy, 1973). A canyon separates Mud Spring from Olsens Spring.

Climate

The nearest weather observation station (Table 4) is in Brigham City, about 6.5 km (4 mi) west of Olsens Spring. Brigham City is 290 m (950 ft) lower in elevation than Olsens Spring and is located in a very large valley west of the Wellsville Mountains. The weather station at Pineview Dam is in a geographic setting similar to Olsens Spring and is only 73 m (240 ft) lower in elevation than the spring. Pineview Dam is located about 30 km (19 mi) southeast of Olsens Spring.

Table 4. *Climatic Data from Weather Stations Near Olsens Spring (Ashcroft and others, 1992)*

	Brigham City Waste Plant	Pineview Dam
Elevation	1290 m (4230 ft)	1506 m (4940 ft)
Average mean temperature	9.7 °C (49.5 °F)	6.5 °C (43.7 °F)
Average precipitation	48 cm (18.8 in)	79 cm (30.9 in)
Average snowfall	75 cm (29.7 in)	301 cm (118.4 in)
Maximum evapotranspiration	118 cm (46.6 in)	110 cm (43.2 in)

Average annual precipitation at Brigham City and Pineview Dam is 48 cm (18.8 in) and 79 cm (30.9 in), respectively. Average annual evapotranspiration of rangeland and mountains at the altitude of Olsens Spring is estimated to be about 35 cm (13.6 in) (Bjorklund and McGreevy, 1974).

Previous Hydrogeologic Investigations

A thesis describing the hydrogeology and hydrochemistry of several springs in Mantua Valley was completed by Rice (1987). Olsens Spring was not directly included in that study, but the "West Hallings Spring" samples collected during that investigation were actually composite samples of West Hallings Spring and Olsens Spring discharges (K. Rice, personal communication to M. Jensen, 1992). Rice (1987) mapped the general geology of the entire Mantua Valley area, but did not differentiate the individual stratigraphic units. Therefore, it was necessary to re-map the Olsens Spring area as part of this investigation. The Mantua Valley area is also included in a report on ground-water resources of the lower Bear River by Bjorklund and McGreevy (1974).

The Olsens Spring study area is included on the Box Elder County geologic map produced by Doelling (1981). The stratigraphy and structure of Mantua Valley are discussed in a regional geologic study by Williams (1948). The Mantua Valley area is also included on a surficial geologic map by Personius (1990) and in a soil survey of eastern Box Elder County conducted by Chadwick and others (1975).

Geologic Setting

Olsens Spring is situated at the base of the Wellsville Mountains. The area around the spring is underlain by a Paleozoic stratigraphic section of interbedded limestone, dolomite, shale, and quartzite (Hintze, 1988). These rocks were faulted and fractured by east-directed thrust faulting during the Cretaceous Sevier orogeny. Eastward tilting and further faulting and fracturing are a result of normal faulting during the last 10 million years along the Wasatch and related faults (Hintze, 1988).

Surface Mapping of Hydrogeologic Features

Prior to geologic mapping, topographic maps (1:24,000 scale) were studied to identify the topographic divides of the surface drainage basin surrounding Olsens Spring and West Hallings Spring. Ground-water divides, which represent hydrologic boundaries to shallow ground-water flow, may be coincident with the topographic divides. For this reason, the surface drainage basin was defined as the area for detailed study. No perennial streams or other surface water bodies that may infer the presence of hydrologic boundaries are located within the basin.

Fracture traces and lineaments were identified on 1:20,000-scale aerial photographs in conjunction with the geologic mapping. The most extensive feature mapped on the aerial photographs was a lineament found to coincide with a strike-slip fault located about 3 km (2 mi) northwest of Olsens Spring. No fracture traces were mapped through Olsens Spring or West Hallings Spring. Geologic field mapping was conducted in the surface drainage area above Olsens Spring. Lithologic contacts, bedding attitude, and structural features (e.g., joint location and attitude) mapped in the field were plotted on 1:20,000-scale black-and-white stereo-pair aerial photographs. These data were then transferred to a 1:24,000-scale topographic base map.

Results of this mapping (Figure 3) indicate that Olsens Spring discharges from dolomite of the Cambrian-age Nounan Formation. The Nounan consists of medium to thick-bedded dolomite and sandy dolomite. The

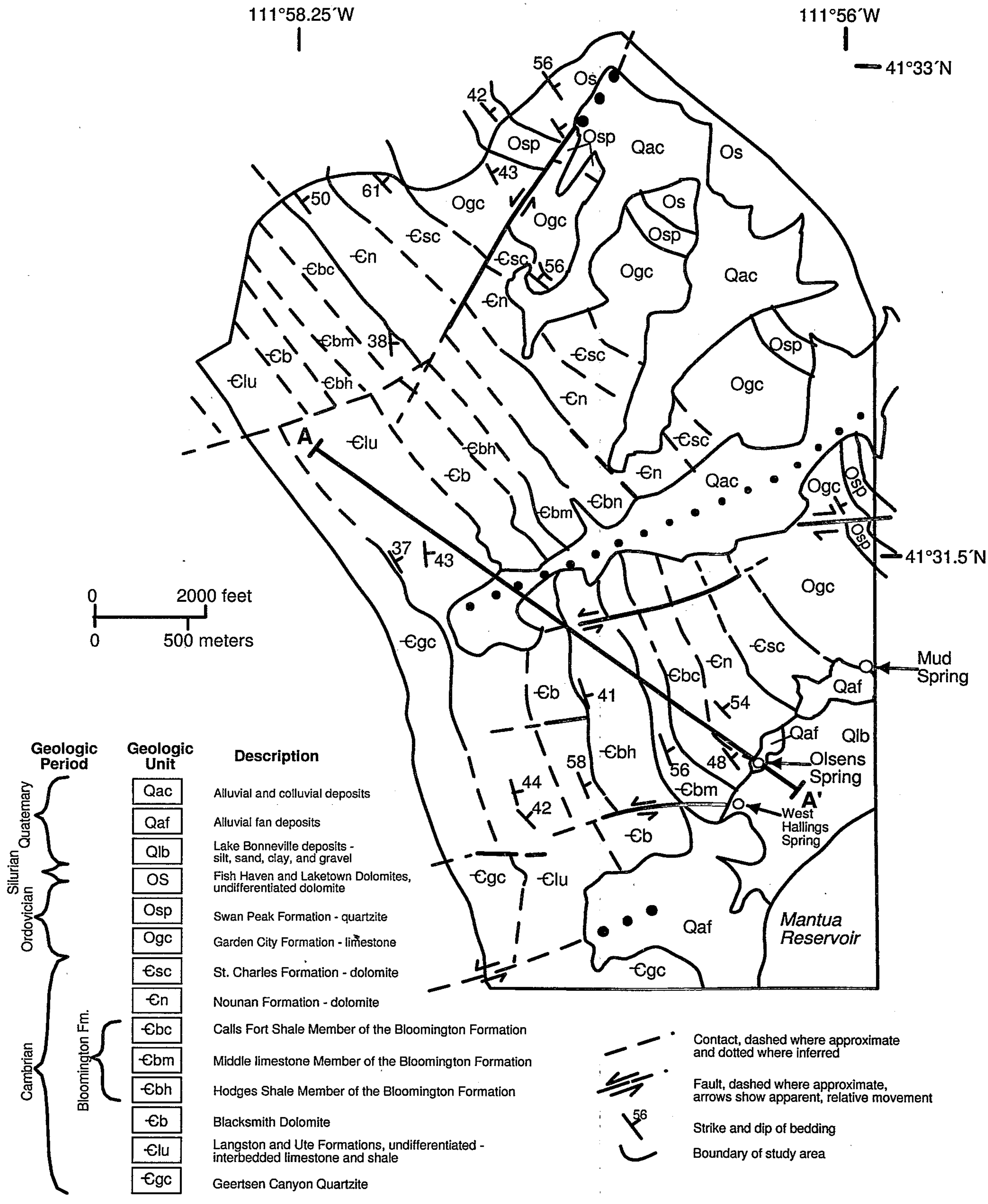


Figure 3. *Geologic map of the Olsens Spring study area. Field mapping was performed by M. Jensen and M. Lowe, with modifications after Williams (1948) and Gelnett (1958). The bedding orientations presented on this figure are provided for illustration. Some measurements were not posted for clarity.*

Nounan Formation is about 250 m (825 ft) thick in this area. The spring collection area is located approximately 9 m (30 ft) stratigraphically above the contact between the Nounan and the underlying Calls Fort Shale Member of the Bloomington Formation. The Calls Fort Shale Member consists of well-indurated shale with interbedded pods of limestone. The contact between the Calls Fort Shale and the Nounan Formation dolomite is a gradational or interbedded contact. Stratigraphically below the Calls Fort Shale is an unnamed middle limestone member from which West Hallings Spring discharges and a basal shale member similar to the Calls Fort.

Mud Spring is located approximately 800 m (2600 ft) northeast of Olsens Spring, at approximately the same elevation as Olsens and West Hallings Springs. Mud Spring discharges at the base of Round Hill from limestone of the Ordovician-age Garden City Formation at the contact with Quaternary-age valley-filling sediments. The Garden City Formation consists of medium dark gray, finely crystalline limestone, which weathers medium light gray and forms slopes and ledges. The Garden City Formation in this area contains intraformational conglomerate, which is an identifying characteristic. This unit is approximately 460 m (1500 ft) thick beneath Round Hill and contains pods and layers of black chert in the upper portion. The Garden City Formation is separated from the Nounan Formation at Olsens Spring by the Cambrian-age St. Charles Formation. No fracture traces or faults were mapped through Mud Spring.

The stratigraphic units near the springs strike about N30°W and dip 45° to 60° to the northeast. Section A-A' (Figure 4) was chosen to show the eastward-dipping Paleozoic units and minor faults in the potential zone of contribution. The location of this cross section is shown on Figure 3. Dominant joints near Olsens Spring strike N28°E to N86°E (Figure 5). Joints throughout the surface drainage area strike predominantly north to northeast (Figure 6). Joint spacing mapped in outcrops at most locations in this area is generally 0.3 m to 1.2 m (1 ft to 4 ft). However, joint density near Olsens Spring is higher, with a spacing of about 0.15 m to 0.20 m (0.5 ft to 0.7 ft) in outcrops. Joints were generally traceable across an individual outcrop. However, outcrops generally were limited to 1 m to 3 m (4 ft to 10 ft) wide exposures and surface exposures were often weathered. Thus, joint lengths could not be reliably mapped. Joint apertures appeared to range from about 0 to 2 cm (0 to 0.75 in) in diameter. Thin calcite fillings were found in many joints with localized deposits of coarsely crystalline calcite.

Several small faults were mapped in the drainage basin (Figure 3). Many of these faults appeared to have a transverse component of motion. Some of this apparent movement may be the result of dip-slip motion across dipping beds. Mapped fault lengths within 1000 m (3300 ft) of the springs ranged from approximately 600 m to 1200 m (2000 ft to 4000 ft). The closest fault appears to intersect West Hallings Spring. Faulting was not identified in the immediate vicinity of Olsens Spring. In addition, no surface expressions of karst features were observed during the field mapping.

Based on the geologic mapping, it appears that the area surrounding Olsens Spring and West Hallings Spring is extensively fractured. From these results, several hypotheses regarding the hydrogeologic system may be formulated:

1. Extensive fracturing in the exposed bedrock may indicate a relatively high degree of vulnerability of the ground water to surface contamination sources.
2. The hydrologic system may behave as a porous media on the scale of the surface drainage basin. However, sufficient data are not available to test this hypothesis.
3. Hydraulic conductivity and, potentially, ground-water velocity may be relatively high in this system.

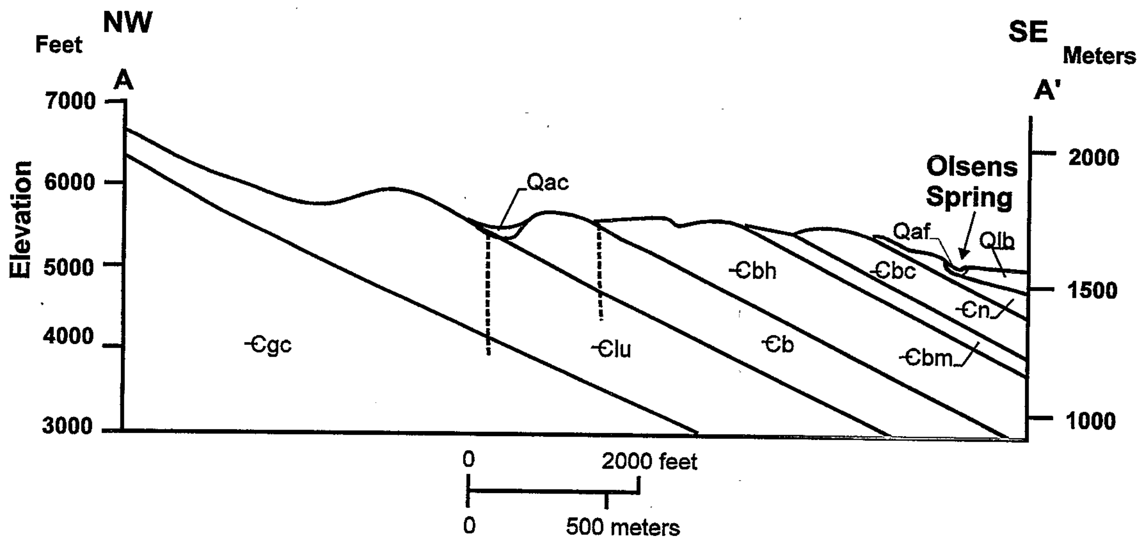


Figure 4. *Geologic cross section through the Olsens Spring area. Projections of mapped faults are dashed. Elevations are referenced to mean sea level.*

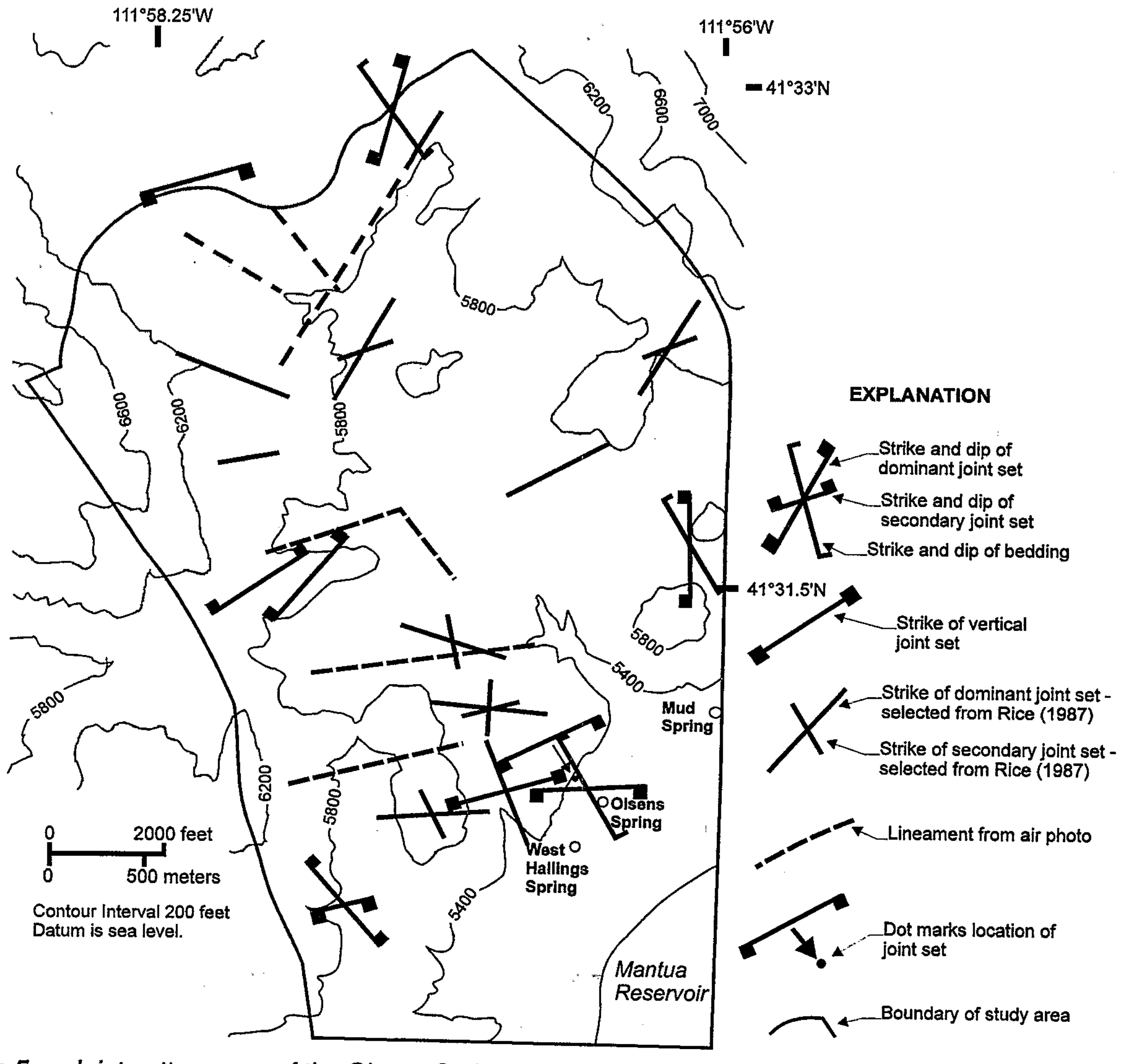


Figure 5. *Joint pattern map of the Olsens Spring study area. Field mapping was performed by M. Jensen and M. Lowe, with selected joint measurements from Rice (1987).*

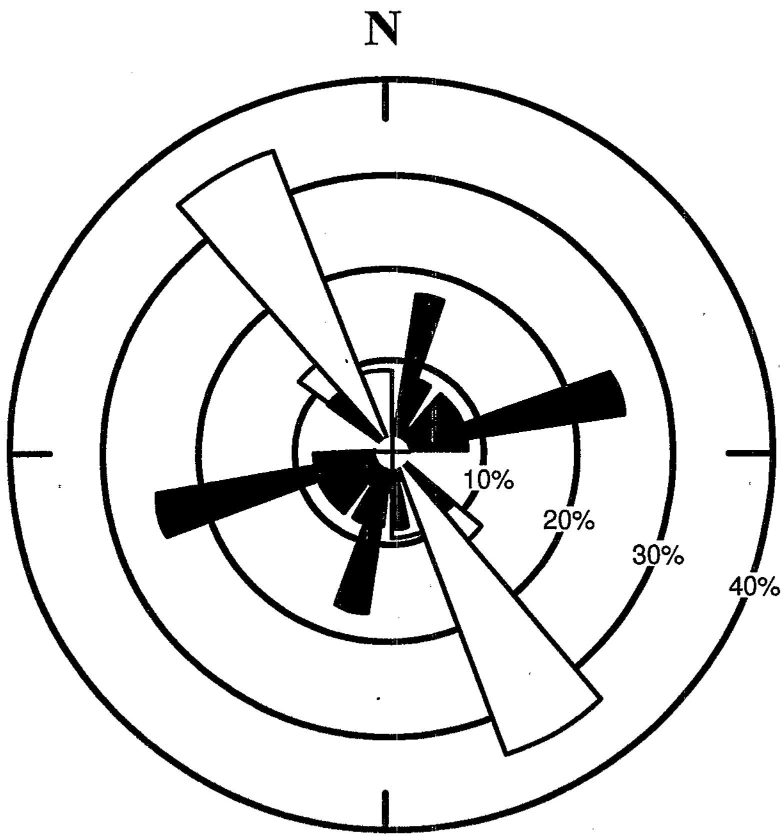


Figure 6. *Rose diagram depicting strike of joints in the surface drainage area surrounding Olsens Spring. Unshaded sectors represent strike of bedding. Number of joint measurements = 12, number of bedding plane measurements = 34.*

4. The system may be anisotropic with an increased potential for ground-water movement to the springs from the north to east-northeast based on joint orientations.
5. The Calls Fort Shale Member separating Olsens Spring and West Hallings Spring does not appear to be a boundary to ground-water flow based on the jointing mapped in outcrops of this unit.
6. Ground-water flow to West Hallings Spring may be controlled to some degree by the fault on which it is located. However, no evidence of significant fault control of ground-water flow to Olsens Spring could be inferred from geologic mapping. It is also conceivable that these springs are controlled by a single structural feature not identified in this study.

Borings and Well Installations

Two monitoring wells were installed topographically upgradient from Olsens Spring (Figure 7) to provide additional subsurface lithologic information and for use in estimating hydraulic parameters. Potential well locations in this area were limited by the relatively steep topography. Of the locations accessible to the drilling rig, these sites were chosen on the premise that hydraulic gradients may mimic topographic gradients and that the formations may be sufficiently fractured to provide flow paths to Olsens Spring or West Hallings Spring. Well 1 was located approximately 114 m (373 ft) topographically upgradient from Olsens Spring. Well 2 was located approximately 35.4 m (116 ft) northeast of Well 1. The wells were drilled using air rotary technology. Well 1 is 44.2 m (145 ft) deep (Figure 8). The upper 33.5 m (110 ft) of the rock strata consists of dolomite

(Nounan Formation), with interbedded shale and some limestone in the bottom 10.7 m (35 ft) of the well. Intervals of fine sand-sized fractions of dolomite were encountered above the water table in both holes, which may indicate solutional enlargement of fractures. However, it is not known if the dolomitic sand was formed in place or was deposited in fractures. Well 2 is 38.1 m (125 ft) deep, all in dolomite of the Nounan Formation (Figure 8). Both borings indicate highly jointed dolomite through most of the drilled intervals. Geophysical logging of the boreholes, which was originally proposed, was not conducted. Open-hole logging potentially could have provided information regarding solution channeling and fracture density at depth. However, such logging would not have been feasible due to a lack of competency of some materials and restrictions on drilling methods that could be used upgradient of the water-supply springs. Based on the results of geologic mapping, the stratigraphic information obtainable from the proposed logs was not considered essential at this location. Sufficient stratigraphic control was available from the samples obtained during drilling.

The wells were constructed using 10.2 cm (4 in) diameter schedule 40 PVC casing and screened immediately below the water table. An appropriately graded filter pack was installed within the screened zone, a bentonite seal was placed above the sand pack, and the annulus was grouted to the surface. The location and elevation of the top of each well casing and the top of the north collection box at Olsens Spring were surveyed with respect to the top of the south collection box.

Ground-water elevations in the wells and spring boxes were measured to 0.3 cm (0.01 ft) on three occasions following well installation. Continuous monitoring may have provided much valuable information regarding site hydrology. However, such monitoring was not feasible at this site due to a lack of security and

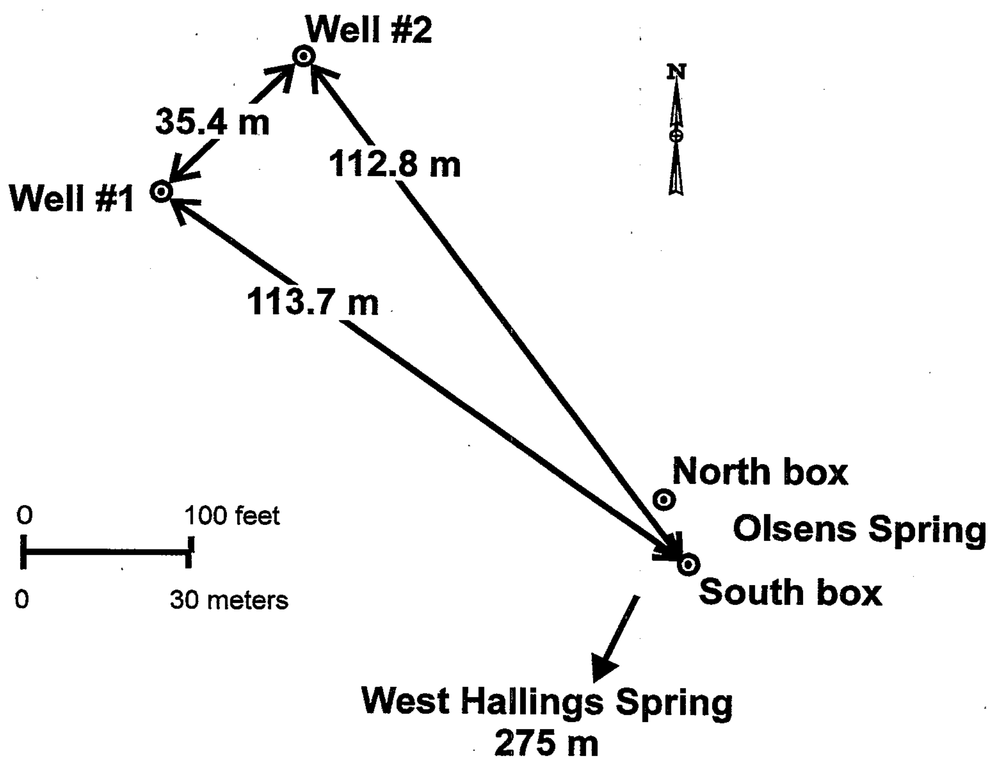


Figure 7. *Monitoring well locations relative to Olsens Spring.*

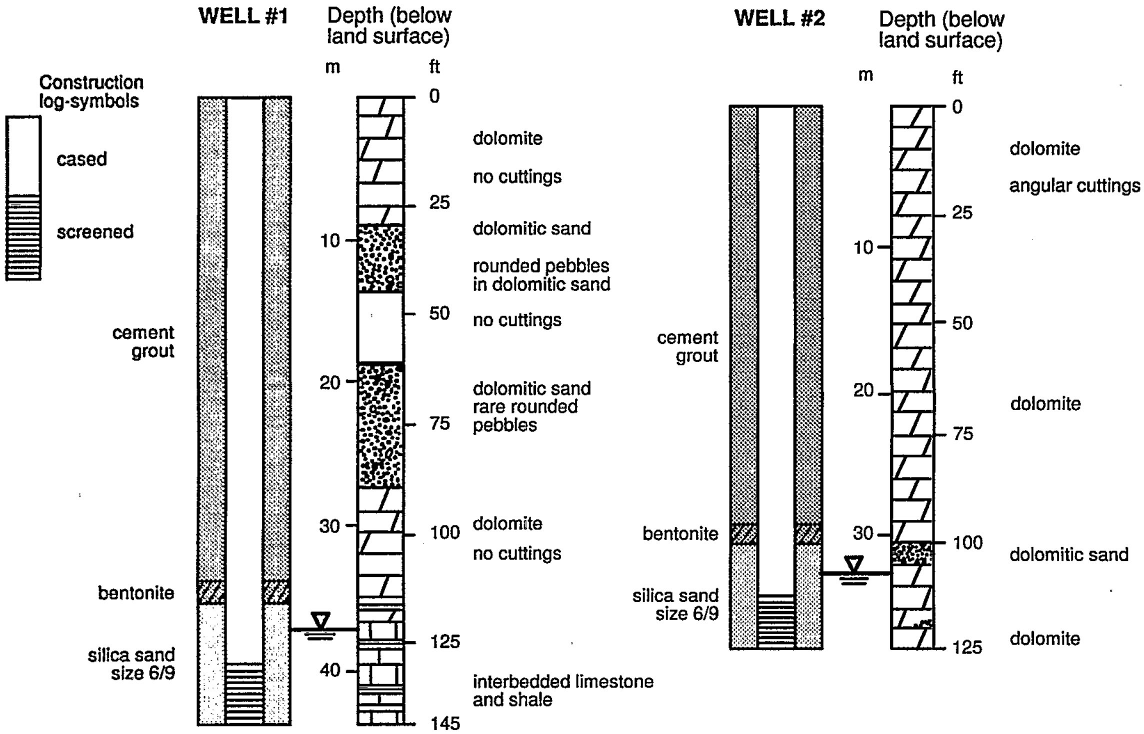


Figure 8. Geologic and construction logs of monitoring wells at Olsens Spring site.

concerns about vadalism of equipment. The hydraulic gradient near Olsens Spring was estimated using ground-water elevations in the two monitoring wells relative to that in the north spring collection box (Table 5). Based on the construction of this collection box, it appears that water elevation measured in this box may be representative of ground-water elevation at the discharge point. However, construction plans for the collection structures at the spring were not available. Therefore, gradient estimates using this monitoring point may not be reliable. Calculated maximum gradients ranged from 0.007 to 0.014 for the three sets of measurements that were obtained. The direction of maximum gradient was toward the south-southeast (152°) and did not vary significantly. The monitoring wells were installed during the final phase of field studies to allow use of information acquired during previous phases in siting the wells. Thus, limited information on temporal fluctuations in ground-water elevations was obtained.

Physical and Chemical Properties of Discharge

Field Parameters

Temperature, pH, specific conductance, and turbidity of water from Olsens Spring were monitored in the field on a monthly basis from November 1990 through October 1991 (Table 6). Due to the manner in which Olsens Spring was developed, it was not possible to monitor the discharge rate at the spring. Records of total discharge from seven springs in Mantua Valley, including Olsens and West Hallings Springs, were maintained by Brigham City personnel (Figure 9). These data indicate significant seasonal variation in discharge rate related to differences in recharge. Although data regarding daily fluctuations in discharge are not maintained,

Table 5. *Ground-Water Elevations in Wells Relative to Ground-Water Elevations in North Spring Collection Box and Estimated Hydraulic Gradients Near Olsens Spring*

Date	*Monitoring Point*	*Relative Elevation (m)*	*(ft)*	*Gradient (m/m)*	*Azimuth (°)*
22-Oct-92	*Well 1*	*0.59*	*1.92*	*0.007*	*152*
	Well 2	*0.65*	*2.12*		
4-Nov-92	*Well 1*	*0.66*	*2.17*	*0.007*	*152*
	Well 2	*0.73*	*2.40*		
17-May-93	*Well 1*	*1.24*	*4.06*	*0.014*	*151*
	Well 2	*1.35*	*4.42*		

discharge does not appear to noticeably fluctuate daily or following precipitation events (L. Pebley, Brigham City Corporation, personal communication to M. Jensen , 1993).

The temperature of the water varied from 9.5 °C to 10.6 °C (49.1 °F to 51.1 °F) during the 12-month sampling period (Figure 10). This small variation did not appear to be significant. Specific conductance did not change significantly during the year, ranging from 498 µmhos/cm to 528 µmhos/cm (Figure 11). The turbidity of the spring water was low, ranging from 0.04 NTU to 0.08 NTU (nephelometric turbidity units). The range of observed pH values was between 7.54 and 7.82, potentially displaying a slight seasonal trend. However, additional data would be required to evaluate the existence or significance of such a trend. The lack of distinct seasonal variations in water temperature, specific conductance, and turbidity indicate that residence time of the discharge from Olsens Spring is sufficiently long to mask the changes in these parameters due to seasonal variation in recharge. More frequent measurements (e.g., hourly or daily) of these parameters and discharge from individual springs would be required to determine whether some component of conduit flow results in rapid transport of surface water to the spring following precipitation.

Geochemical Characterization

Water samples from Olsens Spring were collected quarterly for one year and analyzed by standard laboratory techniques at the Utah State Health Laboratory for major ions, selected trace elements, radiological constituents, and total dissolved solids (Table 7). Water samples from West Hallings Spring and the two monitoring wells near Olsens Spring were also analyzed (Table 8). The laboratory methods that were used are listed in Table 9. Samples were collected from the wells for comparison with results from Olsens Spring and West Halling Spring in an attempt to identify differences in flow paths to these springs.

Ion balances for the first two samples from Olsens Spring have calculated errors of approximately 0% and 2%, respectively, indicating reliable analyses. Ion balances for the third and fourth analyses display differences between cations and anions of approximately 16% and 11%, respectively. Calcium and sulfate concentrations in the third and fourth analyses, respectively, were significantly different from previous values. These anomalous values may be the result of laboratory error. If the two values are changed to match values of the same constituents in the other samples, the calculated ion balances for the third and fourth samples would be in error by less than 5%. Based on the ion balances, and because there are no significant variations in other parameters, the two indicated values appear to be in error and may be laboratory errors. The total dissolved solids (TDS) concentra-

Table 6. *Field Parameters for Water Samples from Olsens Spring*

Date	Temperature (° C)	Specific Conductance (µmhos/cm)	Total Dissolved Solids (mg/1)	pH	Turbidity (NTU)
29-Nov-90	9.5	498	258	7.76	--
17-Dec-90	--	499	249	7.82	0.07
22-Jan-91	9.6	517	259	7.70	0.04
12-Feb-91	10.1	506	254	7.66	0.06
14-Mar-91	10.2	504	252	7.66	0.04
22-Apr-91	10.1	523	261	7.61	0.05
22-May-91	10.3	505	253	7.57	0.08
13-Jun-91	10.4	507	254	7.54	0.04
10-Jul-91	10.3	500	250	7.57	0.04
22-Aug-91	10.4	528	265	7.57	0.04
18-Sep-91	10.2	521	261	7.71	0.04
21-Oct-91	10.6	523	262	7.72	0.04

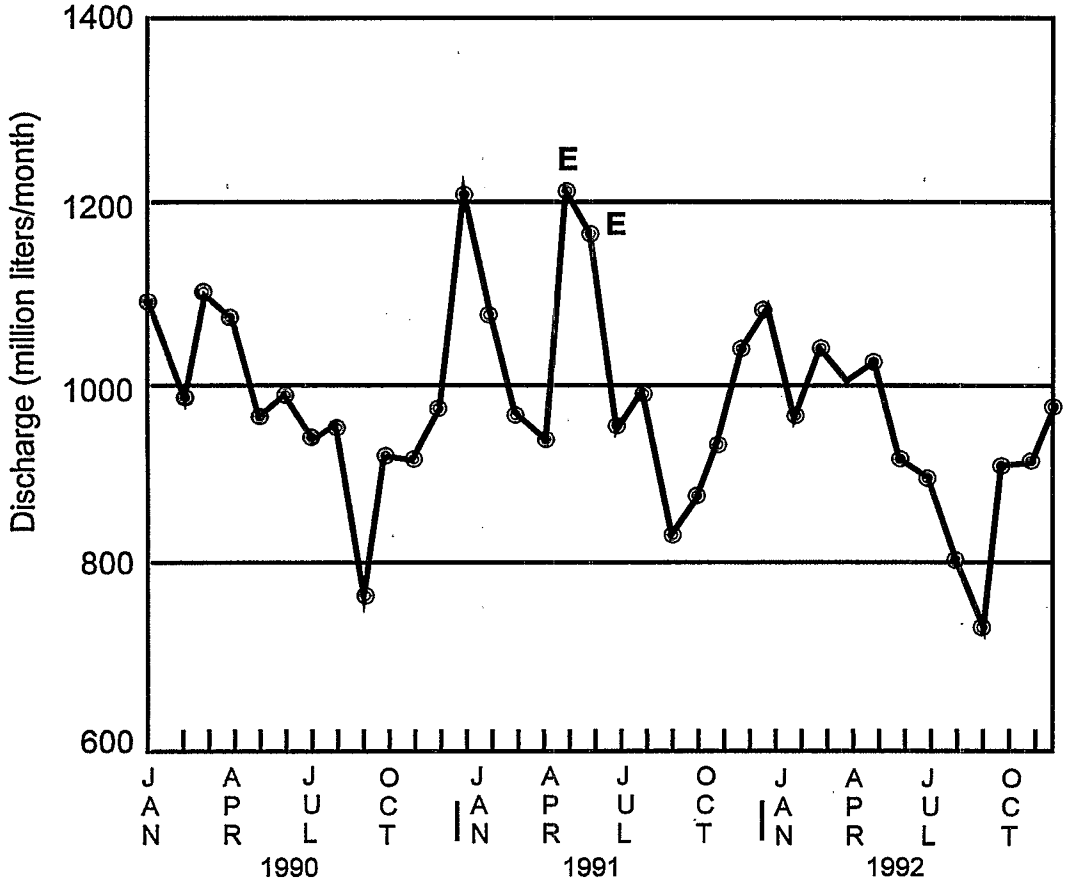


Figure 9. *Combined monthly flow of seven springs in Mantua Valley. E = estimated flow, due to malfunction of flow recorder. (B. Bingham and L. Pebley, Brigham City Corporation, personal communication to M. Jensen, 1993.)*

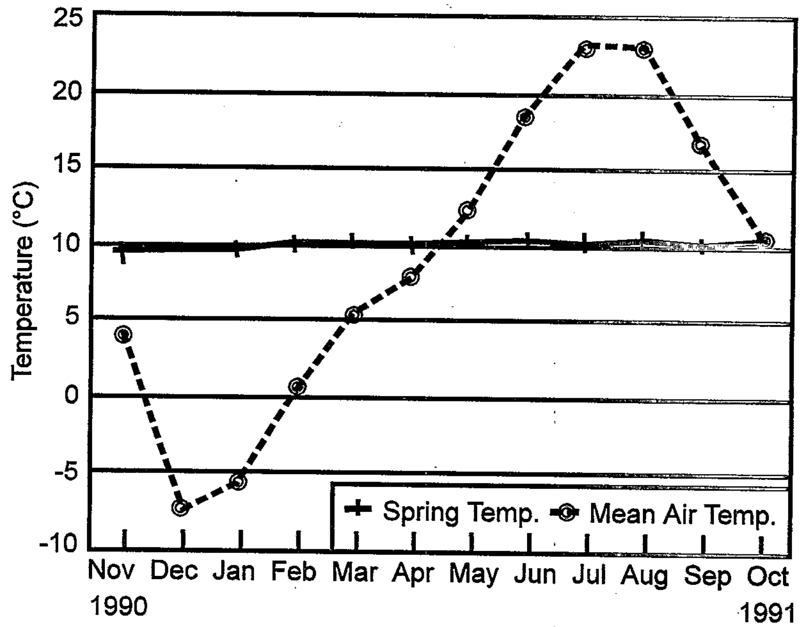


Figure 10. *Temperature of water samples from Olsens Spring and mean monthly air temperature in Brigham City (various personnel, Utah Climate Center, personal communication to M. Jensen, 1993).*

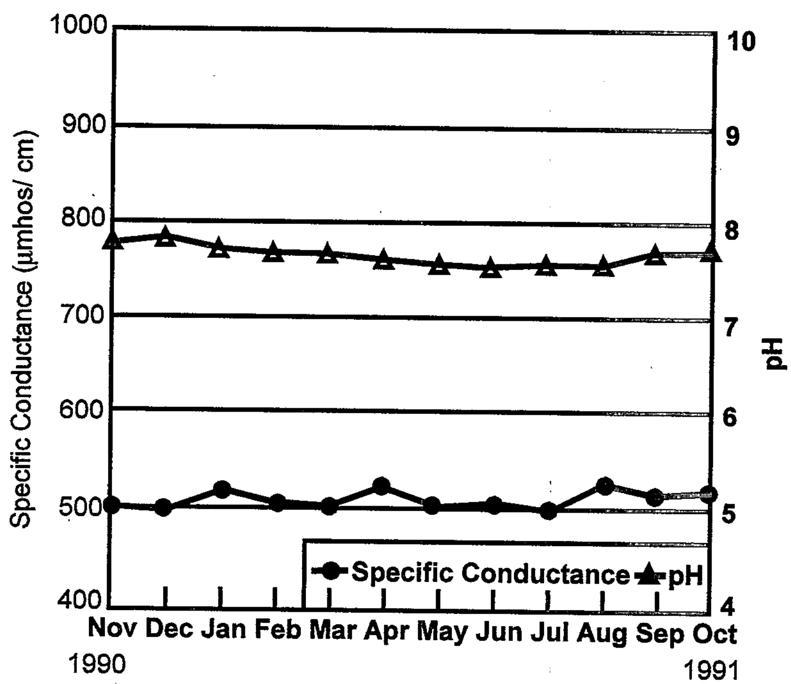


Figure 11. *Specific conductance and pH of water samples from Olsens Spring.*

Table 7. *Analytical Results for Water Samples from Olsens Spring*

Sample Date	17-Dec-90	14-Mar-91	13-Jun-91	18-Sep-91
Calcium (mg/l)	51	53	33*	50
Magnesium (mg/l)	26	27	26	26
Potassium (mg/l)	<1	<1	<1	<1
Sodium (mg/l)	15	15	16	17
Bicarbonate(mg/l)	268	269	257	267
Carbonate (mg/l)	0	0	0	0
Chloride (mg/l)	25	26	28	29
Silica (mg/l)	9	9.8	9.4	9.4
Sulfate (mg/l)	10	10	9.6	37.4*
Nitrate+Nitrite(mg/l)	1.25	1.40	1.39	1.31
Phosphate(mg/l)	0.02	0.34	0.02	0.01
Iron (mg/l)	<0.02	<0.02	<0.02	<0.02
Alkalinity(mg/l)	220	221	211	219
Specific Conductance(μmhos/cm)	492	486	492	509
Total Dissolved Solids (mg/l)	370*	268	274	282
Gross Alpha (pc/l)	<1	3	<1	<1
Gross Beta (pc/l)	<5	<5	<5	<5
Cations (meq/l)	5.4	5.5	4.5	5.4
Anions (meq/l)	5.4	5.4	5.2	6.0

* These values appear to be errors.

Table 8. *Analytical Results for Water Samples from West Hallings Spring and Monitoring Wells Near Olsens Spring*

Sample Date	West Hallings Spring 28-May-92	Well 1 4-Nov-92	Well 2 4-Nov-92
Calcium (mg/l)	44	51	51.5
Magnesium (mg/l)	22	26.2	26.3
Potassium (mg/l)	1.2	<1	<1
Sodium (mg/l)	8.7	14.7	15.1
Bicarbonate(mg/l)	242	271	272
Carbonate (mg/l)	0	0	0
Chloride (mg/l)	11	24	25
Silica (mg/l)	11.7	9.5	9.1
Sulfate (mg/l)	15.9	10.2	10.2
Nitrate+Nitrite (mg/l)	0.9	1.5	1.47
Phosphate (mg/l)	<0.01	0.02	0.02
Iron (mg/l)	<0.02	0.02	<0.02
Alkalinity(mg/l)	198	222	223
Specific Conductance(μmhos/cm)	402	504	506
Total Dissolved Solids (mg/l)	222	268	268
Gross Alpha (pc/l)	<2	<2	<2
Gross Beta (pc/l)	<10	<10	<10
Cations (meq/l)	4.4	5.4	5.4
Anions (meq/l)	4.6	5.3	5.3

Table 9. Analytical Parameters and Laboratory Methods

Analyte	Laboratory Method
Calcium	EPA 200.7
Magnesium	EPA 200.7
Potassium	EPA 200.7
Sodium	EPA 200.7
Bicarbonate	Calculated
Carbonate	Calculated
Chloride	Standard Method 407A
Silica	EPA 370.1
Sulfate	EPA 375.2
Nitrate+Nitrite	EPA 353.2
Phosphate	EPA 365.1
Iron	EPA 200.7
Alkalinity	EPA 310.1
Specific Conductance	EPA 120.1
Total Dissolved Solids	EPA 160.1
Gross Alpha	EPA 900.0
Gross Beta	EPA 900.0

tion of the sample obtained in December 1990 also appears to be in error based on the lack of correlation with other parameters. Ion balances for samples from West Hallings Spring and the monitoring wells have balance errors between 2% and 5% (Table 8).

Major ion concentrations from these analyses were plotted on a trilinear diagram as developed by Piper (1944) (Figure 12). Based on the hydrochemical facies of Back (1966), the trilinear diagram indicates a calcium magnesium bicarbonate type ground water. Bicarbonate is the dominant anion in all of the analyses. The predominance of bicarbonate and calcium is indicative of the carbonate aquifer through which the water flows (Hem, 1985). No significant temporal variations in water type were noted. The small shifts in relative ion concentrations observed in Figure 12 are due to the anomalous values which are believed to be errors. Analyses of water from the two monitoring wells correlate closely with samples from Olsens Spring. Several of the constituents in the sample from West Hallings Spring are present at relatively lower concentrations than in samples from Olsens Spring or the monitoring wells. However, the samples were not obtained synchronously and temporal variations in constituent concentrations may occur.

Saturation Indices and Calcium/Magnesium Molar Ratios

Analytical results of discharge from Olsens Spring were used by L. Spangler (U.S. Geological Survey, Water Resources Division, Salt Lake City, Utah) to determine the relative states of saturation with respect to calcite and dolomite and calcium/magnesium (Ca/Mg) molar ratios. Saturation indices for calcite and dolomite (Table 10) were calculated using the U.S. Geological Survey WATEQ4F model (Ball and Nordstrom, 1991). As analytical results for calcium (June 1991), sulfate (September 1991), and total dissolved solids (December 1990) are believed to be errors, alternate values were used for these calculations. Average concentrations for calcium, sulfate, and total dissolved solids in the remaining three samples were used to calculate mineral satura-

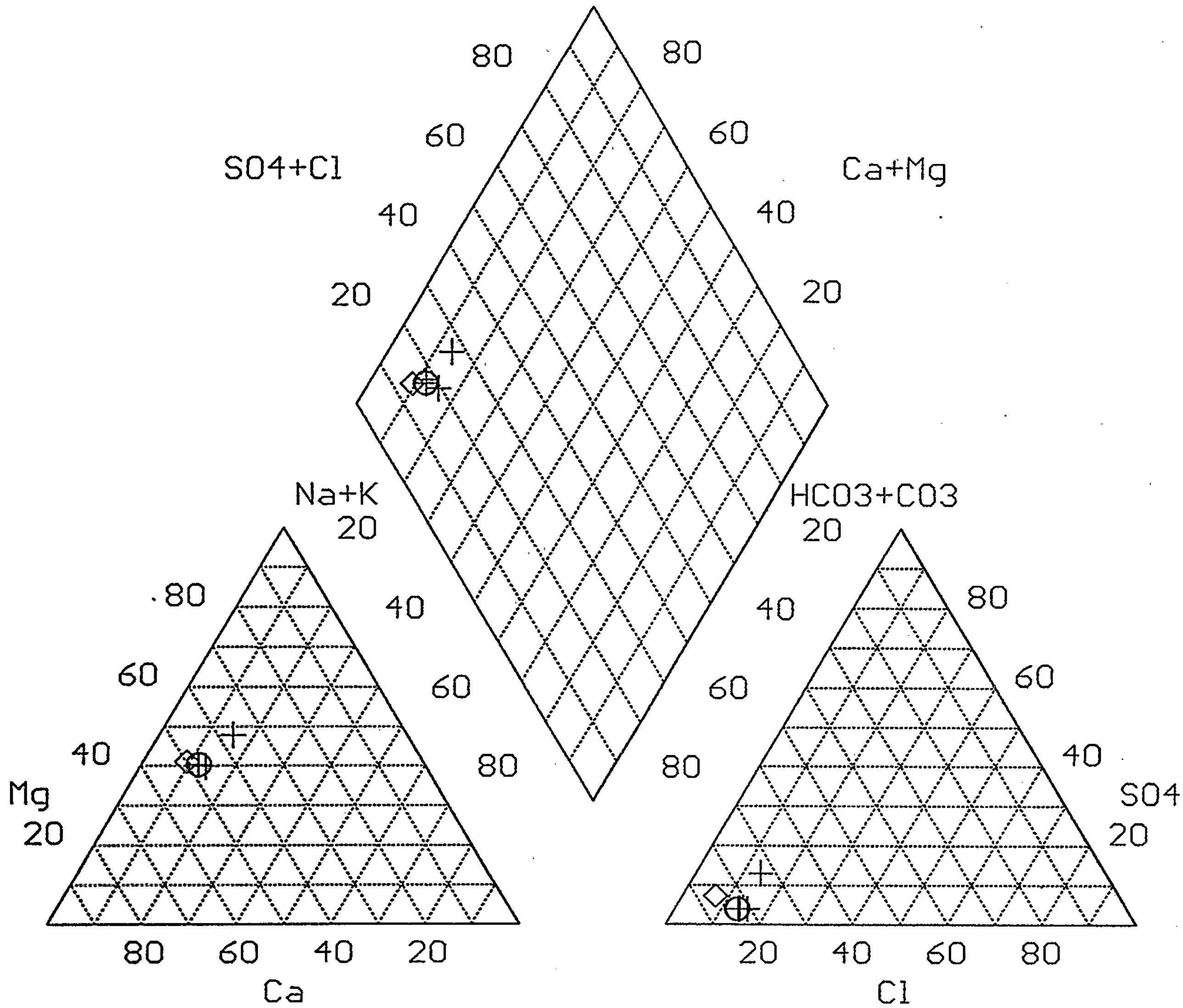


Figure 12. *Trilinear diagram depicting geochemical analyses of water samples from OlsensSpring (+), monitoring wells (O), and West Hallings Spring (◊).*

tion indices. Although there is uncertainty in using alternate values, the alternate values are believed to be closer than the reported laboratory values to the actual values and resulted in better ion balances. It is noted that no errors are suspected for the samples collected on March 14. The results of calculations using these data are assumed to be reliable. The results from this sample indicate calculations using the alternate values are reasonable and may be representative of site conditions. However, results of calculations using these data should be considered preliminary. Additional data would be required to test significant conclusions.

Saturation indices for calcite ranged from -0.008 to +0.272, indicating that most samples were saturated or near equilibrium with respect to calcite (Table 10). This implies a relatively long residence time indicative of flow through a system of fractures that may have undergone relatively minimal amounts of solutional enlarge-

Table 10. *Calcite and Dolomite Saturation Indices and Ca/Mg Molar Ratios Calculated for Water Samples from Olsens Spring*

Date	Calcite	Dolomite	Ca/Mg Molar Ratio
17-Dec-90	0.272	-0.242	1.19
14-Mar-91	0.133	-0.515	1.19
13-Jun-91	-0.008	-0.801	1.21
18-Sep-91	0.156	-0.461	1.17

ment. Ground water in systems primarily characterized by conduit flow (solutionally enlarged fractures) generally has a relatively short residence time and mineral saturation indices indicating undersaturation (more negative values) (White, 1988). Dolomite saturation indices ranged from -0.242 to -0.801, indicating undersaturation. However, hydrologic (e.g., residence times) and geochemical factors inhibit dolomite from reaching saturation, even in stratigraphic sections composed largely of dolomite (White, 1988).

The calcite saturation index (Figure 13) shows a trend suggestive of seasonal variations in saturation that may be related to discharge. As discharge increases in the spring due to increased recharge, saturation of the ground water with respect to all minerals would be expected to decrease as the result of decreased residence time. In support of this interpretation, it is noted that discharge from seven springs in Mantua Valley (Figure 10) peaked during June 1991. However, longer term monitoring of water chemistry and accurate measurements of discharge from only Olsens Spring are required to verify this trend.

Calcium/magnesium molar ratios can be used to help understand flow paths and source areas for hydrologic systems in carbonate settings (Langmuir, 1971). Ground water that has moved through or been in contact with formations composed predominantly of dolomite has a Ca/Mg molar ratio of approximately 1.00. Water with a Ca/Mg molar ratio exceeding 3.00 has been in contact with, or flowed through rocks composed mostly of limestone. Ratios between 1.00 and 3.00 indicate groundwater flow through interbedded dolomite and limestone, calcareous dolomite, or dolomitic limestone. Ratios of water samples obtained from Olsens Spring (Table 10) ranged from 1.17 to 1.21, indicating that the ground water flowed through formations composed largely of dolomite. Calcium/magnesium ratios calculated for West Hallings Spring discharge and the monitoring wells (Table 11) are in the same range as values for water from Olsens Spring, also implying a ground-water flow system that has been in contact with dolomite or dolomitic lithologies (L. Spangler, U.S. Geological Survey, Water Resources Division, written communication to M. Jensen, 1993). Differences in flow paths to the springs could not be inferred from these data.

Two water samples were collected from Olsens Spring for tritium analysis, one each in June and October 1991. The tritium activities in these samples were 16.8 (±2.8) TU and 22.4 (±2.8) TU. These values indicate that the average age of ground water in these samples is less than approximately 40 years and, possibly, less than 20 years (Fetter, 1988; Hendry, 1988).

Geochemical characterization of water from Olsens Spring, West Hallings Spring, and the monitoring wells supports several hypotheses and conclusions concerning ground-water residence time and aquifer lithology:

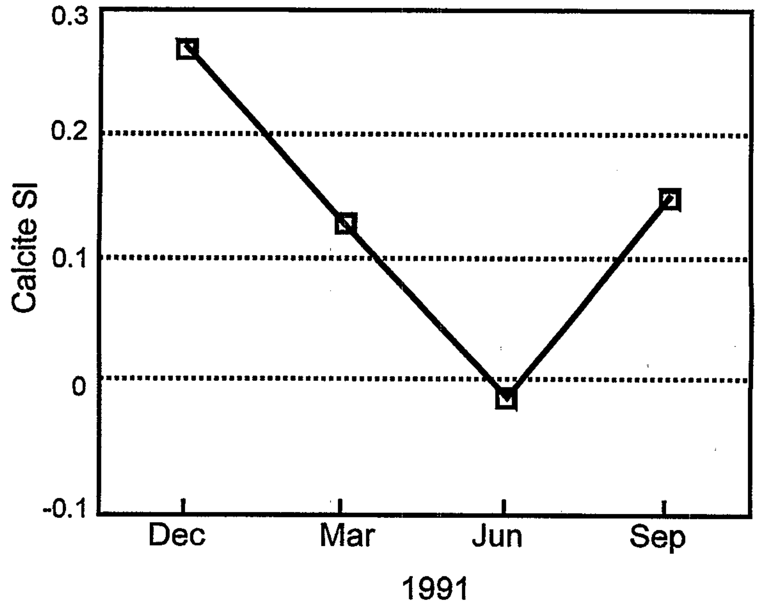


Figure 13. Seasonal variation of calcite saturation index in water samples from Olsens Spring for the sampling period of December 1990 to September 1991.

1. Lack of consistent seasonal variation in physical/chemical parameters, consistently low turbidity, and saturation with respect to calcite indicate that the ground water discharging at Olsens Spring may be migrating predominantly through fractures that have not undergone significant solutional enlargement. However, these observations may also be explained by other conditions such as a lack of recharge from only a few discrete points or by existence of substantial storage in a system with some component of conduit flow (Quinlan and others, 1991). Additional studies, including more frequent monitoring of parameters such as specific conductance, discharge, and turbidity, would be required to better define potential conduit-flow contributions.

2. Potential seasonal fluctuation in the mineral saturation indices may indicate seasonal decreases in residence time related to increased recharge. This is also supported by the increased hydraulic gradient estimated from ground-water elevation data obtained during May 1993.

3. The formations through which ground water migrates are composed largely of dolomite.

4. The average ground-water age is relatively young, less than 40 years. Additional studies (e.g., Busenberg and Plummer, 1992; Solomon and Sudicky, 1991) may provide more information regarding the average time period since recharge.

Table 11. Calcium/Magnesium Molar Ratios in Water Samples from West Hallings Spring and Monitoring Wells Near Olsens Spring

Site	Ca/Mg Ratio
West Hallings Spring	1.21
Well 1	1.19
Well 2	1.18

Catchment Area Calculations

A value of the average annual precipitation of 79 cm (30.9 in) was used (Table 4) for purposes of catchment area estimation, assuming that the climatic setting is similar to the location at Pineview Dam. However, it is noted that these recharge area calculations are sensitive to the recharge estimate, which is not well defined. Thus, the results are subject to significant uncertainty. Data from Bjorklund and McGreevy (1974) indicate that evapotranspiration on rangeland and mountainous areas at the elevation of Olsens Spring is about 35 cm (13.6 in) per year. If it is assumed that the difference between precipitation and evapotranspiration is recharge to ground water within the catchment area, then recharge would be about 44 cm (17.3 in) per year. This represents about 56% of total precipitation and is considered to be a significant overestimate of recharge for the assumed precipitation. Insufficient data were available to conduct a detailed water balance. In order to bracket the catchment area calculations, a recharge value of 10% of total precipitation was used as a lower bound. Based on the fractured nature of the site, it is believed that 10% of available precipitation may represent the minimum recharge at this site. Use of the minimum recharge estimate will result in conservative calculation of maximum catchment areas. The discharge of Olsens Spring is about 1700 l/min (1.0 ft^3/s). Using the method of Todd (1980), a plot of catchment area as a function of estimated recharge rate and discharge for Olsens Spring indicates that a catchment area ranging from about 2 km^2 (0.8 mi^2) to 10 km^2 (4 mi^2) would be necessary to supply enough water for the spring to yield 1,700 l/min (450 gal/min) (Figure 14). West Hallings Spring, which is located 270 m (900 ft) southwest of Olsens Spring, discharges from a limestone unit that is stratigraphically below the discharge point of Olsens Spring. However, this spring is within the same surface drainage basin. Total discharge from both springs is about 8100 l/min (4.8 ft^3/s). A catchment area ranging from about 10 km^2 (4 mi^2) to 54 km^2 (21 mi^2) would be required to provide enough water for the total discharge of these two springs. This indicates that the zone of contribution for the two springs may be significantly larger than the topographic basin surrounding the springs, which has a surface area of about 17 km^2 (6.5 mi^2). Mud Spring also may receive at least a part of its recharge from the same surface drainage basin, but there are no data available to determine to what degree Mud Spring affects Olsens Spring. If discharge from Mud Spring was included in the catchment area calculations, the catchment area would be even larger than that calculated for Olsens and West Hallings Springs. In addition, if the actual precipitation and, thus, recharge rates were less than the assumed values, the required catchment area would increase.

Tracer Study

A tracer study was conducted to estimate ground-water velocity in the vicinity of the springs. The springs were monitored for background fluorescence for ten consecutive days before injecting 500 ml of rhodamine WT dye into Well 1, which is located 114 m (373 ft) from the Olsens Spring collection area (Figure 7). Approximately 1900 liters (500 gal) of water were injected to force the dye out of the well and into the aquifer.

Dye detection was by adsorption on activated charcoal contained in nylon screen packets that were placed in Olsens Spring, the combined flow of Olsens Spring and West Hallings Spring, and in Mud Spring, which is northeast of Olsens Spring (Figure 3). Due to the design of the water system, waters from Olsens Spring and West Hallings Spring mix before water from West Hallings Spring can be sampled under normal conditions. The charcoal detectors were initially changed as frequently as every 4 days, and were left in place for periods as long as 40 days as the test continued. After removal from the springs, the dye detectors were eluted using a mixture of ammonium hydroxide, 1-propanol, and distilled water and the solution was analyzed on a Turner

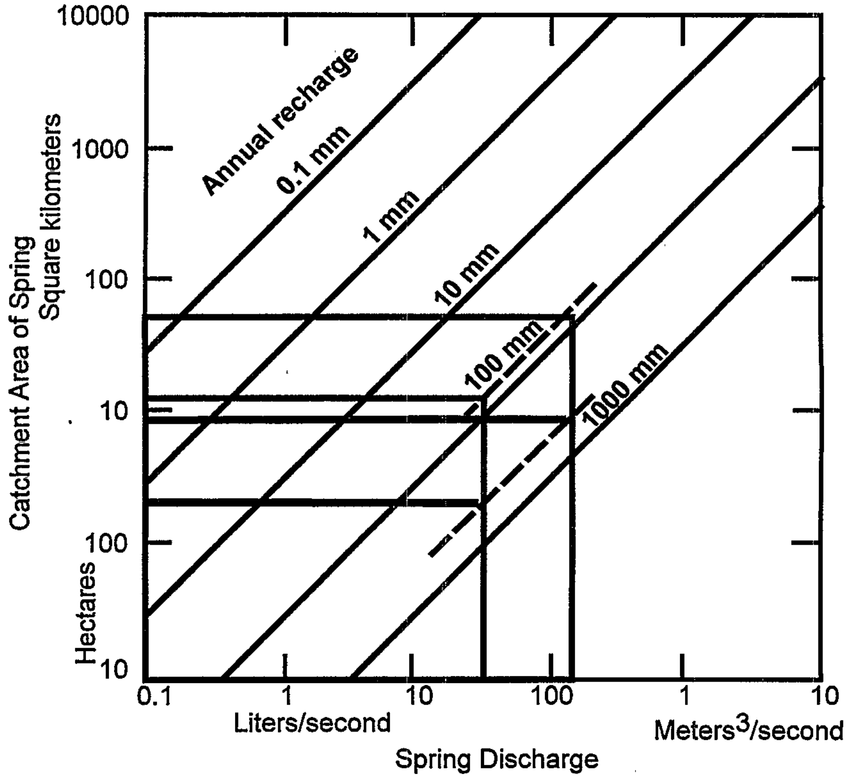


Figure 14. *Estimation of catchment areas for Olsens Spring and combined Olsens Spring and West Hallings Spring. (Modified from Groundwater Hydrology, D.K. Todd, copyright © 1980, John Wiley & Sons, Inc., reprinted by permission of John Wiley & Sons, Inc.)*

filter fluorometer for the presence of dye (L. Spangler, U.S. Geological Survey, Water Resources Division, written communication to M. Jensen, 1992). Analysis of dye recovery hydrographs (Figure 15) indicates that dye was present on charcoal samples from the combined flow of Olsens Spring and West Hallings Spring. Dye was not detected in discharge from only Olsens Spring or from Mud Spring during the test. This indicates that dye only appeared in the discharge from West Hallings Spring. First arrival of the dye in the combined flow of the two springs appeared to be between 12 days and 18 days after dye injection. Maximum response in the combined flow of the two springs was approximately two months after dye injection. The quantity of dye in the charcoal detectors in the combined discharge appears to have gradually decreased since that time.

Fluctuations in instrumental intensity producing secondary peaks during the test are related to length of time that detectors were in the springs and are not the results of dye pulses. Peaks are also due, in part, to adsorption of organic material. More consistent and accurate results would be obtained using an automatic water sampler and fluorometer.

West Hallings Spring is approximately 350 m (1150 ft) southwest of Well 1, where the dye was injected. Based on first inferred arrival of the dye and the straight-line distance from the spring to the well, ground-water velocity potentially ranged from approximately 29 m/d to 19 m/d (96 ft/d to 64 ft/d) (i.e., dye arrival between 12 days and 18 days). Assuming the first arrival of the dye was in 15 days, the velocity would be 23 m/d (77 ft/d). This calculated velocity range is subject to uncertainty resulting from assumptions involved in the calculations (e.g., actual path length and arrival time). In addition, this velocity represents conditions at only

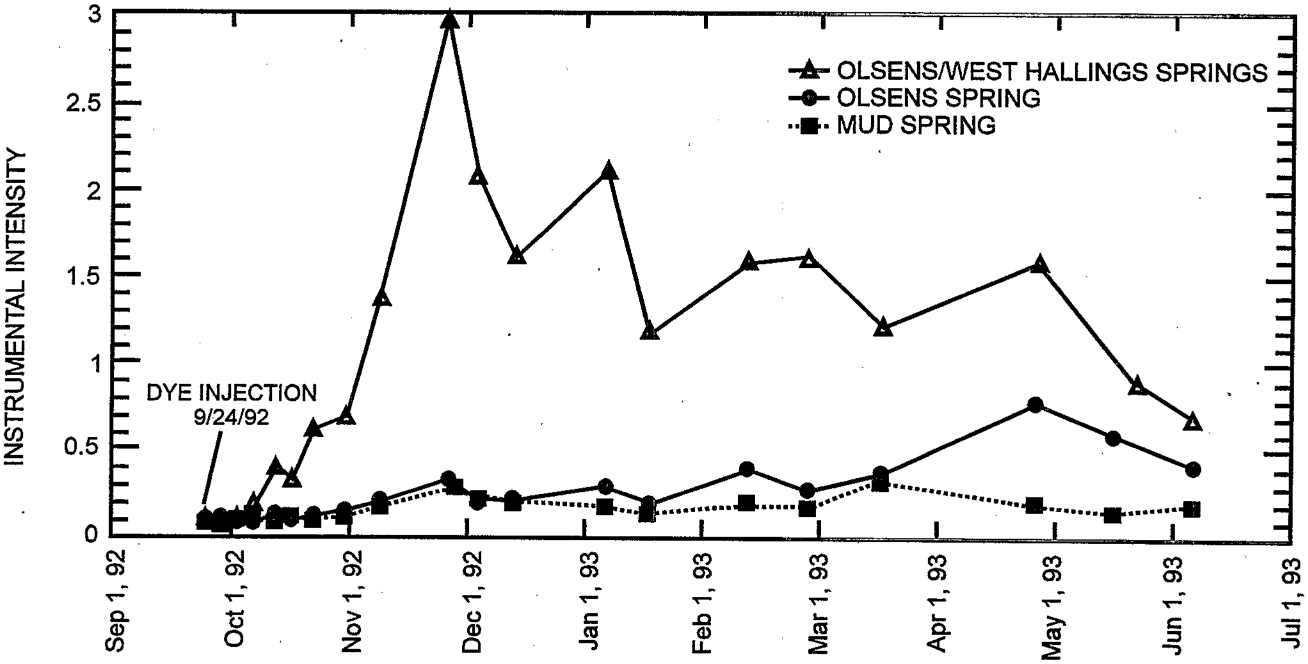


Figure 15. *Relative instrumental intensity of eluted samples from Olsens Spring, Mud Spring, and the combined discharge of Olsens Spring and West Hallings Spring.*

one point in time. Seasonal fluctuations in hydraulic gradients may result in different ground-water velocities. The azimuth from Well 1 to West Hallings Spring is approximately 190° and may indicate preferential flow along north to northeast striking joints. This general flow direction appears to be consistent with the direction of maximum gradient indicated from water levels in the monitoring wells and spring collection area (Figure 7) within the limits of the uncertainty involved in each of these determinations. Results of this test indicate that the ground-water velocity is relatively high in the vicinity of the springs and that Well 1 appears to be within the zone of contribution to West Hallings Spring.

Hydrogeologic Conceptual Model

The conceptual model for site hydrology is that of a fracture-type spring (Figure 1) dominated by flow through a fracture network and receiving recharge from a relatively large area. The potential recharge area may be larger than the surface drainage area in which the spring is situated, indicating that these springs may be more regional discharge points. This implies that the zone of contribution may extend beyond the boundaries of the topographic basin. Additional studies would be required to confirm these hypotheses.

The evidence for this model includes highly fractured bedrock units, stable field parameters, relatively stable water chemistry, very low turbidity, estimated size of required catchment basin, discharge that does not appear to fluctuate daily or after storm events, and lack of observed surface karst features. In addition, the carbonate units at this site are primarily dolomitic and diffuse-flow conditions (i.e., flow through closely-spaced fractures interconnected, potentially, on the scale of centimeters) are common in dolomitic bedrock (Fetter, 1988).

The only surface water body in the vicinity of the springs is Mantua Reservoir, which is located approximately 610 m (2000 ft) southeast of Olsens Spring and is outside the surface drainage basin. Surface water elevations in Mantua Reservoir were several feet lower than ground-water elevations in the springs and monitoring wells based on records maintained by Brigham City Corporation (Alan Wright, Brigham City Corporation, personal communication to M. Jensen, 1997). The reservoir is not considered to have significant influence on the zone of contribution to the springs in this study.

Based on the relatively high velocity, ground-water movement through the aquifer appears to be predominantly through fractures. The system appears to be anisotropic due to the existence of preferred joint orientations. More extensive aquifer testing may aid in determining hydraulic conductivity tensors. Rock units throughout the surface drainage area may be sufficiently fractured to potentially allow the aquifer to behave as a porous medium at the scale of the zone of contribution. However, this hypothesis is difficult to assess. A study conducted by the Wisconsin Geological and Natural History Survey (1991) describes subjective criteria for determining whether fractured-rock aquifers behave as porous media. These criteria are based on fracture density, aquifer testing, spatial variations in water quality, and potentiometric information. Sufficient data from subsurface investigations are not available, however, to make this determination at this site.

Aquifer Tests

A short-term, constant-rate pumping test was performed in each monitoring well at Olsens Spring. A step-drawdown test was performed to determine sustainable discharge rates prior to the tests. Each well was pumped for two hours and drawdown was monitored in the pumping well and in the observation well. Maximum drawdown in the pumping wells was approximately 0.3 m (1 ft) during each test. Significant drawdowns were not noted in the observation wells. This probably indicates that the aquifer was not stressed sufficiently to expand the cone of depression to the monitoring well during the short test. However, the tests also may have been complicated by a lack of hydraulic communication between wells due to a lack of interconnecting fractures.

Complications with pumping and metering equipment were encountered during the pumping portion of the tests. Although the discharge rate appeared relatively stable by visual observation, the flowmeter used to monitor discharge rates fluctuated significantly during the first 10 minutes of the test conducted in Well 2. The discharge rate indicated by the meter during the remainder of the test varied approximately 8% with an average rate of 132 l/min (35 gal/min). The average discharge rate for the test performed in Well 1 was 285 l/min (70 gal/min) with a variation of approximately 11%. In addition, it appeared that Well 1 may have been poorly developed prior to the test as well efficiency appeared to increase during the test. Based on these complications, only recovery data from the pumping wells (Figure 16) and time-averaged pumping rates were analyzed to obtain a preliminary estimate of transmissivity. As it did not appear that actual discharge fluctuated significantly during these tests, methods for analyzing recovery from stepped discharge tests (e.g., Kawecki, 1993; Kruseman and de Ridder, 1990) were not applied.

The recovery data were interpreted using several solutions applicable to porous media. It is recognized that the aquifer may not behave as a porous medium at the scale of these tests. Therefore, the results of these tests are viewed as preliminary due to uncertainties associated with well losses, equipment complications during the tests, subsurface anisotropy and heterogeneity, and application of porous-media assumptions. Extensive

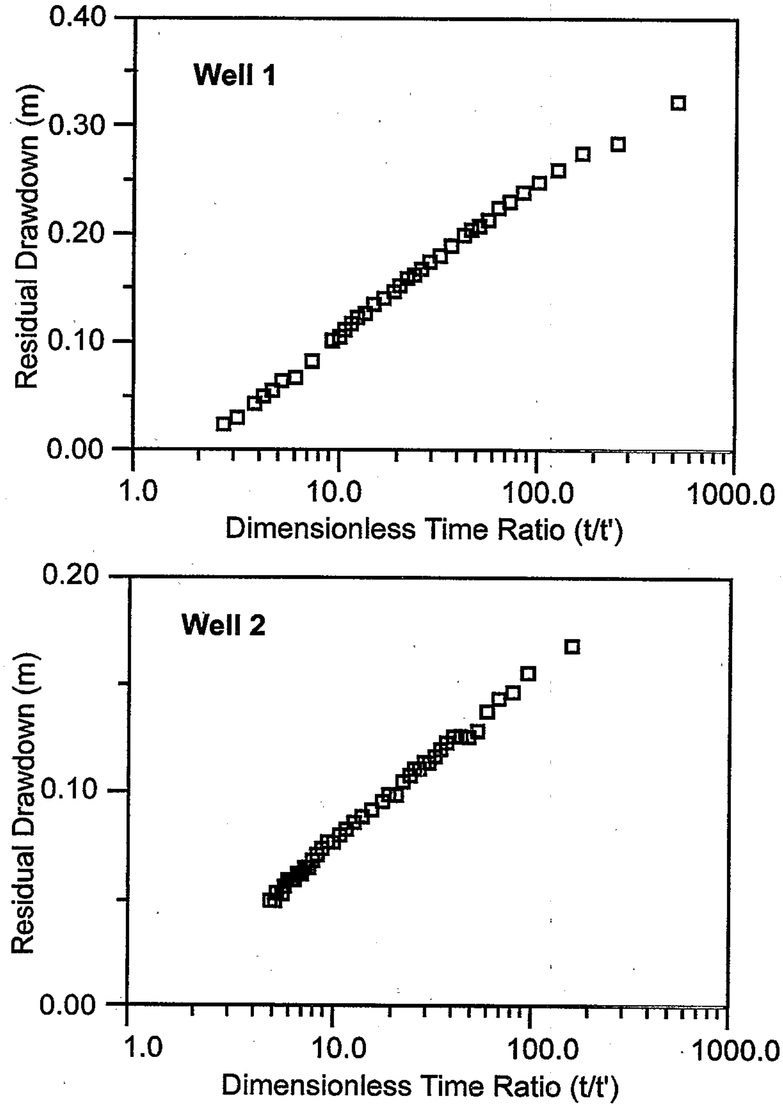


Figure 16. *Residual drawdown plotted as function of the ratio of the elapsed time since pumping began (t) to the elapsed time since pumping ended (t').*

aquifer testing beyond the scope of this investigation would be required to test the porous-media assumptions and better define hydraulic parameters. The interpretation methods used and transmissivity values calculated from these tests are presented in Table 12.

Calculated transmissivity ranged from 270 m^2/d (2900 ft^2/d) to 550 m^2/d (5900 ft^2/d). Based on these results, the aquifer in the area of these wells appears to be relatively transmissive, in support of qualitative hypotheses derived from hydrogeologic mapping. Calculation of hydraulic conductivity from these data involves additional assumptions, including aquifer thickness influenced during pumping. In a porous medium with fully penetrating wells, this thickness would generally be the saturated thickness within the well screen. However, this thickness is unknown in a fractured setting with no definable confining units and may be considerably greater than the screened interval of the well.

Table 12. Interpretation Methods and Estimated Transmissivity from Aquifer Tests in Monitoring Wells at Olsens Spring

Solution	Well 1	Well 2
Theis (1935)	550 m^2/d (5900 ft^2/d)	270 m^2/d (2900 ft^2/d)
Theis Recovery Method (Theis, 1935)	510 m^2/d (5500 ft^2/d)	440 m^2/d (4700 ft^2/d)
Neuman (1975)	440 m^2/d (4800 ft^2/d)	Not Applied
Jacob (1950)	540 m^2/d (5800 ft^2/d)	330 m^2/d (3600 ft^2/d)

Zone of Contribution and Delineation of Spring Protection Zones

General site hydrogeology and certain geologic and hydrologic controls on ground-water flow to the springs were qualitatively defined using the mapping techniques applied in this study. However, definitive ground-water flow boundaries were not delineated. In addition, the zones of contribution to Olsens Spring, West Hallings Spring, and Mud Spring could not be differentiated using these techniques. At a minimum, it appears that the zone of contribution to these springs may encompass the entire surface drainage basin (Figure 17), an area of approximately 17 km^2 (6.5 mi^2), and that ground-water velocities may be relatively high near the springs. The results also indicate that the zone of contribution may be significantly larger than this area. Extrapolating the velocity calculated from the tracer study, the time-of-travel from the northern extent of the topographic basin to the area of the springs would be less than 200 days. Similar results are obtained using the limited information available from the pumping tests and estimates of hydraulic gradient in the area of the springs. A high degree of uncertainty exists in applying these velocity estimates over this distance as the average velocity is probably significantly lower. However, this analysis indicates that the topographic basin may not be sufficient for use as the 15-year time-of-travel boundary specified in the regulations promulgated by the State of Utah. Additional studies, potentially including installation of a piezometric network, would be required to better determine the zone of contribution and protection zones outside the surface drainage basin.

SHEEP SPRING

Location and Description

Sheep Spring is located in southwestern Utah, approximately 2.7 km (1.7 mi) northwest of St. George and 2.0 km (1.25 mi) east of Santa Clara in Washington County (Figure 18 and Table 13). This area of Utah is near the western edge of the Colorado Plateau physiographic province (Stokes, 1986). The Santa Clara city water system serves about 1,500 people, and uses Sheep Spring as a water source. The spring is at an elevation of about 890 m (2920 ft). Sheep Spring discharges approximately 7.6 l/min (2 gal/min), making it a sixth magnitude spring (Table 2).

Sheep Spring was developed by digging a collection tunnel about 10.7 m (35 ft) eastward into a low bench. The collection tunnel is roughly 1 m (3 ft) wide and 1.8 m to 2.1 m (6 ft to 7 ft) high. Ground water seeps and flows from the north wall of the collection tunnel, ponds on the floor of the tunnel, and exits through a discharge pipe to the water system.

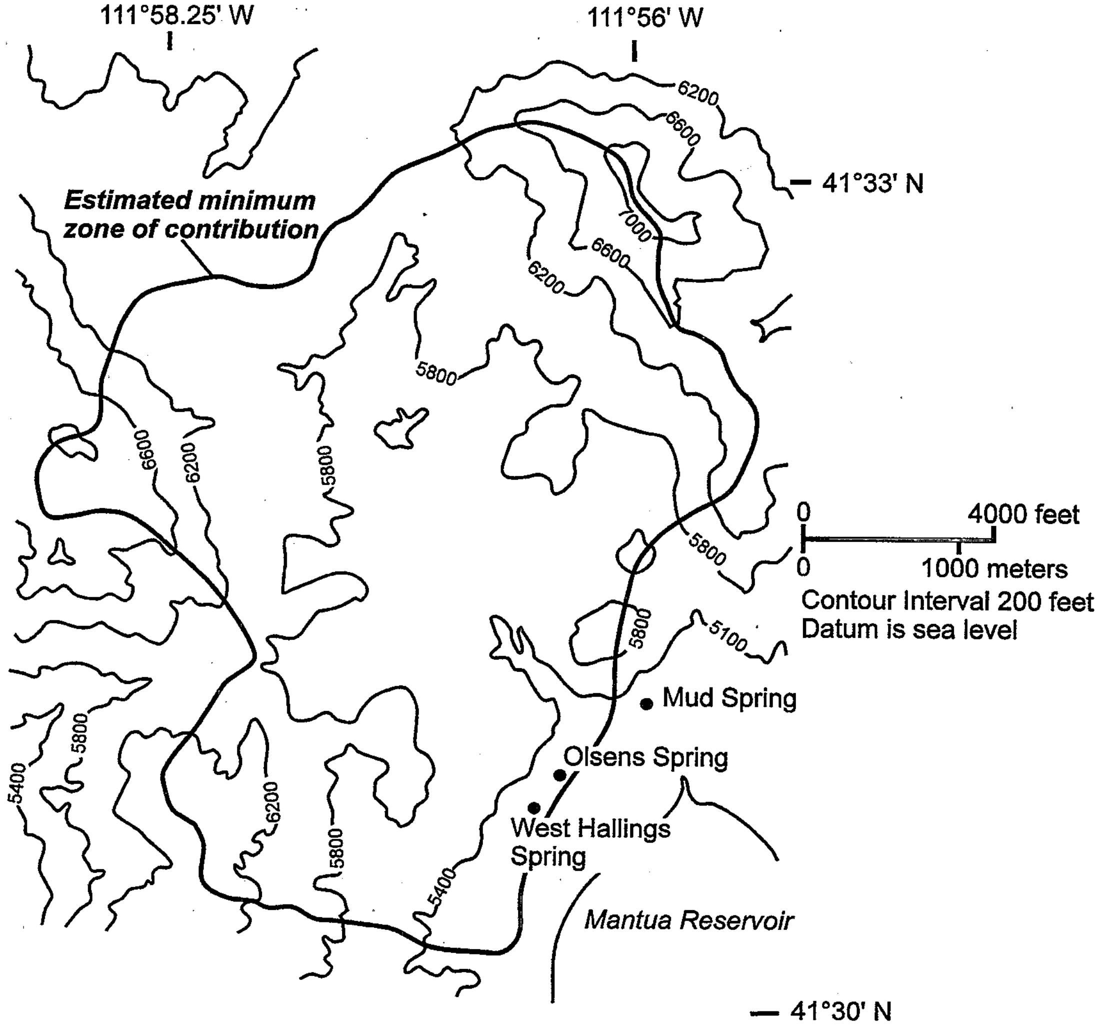


Figure 17. *Surface drainage basin and estimated minimum zone of contribution to springs in this basin.*

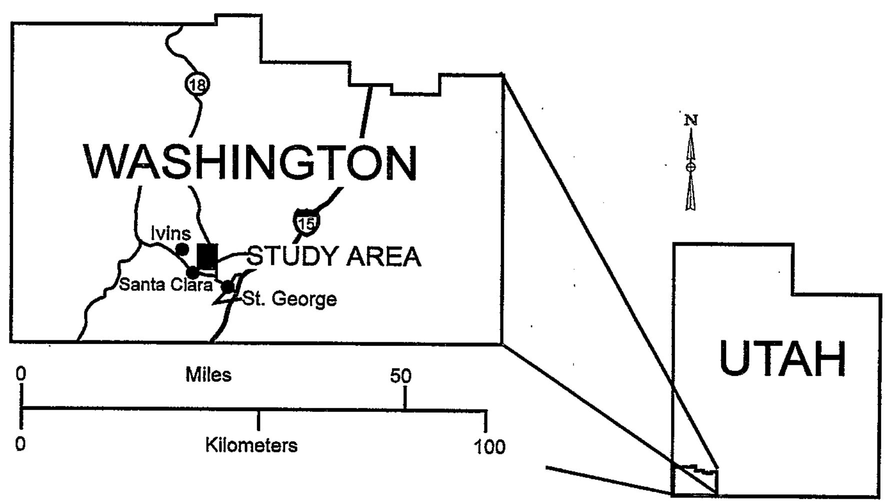


Figure 18. *Location of Sheep Spring study area.*

Table 13. Location and Elevation of Sheep Spring and Associated Monitoring Wells Interpolated from 1:24,000-Scale Topographic Map

	Latitude	Longitude	Elevation
Sheep Spring	37°08'17" N	113°37'06" W	890 m (2920 ft)
Monitoring Wells 1,2,3	37°08'18" N	113°37'06" W	893 m (2930 ft)

Climate

Average annual precipitation in the St. George area is about 21 cm (8.1 in), of which, 6.1 cm (2.4 in) is snow. Average annual mean temperature is 16.8 °C (62.3 °F) and maximum annual evapotranspiration is approximately 160 cm (63 in) (Ashcroft and others, 1992). Vegetation in the Sheep Spring area is sparse desert vegetation. The area is underlain by thin eolian and alluvial deposits, and exposed bedrock.

Previous Hydrogeologic Investigations

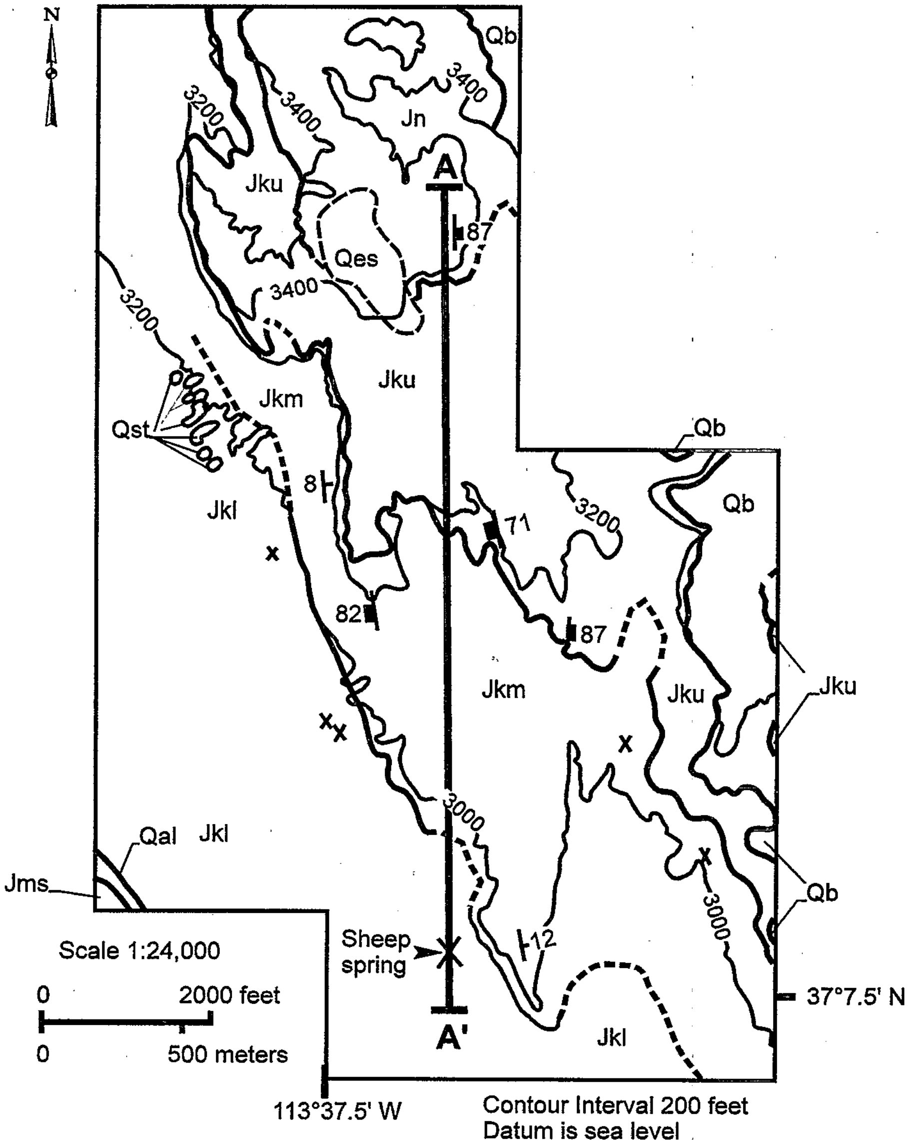
The first ground-water investigation that included the Sheep Spring study area was that of Cordova and others (1972). This regional study presented information on the principal aquifers and on the recharge, movement, discharge, chemical quality, storage, and utilization of ground water. A more recent study by Cordova (1978) presents hydrologic data and a hydrologic evaluation of the Navajo Sandstone, a major aquifer in the St. George area. A preliminary evaluation of the hydrogeology of Sheep Spring was completed by Jensen and Lowe (1992). Clyde (1987) compiled a regional report which covers the St. George area, including the geology, recharge, occurrence, movement, use, quality, and potential development of the ground-water resources. An assessment of the low-temperature geothermal potential of the Santa Clara area, which also includes a geologic map, was completed by Budding and Sommer (1986). Geologic maps that cover the Sheep Spring area include a 1:125,000-scale map by Cook (1960) and a 1:31,250-scale engineering geology map by Christenson and Deen (1983).

Geologic and Hydrologic Setting

Sheep Spring flows from the lower member of the Kayenta Formation, which is interbedded siltstone, sandstone, and shale. The complete Kayenta Formation is approximately 655 m (2150 ft) thick in this area and consists of three members. The Kayenta Formation is underlain by the Moenave Formation and overlain by the Navajo Sandstone. All three formations are Jurassic in age (Hintze, 1988).

The Navajo Sandstone, overlying the Kayenta Formation (Figure 19), is a regional aquifer. The regional ground-water flow direction in the Navajo Sandstone in this area is from north to south, from the Pine Valley Mountains toward discharge at the Santa Clara River and Virgin River (Figure 20). The Pine Valley Mountains are the regional recharge area for much of the St. George and Santa Clara area (Cordova, 1978).

Minor seeps are present along the topographic bench northwest of Sheep Spring, and springs are present 0.8 km (0.5 mi) and farther away, which discharge from a similar stratigraphic interval as Sheep Spring. As the topographic basin in which Sheep Spring is located is large (130 km^2, 50 mi^2), and because other springs discharge from the Kayenta Formation within the basin, Sheep Spring appears to represent only one of many local discharge points for this aquifer system.



Explanation

Qal	younger Holocene alluvium	Jms	Springdale Sandstone Member of Moenave Formation
Qes	eolian sand		
Qb	Quaternary basalt		
Qst	older Holocene spring tufa	X	Spring
Jn	Navajo Sandstone		
Jku	Upper member of Kayenta Formation		
Jkm	Middle member of Kayenta Formation	8	Strike and dip of bedding
Jkl	Lower member of Kayenta Formation	87	Strike and dip of joints

Figure 19. *Geologic map of the Sheep Spring study area.*

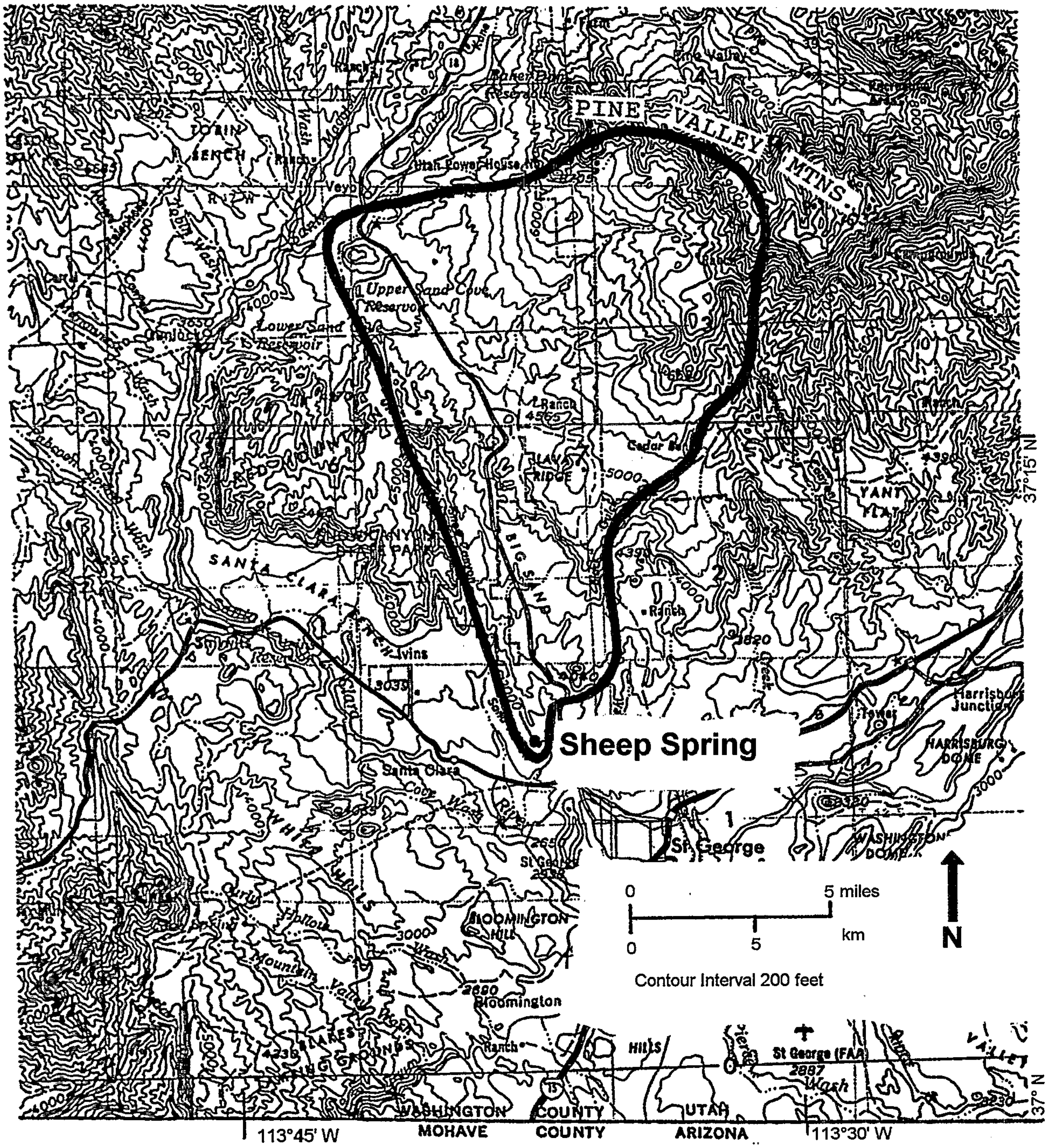


Figure 20. *Potential zone of contribution to Sheep Spring and other springs in this area. (Base map from U.S. Geological Survey 1:250,000 -scale topographic map.)*

Surface Mapping of Hydrogeologic Features

Prior to conducting geologic mapping, topographic maps of the area (1:24,000 scale) were studied to identify the surface drainage divide upgradient of Sheep Spring. The divide, which appears to be the upgradient extent of the potential zone of contribution to Sheep Spring, is approximately 24 km (15 mi) northeast of the spring. Due to the significant distance to this divide, an area of approximately 4.5 km^2 (1.75 mi^2) surrounding the spring was initially designated for detailed field study. Aerial photographs (1:20,000 scale) were also studied to identify fracture traces and lineaments in this area (Figure 21).The photographs displayed a pervasive north-trending grain resulting from the dominant joint orientation (Figure 22). Only one individual feature, approximately 1100 m (3600 ft) in length, was identified within the study area. This feature was located 1160 m (3850 ft) from Sheep Spring.

The Kayenta Formation in this area consists of three members (Figures 19 and 23). Sheep Spring discharges near the top of the lower, slope-forming member. This lower member consists of reddish-brown, interbedded siltstone, sandstone, and shale, and is 90 m to 120 m (300 ft to 400 ft) thick. The middle member is interbedded, moderate-reddish-orange and moderate-reddish-brown, fine- to medium-grained sandstone which forms blocky ledges and slopes. The upper member consists of sandstone and crops out as massive, rounded ledges. The Kayenta Formation dips approximately 8° to 12° east in this area (Figure 19).

The lower member of the Kayenta Formation forms a northwest-trending bench approximately 6 m (20 ft) high. The spring collection tunnel is excavated into this bench. Several minor seeps are scattered along this small bench northwest of Sheep Spring (Figure 19). These small seeps and accompanying vegetation appear to be localized near outcrops of a light-colored sandstone layer. This layer is present in the top of the spring collection tunnel, where it is 0.23 m (0.75 ft) thick, and was also encountered in a detailed measured section approximately 30 m (100 ft) northwest of the collection tunnel, where it is 0.5 m (1.5 ft) thick. In outcrop, this layer consists of moderately well-sorted silty sandstone which is medium-gray with pale bluish-green streaks, has calcareous cement, and appears to have high porosity. This unit, or a similar unit in approximately the same stratigraphic position, can be traced to the northwest of Sheep Spring about 670 m (2200 ft) and southeast about 21 m (70 ft).

Calcareous tufa deposits are associated with outcrops of the sandstone layer along the bench. The tufa is light gray and porous. It is thickest (approximately 1.2 m (4 ft)) in an exposure about 14 m (45 ft) southeast of Sheep Spring and directly overlies the thin sandstone layer. In other outcrops, the tufa is located directly on or up to several feet above the sandstone layer. These tufa deposits, based on porosity and appearance, were apparently deposited at land surface by the evaporation of spring water, then buried by eolian and alluvial sediments.

Older tufa deposits are exposed about 1.9 km (1.2 mi) northwest of Sheep Spring. These deposits are light to very light gray, friable to poorly indurated with blocky weathering, and up to about 1.2 m (4 ft) thick. The tufa forms tabular deposits which cap the tops of sloping hills. The tabular deposits strike N28°W and dip 14° west. The deposits have been eroded, and appear to be remnants of a tufa sheet deposited on the land surface before the slope retreated to its present position. These two levels of tufa deposits indicate that ground water has been discharging from this area for thousands of years.

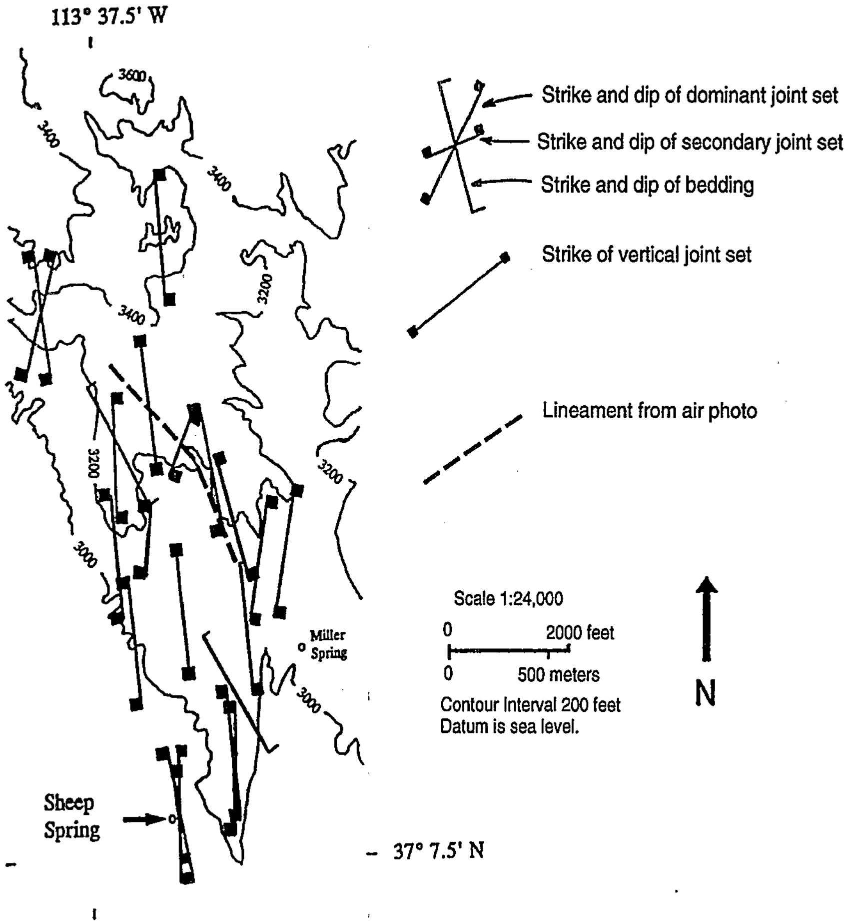


Figure 21. *Joint orientations in the Sheep Spring study area.*

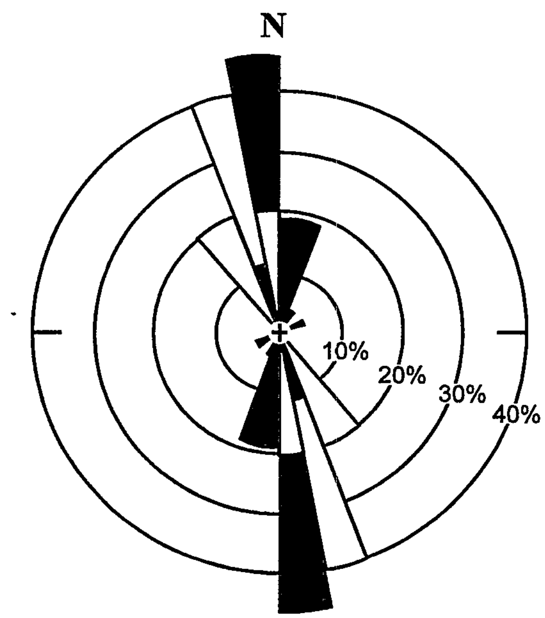


Figure 22. *Rose diagram depicting strike of joints in the Sheep Spring study area. Unshaded sectors represent strike of bedding. Number of joint measurments = 28, number of bedding measurements = 5.*

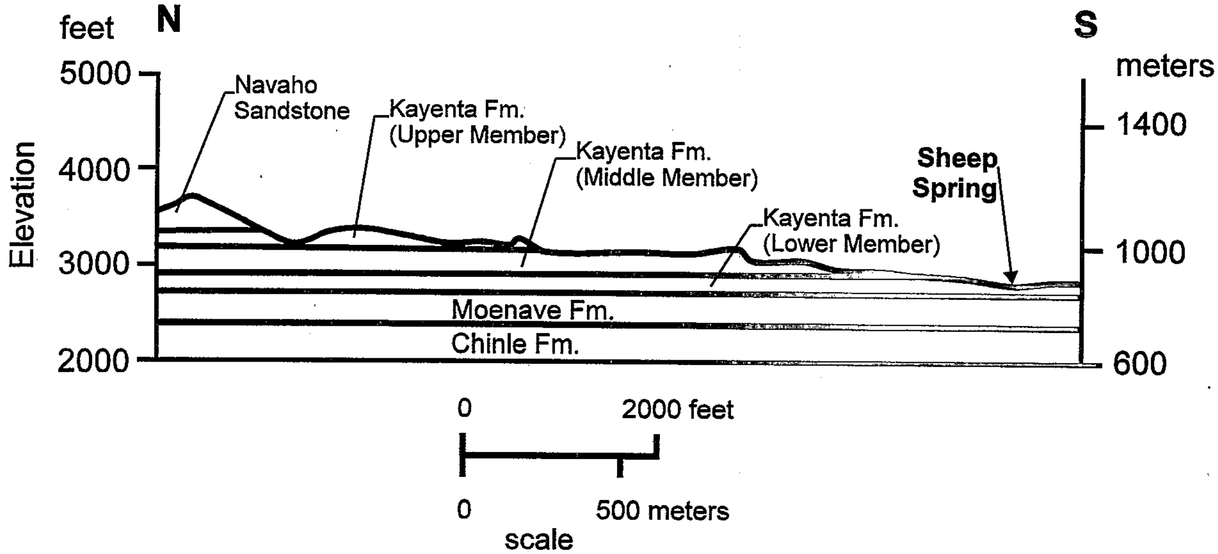


Figure 23. *Generalized geologic cross section through Sheep Spring study area.*

The Kayenta Formation and Navajo Sandstone are highly jointed at Sheep Spring and north of the spring (Figure 21). The dominant joints strike approximately north (Figure 22), and are nearly vertical, dipping steeply to the east and west. The joints are clearly evident in outcrops and in the spring collection tunnel. Spacing of the joints in the collection tunnel ranges from 0.1 m to about 0.9 m (0.3 ft to about 3 ft), with areas of more extensive jointing where the spacing is generally 0.2 m to 0.3 m (0.7 ft to 1 ft). Ground water discharging into the spring collection tunnel appears to flow both from the joints (secondary porosity) and from pore spaces (primary porosity) in the rock matrix on the north wall of the collection tunnel. Apertures of joints in the tunnel are very small, approximately 0.15 cm (0.06 in) and less.

Joint spacing in outcrops in the study area varies from about 0.1 m to about 9 m (several inches to about 30 ft). In outcrop, there are localized areas displaying abundant joints with only 0.1 m to 0.2 m (several inches) of separation between the joints. Joint lengths mapped in outcrops ranged up to the size of the outcrop, approximately 60 m (200 ft). Joint apertures measured in outcrop ranged up to about 1 cm (0.4 in). Thin calcite coatings (0.25 cm (0.1 in)) were observed in some exposed joints. Joints appear to be more abundant in the ledge-forming units, which are usually better indurated sandstone. The jointing extends stratigraphically upward, and is well developed in the overlying Navajo Sandstone.

Based on the geologic mapping, several hypotheses regarding the hydrogeologic system may be formulated:

1. Fracturing in the exposed bedrock may indicate a relatively high degree of vulnerability of the ground water to surface contamination sources.
2. The hydrologic system may be highly anisotropic due to the pervasive north-trending joints. However, as at Olsens Spring, sufficient data are not available to test this hypothesis.
3. Hydraulic conductivity may be relatively low in this system.
4. The potential zone of contribution to springs in this area may be relatively large based on surface drainage divides present in this area.
5. Ground water discharging at Sheep Spring appears to be transmitted mainly, though not entirely,

through a relatively thin sandstone unit in the vicinity of the spring. Flow appears to be through pore spaces and joints.
6. Ground-water flow boundaries could not be inferred from surface mapping of hydrogeologic features within the mapped area.

Borings and Well Installations

Three monitoring wells were installed near Sheep Spring (Figure 24) to provide stratigraphic control and allow performance of aquifer tests. The boreholes were drilled using air rotary technology in interbedded siltstone, sandstone, and shale of the Kayenta Formation. Well 1 was cored from 4.3 m to 7.9 m (14 ft to 26 ft) below land surface and Well 2 from 6.1 m to 7.6 m (20 ft to 25 ft) to provide stratigraphic control across the thin sandstone unit (Figure 25). Wet core and cutting samples were noted in Well 2 between 6.4 m and 7.3 m (21 ft and 24 ft). The material consists of very fine- to fine-grained sandstone and interbedded siltstone. This appears to be the same light-colored layer present in the top of the collection tunnel. A fracture observed in a core sample through the sandstone layer in Well 2 is approximately 0.5 mm (0.02 in) wide, but is mostly filled with silty material. Geophysical logging to aid in determining lithology was not conducted as originally proposed. Coring was chosen over use of borehole geophysics as a more appropriate means of obtaining tight stratigraphic control and collecting samples for laboratory tests of physical properties.

Horizontal hydraulic conductivity was estimated to be 2×10^{-7} cm/s (6×10^{-4} ft/d) using U.S. EPA Method 9100 for laboratory permeameter analyses (U.S. EPA, 1986) by the falling head method. The estimated hydraulic conductivity is quite low but does not account for ground-water movement through fractures. However, it does indicate that significant heterogeneity may exist in the sandstone matrix based on observations of flow through primary porosity in the spring collection tunnel. An effective porosity of 13% was also estimated for a sample from the base of this sandstone unit using Boyle's Law (Brown, 1981).

The wells were constructed using 10.2 cm (4 in) diameter schedule 40 PVC casing and screened across the water table. A graded filter pack was installed within the screened zone, a bentonite seal was placed above the sand pack, and the annulus was grouted to the surface. Each well was developed by alternately pumping and injecting water obtained from the spring. The well was then pumped to remove the injected water following development. The top of each well casing was surveyed with respect to the top of the spring collection tunnel.

The monitoring wells were installed during the final phase of field studies to take advantage of information acquired during previous phases. Thus, limited information on temporal fluctuations in ground-water elevations was obtained. As at the Olsens Spring site, continuous ground-water elevation data may have been useful in understanding site hydrology and should often be obtained, if feasible. However, concerns for the security of monitoring equipment prevented acquisition of such data at this site. Ground-water elevations in Well 2 and Well 3 rose approximately 0.13 m (0.4 ft) between May 1992 and May 1993. The ground-water elevation in Well 1 rose 0.25 m (0.8 ft) during the same period (Figure 26).The similarity in water level fluctuations between Wells 2 and 3 and the difference between these wells and Well 1 may indicate that Well 1 was notat equilibrium at the time of the measurement in May 1992, immediately following well installation. However, it is possible that water levels in any of the wells may not have returned to static conditions at the time of the May 1992 measurements. Therefore, these measurements are considered to be suspect. Reliable hydraulic gradients could not be calculated from relative ground-water elevations due to the close well spacing and the

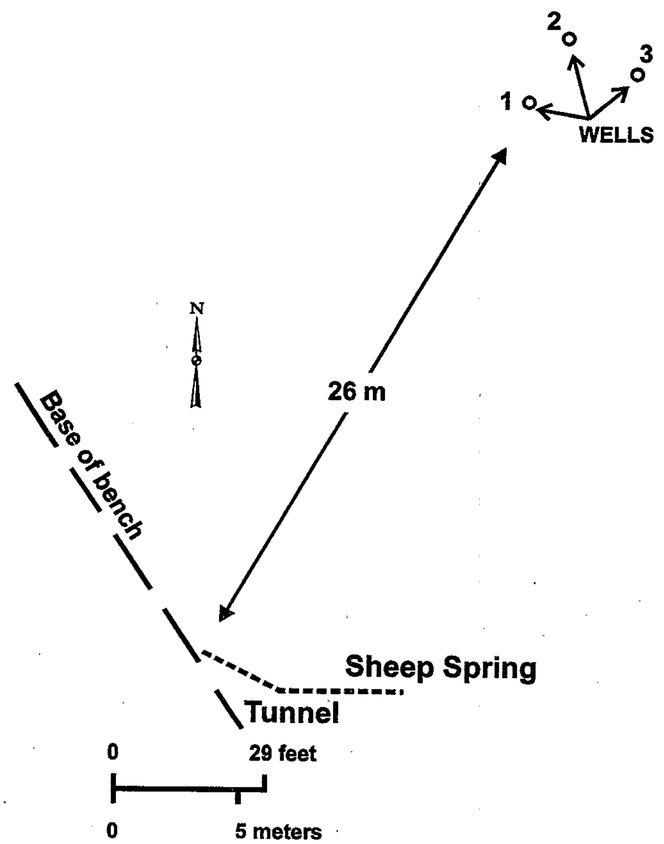


Figure 24. *Location of monitoring wells near Sheep Spring.*

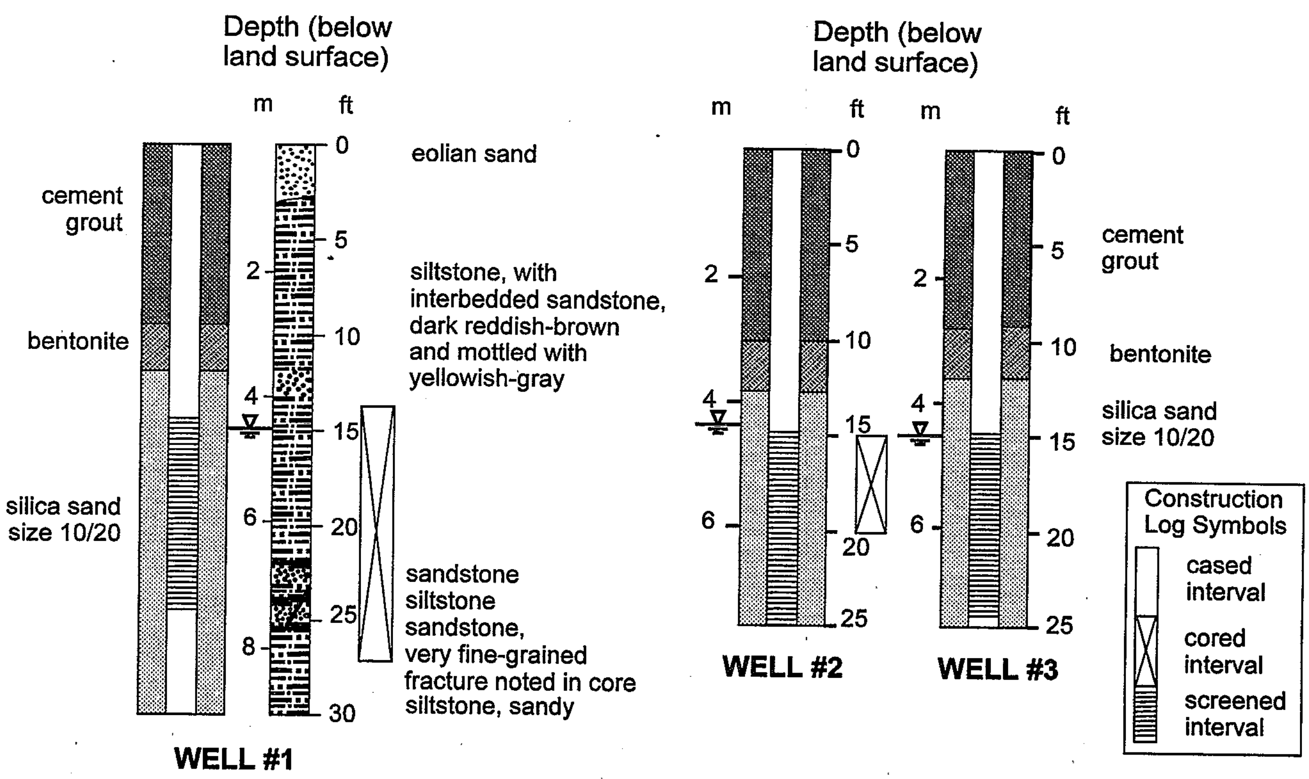


Figure 25. *Construction logs of montoring wells near Sheep Spring and representative geologic log of Well 1.*

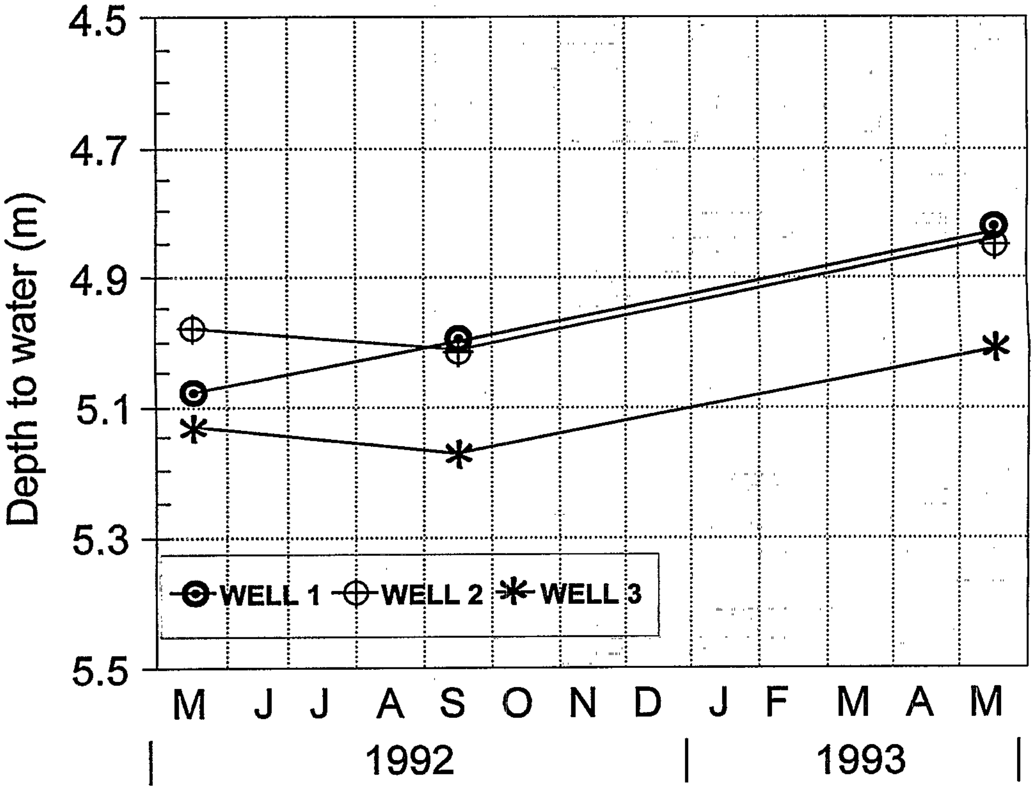


Figure 26. Depth to ground water below the top of casing in monitoring wells at Sheep Spring.

small differences in hydraulic head. The well spacing was chosen to allow performance of a multi-well pumping test for evaluation of transmissivity and anisotropy. Greater well spacing would be required to estimate hydraulic gradient in this area. Ground-water elevations in the spring collection tunnel could not be used in the calculations due to the manner in which the spring was developed. It appeared that the roof of the collection tunnel was below the water table.

Physical and Chemical Properties of Discharge

Field Parameters

Discharge rate, temperature, pH, specific conductance, and turbidity of the water from Sheep Spring were monitored on a monthly basis from November 1990 through October 1991 (Table 14). The discharge rate was relatively constant at approximately 7.6 l/min (2.0 gal/min) throughout the 12-month sampling period. The flow rate was determined at a drain outside the collection tunnel using a 7.6 l (2.0 gal) bucket and stopwatch. The temperature of the water varied through the sample period from a low of 16.8 °C (62.2 °F) in April 1991 to a high of 21.3 °C (70.3 °F) in September and October 1992 (Figure 27). Based on regional studies and the climate in this area of Utah, the relatively high ground-water temperature does not appear to be related to geothermal activity. The observed fluctuation appeared to be related to the mean monthly air temperatures recorded at the weather station in St. George and may indicate seasonal changes in near-surface ground temperatures. Specific conductance fluctuated only about 10% and ranged from 698 µmhos/cm to 770 µmhos/cm (Figure 28). These

Table 14. *Field Parameters for Water Samples from Sheep Spring*

Date	Temperature (°C)	Specific Conductance (µmhos/cm)	Total Dissolved Solids (mg/1)	pH	Turbidity (NTU)	Flow (l/min)	Flow (gpm)
28-Nov-90	19.7	770	371	8.12	--	--	--
20-Dec-90	19.5	727	363	8.10	0.11	7.9	2.1
17-Jan-91	17.9	720	360	8.02	0.13	7.9	2.1
14-Feb-91	17.2	741	370	7.85	0.05	7.9	2.1
19-Mar-91	16.9	698	348	7.83	0.07	7.9	2.1
16-Apr-91	16.8	736	370	7.92	0.04	7.9	2.1
16-May-91	17.3	733	367	7.81	0.05	7.6	2.0
12-Jun-91	18.0	735	368	7.65	0.07	7.9	2.1
9-Jul-91	19.1	730	365	7.76	0.04	7.2	1.9
21-Aug-91	20.7	762	381	7.74	0.08	7.9	2.1
17-Sep-91	21.3	727	364	7.85	0.16	7.2	1.9
29-Oct-91	21.3	729	365	8.08	0.07	7.2	1.9

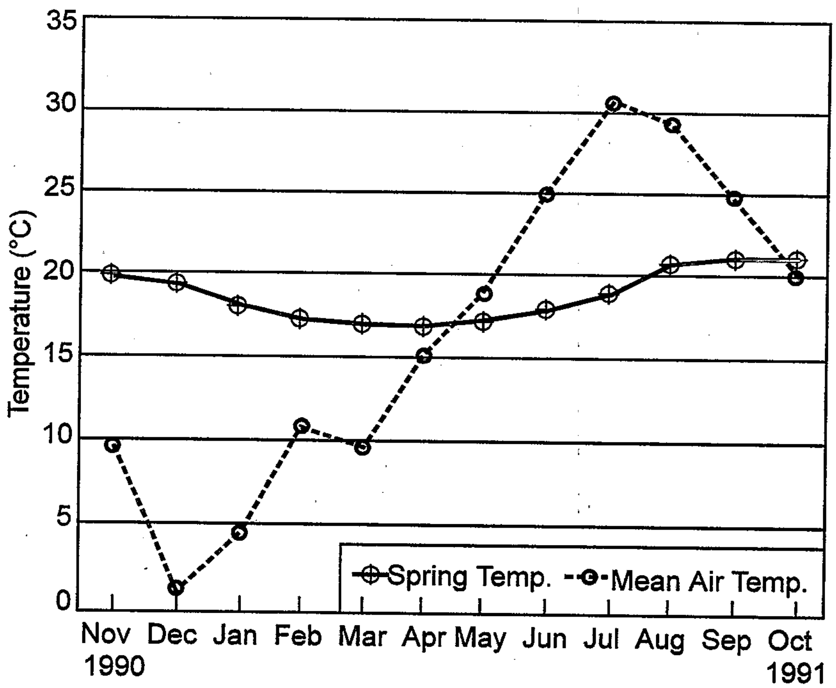


Figure 27. *Temperature of water samples from Sheep Spring and mean air temperature at St. George.*

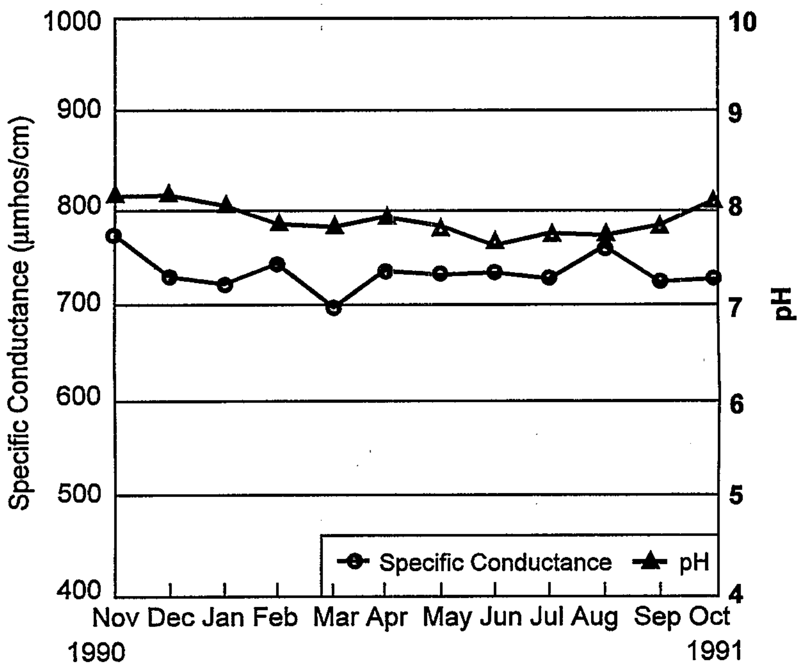


Figure 28. Specific conductance and pH of water samples from Sheep Spring.

minor fluctuations do not appear to correlate with changes in temperature. The pH ranged from 7.65 to 8.12 with only minor seasonal trends potentially present. The potential trends were not considered to be significant. Turbidity was low and ranged from 0.04 NTU to 0.16 NTU.

Geochemical Characterization

Water samples from Sheep Spring were collected quarterly for one year and analyzed by standard laboratory techniques (Table 9) for major ions, selected trace elements, radionuclides, and total dissolved solids (Table 15). Analyses of ground-water samples from the wells were not performed at this site as at the Olsens Spring study site. The decision not to obtain samples from these wells was based on a high degree of confidence in discharge location (i.e., Sheep Spring) for ground water at the location of the wells. Water samples from Sheep Spring provide results that are more representative of average conditions within the zone of contribution. Major ion concentrations from these analyses were used to construct a trilinear diagram as developed by Piper (1944) (Figure 29). Based on the hydrochemical facies of Back (1966), the trilinear diagram indicates calcium-sulfate type ground water. The ion balances for the four analyses had 0% to 4% error. The cations plot in the field between calcium and sodium water types. Sulfate and bicarbonate are the predominant anions in all of the samples. The relatively high concentrations of dissolved sodium, calcium, and magnesium may be due to the fine-grained sediments in the Kayenta Formation.

The high sulfate content may indicate the presence of gypsum in the Kayenta Formation. The nearest lithologic unit that contains large amounts of gypsum is the Carmel Formation, which overlies the Navajo Sandstone. However, the nearest outcrop is approximately 11 km (7 mi) to the north of Sheep Spring (Cordova and others, 1972). High sulfate content may also be due to sulfate cement in the Kayenta Formation. Sulfate cement has been encountered locally in the Navajo Sandstone in the St. George area (R. Rasely, U.S. Soil

Table 15. *Analytical Results of Water Samples from Sheep Spring*

Sample Date	20-Dec-90	19-Mar-91	12-Jun-91	17-Sep-91
Calcium (mg/l)	58	59	56	54
Magnesium (mg/l)	24	24	23	23
Potassium (mg/l)	6	6	5.8	6.4
Sodium (mg/l)	63	64	62	62
Bicarbonate(mg/l)	174	177	166	172
Carbonate (mg/l)	0	0	0	0
Chloride (mg/l)	27	28.5	28	27.5
Silica (mg/l)	18	18	16.8	19
Sulfate (mg/l)	200	190	190	172
Nitrate+Nitrite(mg/l)	0.72	0.08	0.77	0.79
Phosphate (mg/l)	<0.01	<0.01	<0.01	<0.01
Iron (mg/l)	<0.02	<0.02	<0.02	<0.02
Alkalinity (mg/l)	143	145	136	141
Specific Conductance(μmhos/cm)	727	737	715	731
Total Dissolved Solids (mg/l)	520	498	480	464
Gross Alpha (pc/l)	3	<1	3	<1
Gross Beta (pc/l)	12.2	<5	11	<5
Cations (meq/l)	7.8	7.9	7.5	7.5
Anions (meq/l)	7.8	7.7	7.5	7.2

Conservation Service, personal communication to M. Jensen, 1992). Significant seasonal variation in hydrochemical facies was not observed. Thus, no information regarding potential seasonal fluctuation in ground-water flow path can be inferred.

Two water samples from Sheep Spring were collected for tritium analyses, one each in June and October 1991. Tritium analyses yielded activities of 2.4 (±2.8) TU and 1.9 (±2.6) TU. These low activities indicate that recharge to the aquifer probably took place before 1953 (Fetter, 1988). These results indicate that the average residence time of water discharging at Sheep Spring may be relatively long. No additional information regarding seasonal fluctuations in ground-water flow paths or residence time could be reliably inferred.

Catchment Area Calculations

Average annual precipitation at the St. George weather station is about 21 cm (8.1 in) (Ashcroft and others, 1992). There are no data from the vicinity of Sheep Spring that may be used to reliably estimate evapotranspiration or recharge. For the purposes of this investigation it is estimated that recharge is equal to 10% of average annual precipitation or about 2.0 cm (0.80 in). However, that value may be an overestimate of recharge in this arid area. The discharge of Sheep Spring is about 7.6 l/min (2 gal/min). Using the method of Todd (1980), a plot of catchment area as a function of estimated recharge and discharge rates for Sheep Spring (Figure 30) indicates that a catchment area of about 0.2 km^2 (0.08 mi^2) would be necessary to provide enough water for the spring to yield 7.6 l/min (2 gal/min). Use of a recharge estimate of 0.2 cm (0.008 in) results in a calculated catchment area of approximately 2 km^2 (0.8 mi^2). Minor, scattered seeps are present along the topographic bench north of Sheep Spring. However, no data are available to estimate the discharge of these seeps. If the seeps were included in the calculations, the catchment area would be larger. The topographic basin in which

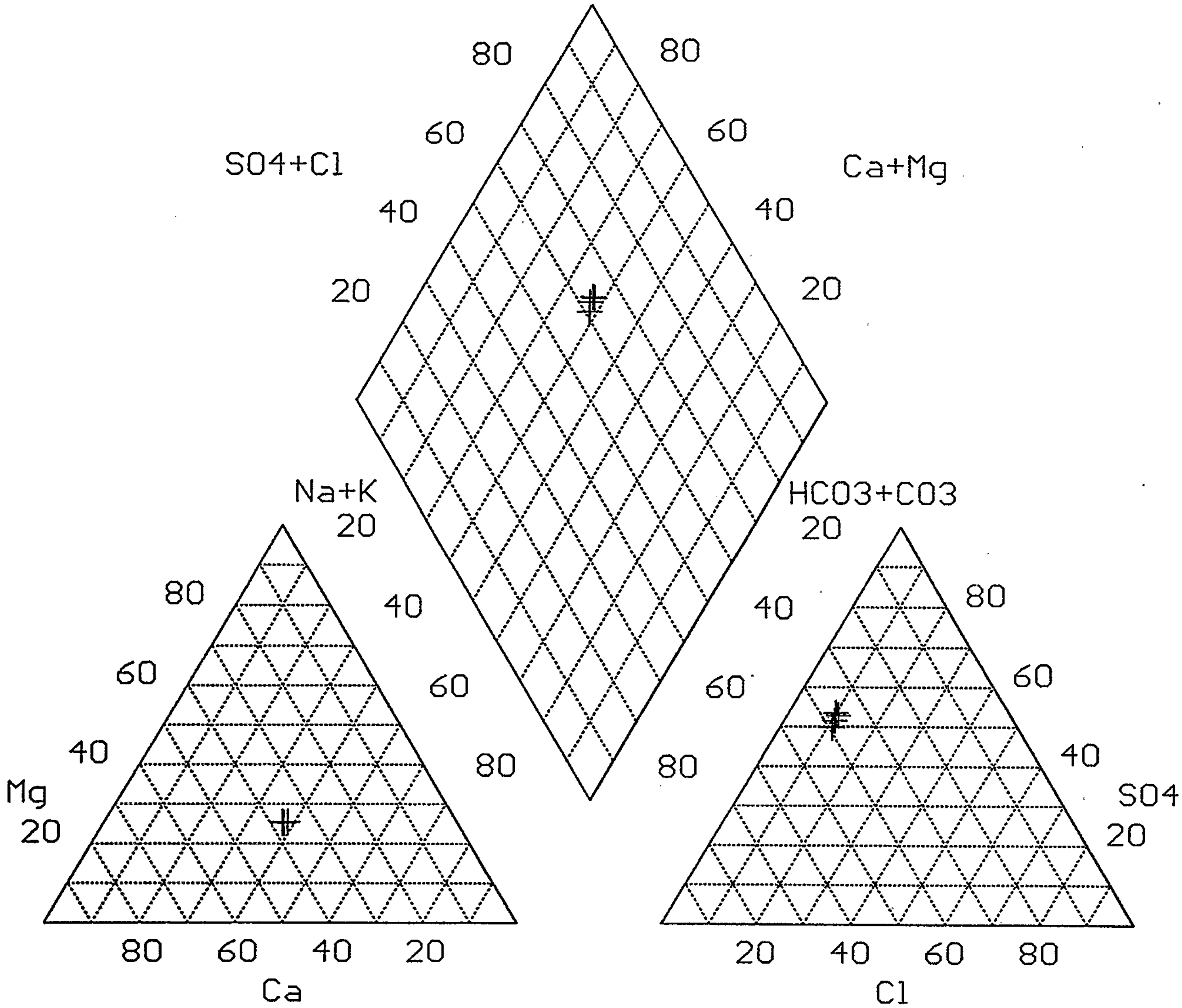


Figure 29. *Trilinear diagram depicting geochemical analyses of water samples from Sheep Spring.*

Sheep Spring is located is significantly larger (130 km^2, 50 mi^2) (Figure 20) than the estimated catchment area. Although the estimated catchment area required to support the discharge is relatively small, little quantitative information concerning the zone of contribution and its location can be inferred from this calculation.

Hydrogeologic Conceptual Model

Sheep Spring appears to be a contact or seep-type spring (Figure 1). The conceptual model, developed from hydrogeologic mapping, is that of a spring discharging from a fractured clastic-rock aquifer of relatively low hydraulic conductivity with ground water moving through primary and secondary porosity. A layer of very fine-grained sandstone within the lower member of the Kayenta Formation appears to be transmitting much of the water in the vicinity of the spring. The presence of spring tufa deposits indicates that ground water has been discharging from the area of Sheep Spring for thousands of years.

Regional ground-water flow is from north to south. However, local flow directions may be affected by the location and discharge rates of other springs in this basin. The system may be anisotropic and may not behave

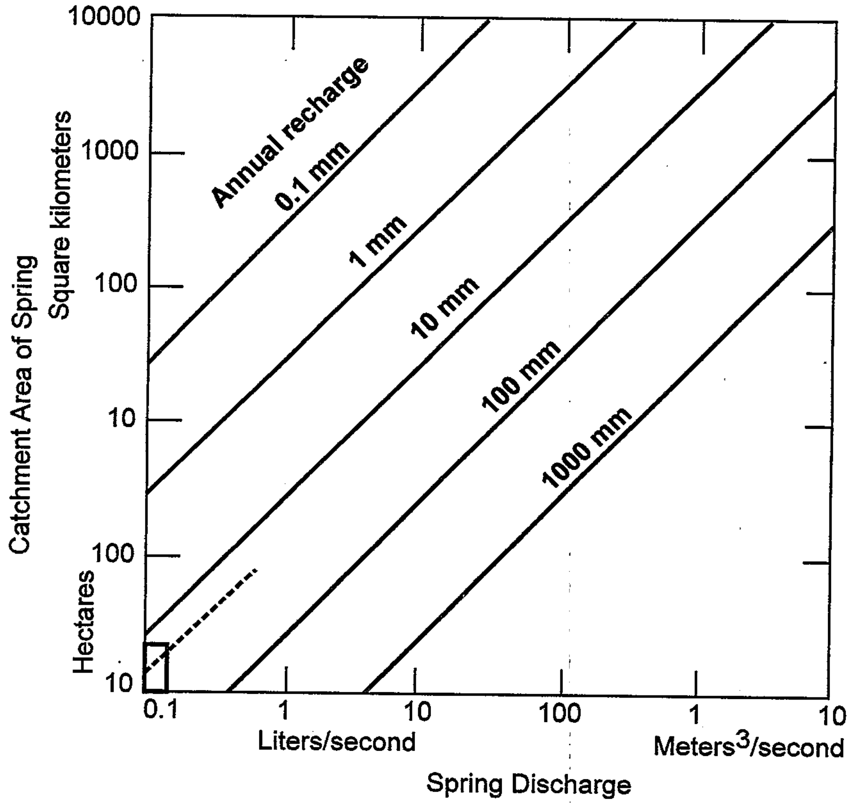


Figure 30. Estimation of catchment area for Sheep Spring. (Modified from Groundwater Hydrology, D. K. Todd, copyright © 1980, John Wiley & Sons, Inc., reprinted by permission of John Wiley & Sons, Inc.)

as a porous medium on the scale of a potential protection zone because essentially all of the fractures strike in the same direction (Figure 21). Sheep Spring appears to be a local discharge point with, potentially, a relatively large zone of contribution based on the locations of topographic divides. Ground-water residence times in this setting may be relatively long. Based on average ground-water age, significant recharge may not be occurring in the vicinity of Sheep Spring. No ground-water flow boundaries near the spring were defined. There are no perennial streams or other major hydrologic features in the potential zone of contribution that may act as hydrologic boundaries.

Aquifer Tests

A short-term constant rate pumping test was attempted in Well 2. However, the sustainable discharge rate was very low, approximately 0.7 l/min (0.18 gal/min). No significant drawdown was detected in either observation well, located 3.0 m (10 ft) and 3.2 m (10.4 ft) from the pumping well during the three-hour test. The data do not appear to adequately fit potential solutions for homogeneous porous media. It appears from a qualitative analysis that the test may have been complicated by subsurface heterogeneity, low hydraulic conductivity, and well storage effects.

Rising-head slug tests were performed in each of the three monitoring wells at Sheep Spring to estimate hydraulic conductivity for comparison with qualitative assessments based on hydrogeologic mapping. The tests were conducted by inserting a submersible pump in the well and allowing the water level to return to its static elevation. Water was then rapidly pumped to lower the water level in the well about 0.9 m to 1.2 m (3 ft to 4 ft).

This took only 5 seconds to 10 seconds and is considered to be approximately instantaneous under the conditions of this test. The rising water level was then monitored using an electric water level indicator until it was within 95% of the original static level.

Data from these tests (Figure 31) were analyzed using the method of Bouwer and Rice (1976), as modified by Bouwer (1989). Data obtained during the first 30 seconds of recovery indicate a much higher recovery rate than subsequent data. This appears to be due to backflow of water from the discharge hose recharging the well as a functioning check valve was not in place in the discharge line during the tests. Drainage from the sand pack, which has a relatively high hydraulic conductivity in comparison to the geologic formation, may also have contributed to this effect. However, the dominant cause appears to be drainage from the discharge line based on comparison of the water volume in the line and the anomalous head rise. Therefore, data collected during the first 30 seconds of each test were not used in the interpretations. An aquifer thickness of 3.1 m (10 ft), which is approximately the saturated thickness of the screened interval, was assumed. However, results of analyses assuming an aquifer thickness of 15.2 m (50 ft) and 30.5 m (100 ft) were not significantly different from those using an assumed thickness of 3.1 m (10 ft). Estimated hydraulic conductivities from the tests (Table 16) ranged from approximately 0.04 m/d to 0.2 m/d (0.1 ft/d to 0.6 ft/d). These results support the conceptual model of an aquifer with a relatively low hydraulic conductivity.

A similar value of 0.3 m/d (1 ft/d) for hydraulic conductivity was estimated by Cordova and others (1972) based on the specific capacity of a well located about 4 km (2.5 mi) from Sheep Spring and aquifer characteristics of the Kayenta Formation. Studies have not been sufficient to adequately define subsurface heterogeneity and the potential range of hydraulic parameters. For example, wells used for hydraulic testing may not intersect transmissive joints and results may not be representative of the most conductive areas or average conditions at the site. More extensive studies would be required to better define hydraulic parameters within the potential zone of contribution to Sheep Spring. However, it is questionable as to whether such studies would provide significantly more reliable estimates in a cost-effective manner. A tracer test was not conducted at this site as it was at Olsens Spring due to the increased confidence in the discharge point for ground water in the vicinity of the wells and the difficulties involved in conducting representative tests in materials of low hydraulic conductivity.

Zone of Contribution and Delineation of Spring Protection Zones

Hydrogeologic mapping indicated that the potential zone of contribution to springs in this area may extend as much as approximately 24 km (15 mi) north of Sheep Spring. This would be considered to be the maximum plausible extent. The actual zone of contribution may be much smaller. Many factors, including ground-water discharge from nearby springs, control local flow directions. Ground-water flow boundaries closer to Sheep Spring could not be inferred from hydrogeologic mapping methods applied in this study. Conservative protection zones that cover the entire surface drainage area (i.e., more than 130 km^2) may be considered too large to be effectively managed by the water supplier. However, a relatively low ground-water seepage velocity may potentially limit the zone of contribution within the 15-year time period to a much smaller area. Therefore, incorporation of additional delineation techniques (e.g., estimates of ground-water time-of-travel) may be required to define more practicable protection zones.

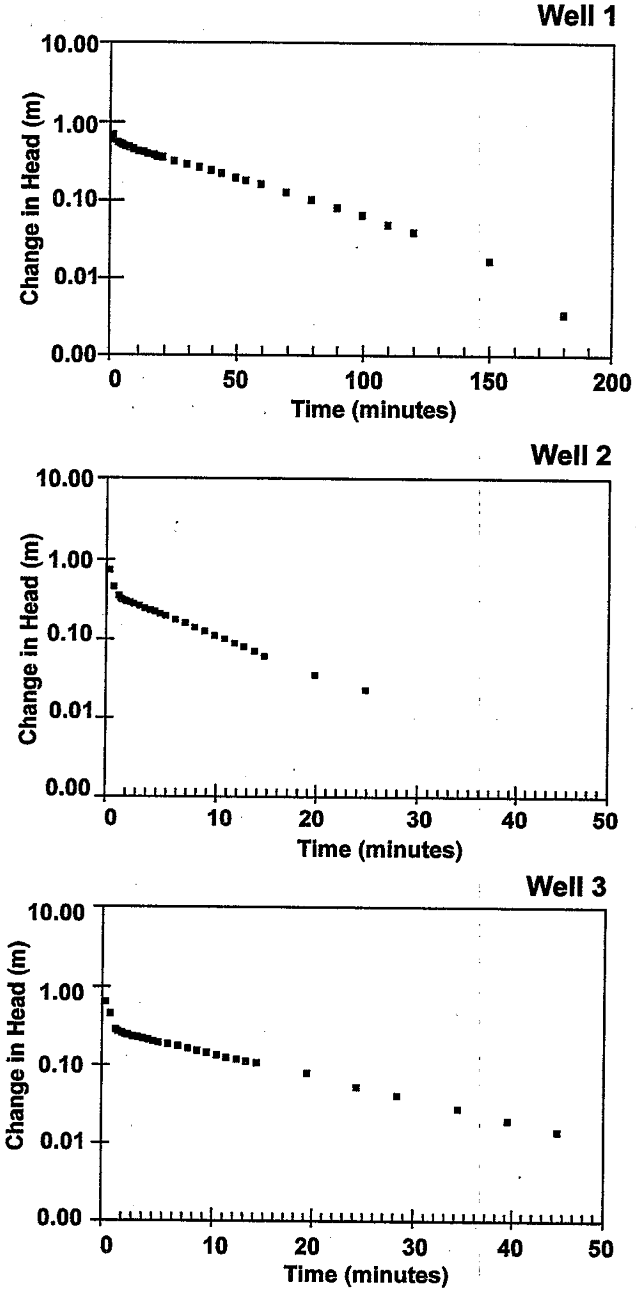


Figure 31. *Recovery data from slug tests in wells near Sheep Spring.*

Table 16. *Hydraulic Conductivity Estimated from Slug Tests in Wells at Sheep Spring*

Well	*Hydraulic Conductivity (m/d)*	*(ft/d)*
1	*0.04*	*0.1*
2	*0.2*	*0.6*
3	*0.1*	*0.4*

The information regarding hydraulic parameters obtained during this study would generally not be sufficient to define reliable protection zones based solely on time-of-travel criteria. However, ground-water seepage velocity was estimated from these data to illustrate the potential use of such criteria in refining the protection zone delineation. For purposes of these preliminary calculations, a hydraulic gradient of 0.05 was assumed based on land surface topography. The implicit assumptions are that the aquifer is unconfined and the water table mimics topography. This is the approximate slope of the relatively flat area directly north of the spring, and also the slope from a surface drainage north of the spring to the spring collection area. This gradient appears to be relatively high and significant uncertainty exists in applying this estimate throughout the potential protection zone over the required time periods.

Minor seeps and discharge into the collection tunnel indicate that much of the ground water at the discharge point is moving through a sandstone layer. Effective porosity estimated from a core sample of this unit was 13%. Using these parameters and a hydraulic conductivity of 0.3 m/d (1 ft/d), the estimated average interstitial velocity is 0.12 m/d (0.4 ft/d). This velocity yields a 250-day travel distance of 30 m (100 ft) and a 15-year travel distance of 660 m (2200 ft) topographically upgradient of the spring (Figure 32). These potential protection zones cover only a part of the potential zone of contribution and would be a more manageable and, perhaps, more reasonable protection zone for the water supplier. As previously noted, significantly more hydrogeologic data from subsurface investigations would be required to estimate more reliable protection zones using such criteria.

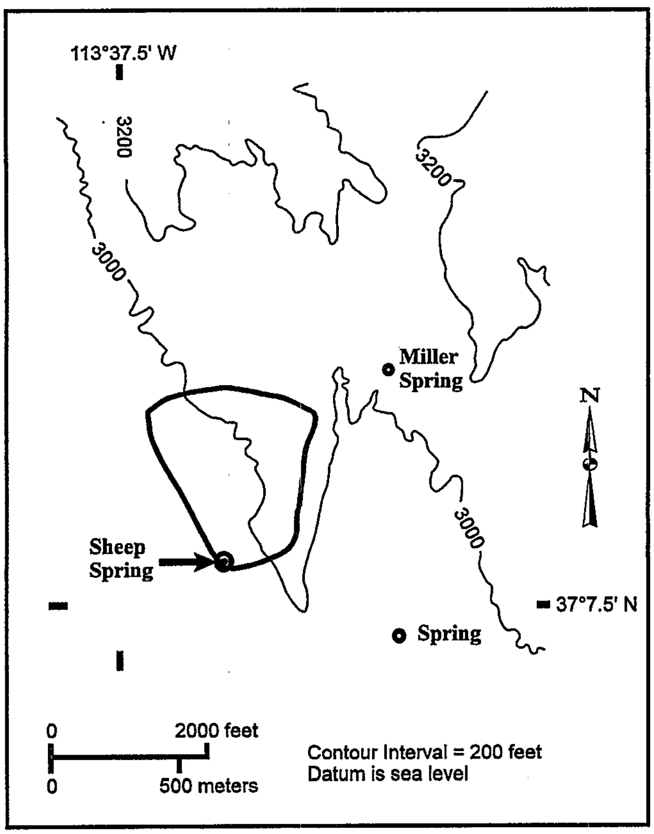


Figure 32. Example of potential protection zones at Sheep Spring based on preliminary time-of-travel calculations. Small zone around spring is 30 m (100 ft) travel distance and large area is 660 m (2200 ft) travel distance.

Chapter 4
CONCLUSIONS AND RECOMMENDATIONS

EVALUATION OF HYDROGEOLOGIC MAPPING METHODS

Results from these case studies support the conclusion that hydrogeologic mapping methods may be used at many sites to provide basic site characterization information. Case studies performed in significantly different settings would be required to evaluate the relative utility of different mapping methods in each setting. This investigation took place in the arid Southwest where vegetation is sparse and population density is relatively low. One of the more powerful hydrogeologic characterization techniques (potentiometric surface mapping) could not be used in this study. This method of ground-water flow field analysis may be very effective in other areas where population density is higher and more ground-water elevation data are available. Other methods used in this study, such as detailed geologic mapping, may not be effective in many areas due to dense vegetative cover.

Potential ground-water flow controls and boundaries may be located using these techniques. These controls may be geologic (e.g., areal limits of aquifer formations) or hydrologic (e.g., ground-water divides). At some sites, hydrogeologic mapping may result in delineation of ground-water flow boundaries suitable for use in directly defining protection zones around springs. Although these methods will not be capable of fully defining the zone of contribution to springs at all sites, results of such investigations will provide information needed for more informed decisions regarding potential zones of contribution and appropriate protection zones at most sites. As in this study, such mapping may also serve as a basis from which additional investigations to better define the zone of contribution, hydraulic parameters, and subsurface heterogeneity may be assessed.

In these case studies, hydrogeologic mapping provided information on aquifer characteristics including aquifer lithologies, relative ground-water residence times, qualitative evaluation of hydraulic parameters, potential recharge areas, and potential physical/hydrologic controls on ground-water flow. This information was valuable in building the conceptual models for site hydrogeology, estimating potential zones of contribution, and constraining potential protection zones around each spring. However, ground-water flow boundaries suitable for delineating practicable protection zones with a high degree of confidence could not be identified at either site using the techniques applied in this study. Additional subsurface studies may be warranted at each site to support delineation of more reliable and more practicable protection zones.

Application of these techniques in the carbonate aquifer (Olsens Spring study site) provided sufficient information to indicate that the portion of the aquifer supplying water to Olsens Spring may behave predominantly as a diffuse-flow system as termed by Shuster and White (1971). In other words, flow may be predominantly laminar along joints, fractures, bedding planes, and other features interconnected at a scale much smaller than the site scale and, potentially, on the scale of centimeters to meters. However, additional studies would be required to better define potential conduit-flow components.

Geologic mapping indicated the area surrounding the spring may be relatively vulnerable to ground-water contamination from surface sources. Ground-water velocity near the springs, estimated from a tracer study and hydraulic parameters, was relatively high. Results also indicated that the zone of contribution to Olsen's Spring and West Hallings Spring may be larger than the topographic basin surrounding the springs. This implies that reliable protection zones may also extend beyond the basin. Available information indicated that potential protection zones should, at a minimum, encompass the entire surface drainage basin.

Additional studies designed to obtain piezometric data within and surrounding the drainage basin may have provided valuable information for evaluation of the shallow flow system, but were not feasible in this investigation due to a lack of available wells. Evaluation of piezometric information may be more feasible in other areas where private wells are installed into the aquifer under investigation. Information regarding the role of faults in ground-water movement, potential presence of shallow ground-water divides, and improved fracture characterization may be obtained from such studies. Additional tracer and more extensive aquifer tests could be conducted within the potential zone of contribution to better estimate ground-water velocity and anisotropy. Use of naturally occurring isotopes may also be valuable in tracing ground water from recharge to discharge following installation of additional piezometers. Additional monitoring of mineral saturation indices and frequent monitoring of discharge rates of individual springs could be instituted to confirm temporal trends observed in this study. Other evaluation tools include analysis of chlorofluorocarbon compounds and tritium/helium-3 dating techniques to better estimate ground-water residence time. Detailed monitoring of spring parameters such as discharge rate, water temperature, turbidity, or specific conductance before, during, and after recharge events may also provide a better indication of whether a significant component of conduit flow exists.

In the fractured clastic-rock aquifer (Sheep Spring study site), the potential zone of contribution to springs in this area may be bounded by a ground-water divide located about 24 km (15 mi) north of Sheep Spring. However, flow boundaries closer to the spring could not be inferred using the methods applied in this study. Due to the potentially long ground-water residence time and the relatively large extent of the aquifer that was indicated, protection zones delineated using other techniques, such as estimation of ground-water time-of-travel, may be more practicable.

Additional studies designed to provide more detailed information regarding hydraulic parameters and potential subsurface heterogeneity would be required to more reliably estimate ground-water time-of-travel to Sheep Spring. Such studies could include aquifer testing to define anisotropy, potentiometric surface mapping, and use of tools such as the borehole flowmeter to define the relative flow contribution from various lithologic units and from fractures within those units. A more quantitative determination of ground-water residence time, potentially comparing information from radiocarbon, chlorofluorocarbon, and tritium/helium-3 analyses with estimates derived from hydraulic information, may also be useful in evaluating potential protection zones.

GUIDANCE FOR APPLICATION OF TECHNIQUES

The hydrogeologic mapping methods applied at any spring site will depend on such factors as hydrogeologic setting, availability of existing hydrogeologic information, and management objectives to be applied within the delineated protection zones. Integration of multiple mapping techniques will be required to develop the hydrogeologic conceptual model and define potential ground-water flow controls and boundaries to support the delineation. The success of all of these methods depends greatly on the hydrogeologic complexity of the spring site.

A general methodology for aquifer characterization using hydrogeologic mapping begins with a review of site-specific hydrogeologic literature to develop an initial conceptual model for site hydrogeology. The conceptual model is the synthesis of available information designed to convey the assumptions and hypotheses regarding ground-water flow to the spring. Development of a conceptual model is an essential element in the delineation process. The model serves as the focus for characterization efforts.

Important assumptions and potential ground-water flow controls should be identified from this model. Hydrogeologic mapping and subsurface investigations may then be designed and implemented to test these assumptions and delineate specific physical and hydrologic controls. Results from these studies may be used to refine the conceptual model, estimate the zone of contribution, and, potentially, delineate protection zones based on controls. Important information gaps may be identified and additional investigations proposed to fill these gaps.

Review of previous investigations, analyses of topographic maps, fracture trace analyses from aerial photographs, estimation of catchment area, evaluation of geochemical parameters, and construction of potentiometric maps should provide valuable information for model development at most sites. Detailed geologic mapping, supported by surface and borehole geophysical surveys, may be useful at many sites in defining the boundaries of formations acting as aquifers and aquitards and locating features that may serve as pathways or boundaries for ground-water flow (e.g., fracture systems and faults). Isotope studies may be used to constrain ground-water age, residence time, and location of recharge. Specialized tools, such as tracer studies, may provide direct information regarding ground-water velocity and zone of contribution, particularly in karst settings.

Hydrogeologic mapping techniques are relatively low cost characterization tools. Many of these techniques are also simple to apply. The applicability of particular tools at spring sites depends on site conditions and the nature of potential ground-water flow controls. Many of these methods may be most applicable at sites where little information concerning the hydrogeologic setting is available. It is recommended that the potential utility of such techniques be evaluated early in all spring protection zone delineation projects.

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