STATE OF MONTANA
BUREAU OF MINES AND GEOLOGY
E. G. Koch, Director

BULLETIN 50-A

GEOLOGY AND GROUND-WATER RESOURCES OF WESTERN AND SOUTHERN PARTS OF JUDITH BASIN, MONTANA

By
Everett A. Zimmerman

This bulletin has been prepared by the United States Geological Survey, Water Resources Division, under a cooperative agreement with the Montana Bureau of Mines and Geology

MONTANA COLLEGE OF MINERAL SCIENCE AND TECHNOLOGY
Butte, Montana
1966

For sale by
Montana Bureau of Mines and Geology
Room 203-B, Main Hall
Montana College of Mineral Science and Technology
Butte, Montana

Price $1.25

(Part B is free)

CONTENTS

ILLUSTRATIONS

TABLES

See also Part B for additional tables.

GEOLOGY AND GROUND-WATER RESOURCES OF WESTERN AND SOUTHERN PARTS OF JUDITH BASIN, MONTANA

by
Everett A. Zimmerman

ABSTRACT

The Judith Basin, a structural and topographic basin in central Montana, is partly surrounded by isolated mountain ranges. The Little Belt and Big Snowy Mountains bound the western and southern parts of the Judith Basin on the southwest and southeast respectively. The Highwood Mountains bound the area on the northwest. Agriculture is the principal occupation in the Judith Basin. The climate is semiarid; thus, shortages of water for livestock and irrigation are common.

Ground-water recharge in the Judith Basin is derived mainly from precipitation. Many of the water-bearing beds underlying the basin are exposed in the mountains or foothills where precipitation is greater than that in the lower parts of the basin. Some aquifers, notably the aquifer in the Madison Group, are recharged in their outcrop area by several hundred cubic feet per second of water from streams. Infiltration of irrigation water recharges terrace deposits and alluvium in some parts of the area.

Ground water is discharged by wells, springs, effluent streams, evaporation, and transpiration. The discharge by individual wells and springs ranges from less than 1 to about 1,000 gallons per minute, but most yield less than 10 gallons per minute. The towns of Stanford, Geyser, Moore, and Judith Gap obtain their municipal water supplies from ground water. Ground water is the source of domestic water for nearly all residents. One well in the area reportedly flows 1,000 gpm and the owner proposes to use the water for irrigation. Many wells of lesser yield are used for lawn and garden irrigation. Aquifer tests indicate that wells yielding adequate water for irrigation of moderate-size tracts could be constructed in some alluvial deposits.

Rocks ranging in age from Mississippian to Quaternary underlie the study area. Ground water in usable quantities is obtainable from the Madison Group and the Kibbey Sandstone of Mississippian age, the Amsden Formation of Mississippian and Pennsylvanian age, the Swift Sandstone of Jurassic age, the Kootenai Formation, several sandstone beds in the Colorado Shale, the Eagle Sandstone, and the Judith River Formation, all of Cretaceous age, and terrace deposits and alluvium of Quaternary age. Wells finished in most of these formations yield water adequate for domestic and stock needs only, but the amount of recharge observed entering the Madison suggests that it is potentially a source of large quantities of water.

Ground water in the study area is generally hard but of suitable quality for domestic and livestock use. Samples of water from the Madison Group, the Amsden Formation, the Kootenai Formation, and alluvium have a generally low alkalinity hazard but a high salinity hazard for irrigation use.

INTRODUCTION

A cooperative program of ground-water investigations by the Montana Bureau of Mines and Geology and the Ground Water Branch of the U. S. Geological Survey was begun in 1955. The objectives of this program are to appraise the ground-water resources of selected areas of Montana and through publication, to make this information available to the people of the state. Under this program, several investigations have been completed and others are in progress. The status of each investigation under the cooperative program is shown in Figure 1.

An investigation of the southern part of the Judith Basin was begun in 1959. A preliminary report on this area (Zimmerman, 1962) has been published. The report, owing to its preliminary nature, did not include much of the basic information that was available on the southern part of the Judith Basin. In 1962, the area of investigation was extended into the western part of the Judith Basin. This report combines the results of the investigations of the western and southern parts of the Judith Basin and includes all available data on the ground-water resources of the combined areas.

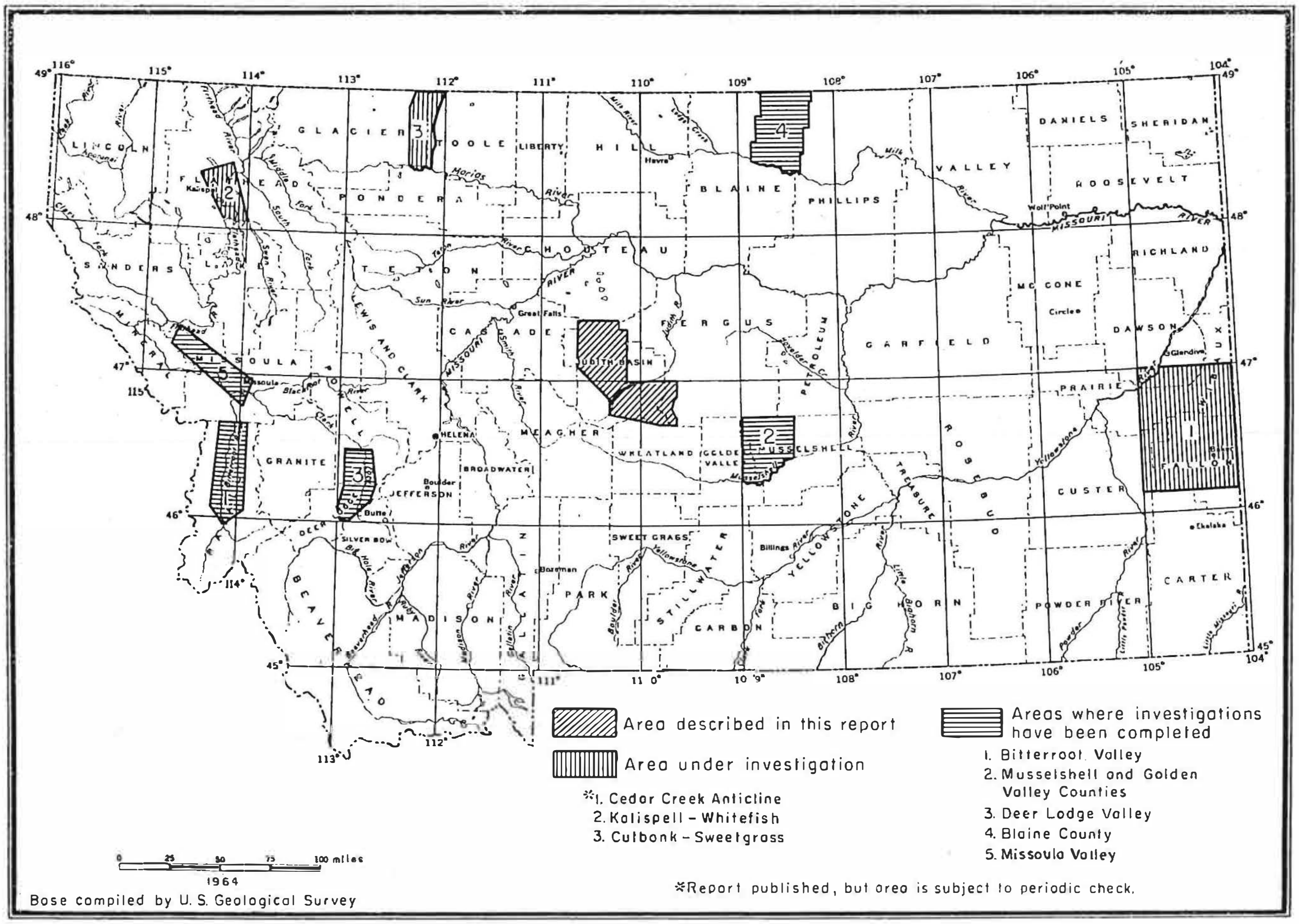

Figure 1.—Map of Montana showing location of cooperative ground-water investigations, 1955-64.

PURPOSE AND SCOPE

The climate of the Judith Basin is semiarid, and agriculture is the major occupation. The stabilizing effect of irrigation on the agricultural economy has been recognized for many years, and the meager surface-water supplies in the area are mostly appropriated. The desire to increase the irrigated acreage and to attract industry, and the need for increased municipal water supplies in the area, has prompted much interest in the development of ground water for these uses. The investigation for this report was undertaken to appraise the ground-water resources of a part of the Judith Basin. The western and southern parts of the basin were selected for study because of the interest of local residents in developing ground water for irrigation and because the potential for such development seemed favorable. An investigation of the ground-water resources of the rest of the basin at a later date is planned.

The factors investigated in the western and southern parts of the Judith Basin included (1) the character, thickness, and distribution of water-bearing rocks; (2) the source, occurrence, and direction of movement of the ground water; (3) the quantity of ground water available; and (4) the chemical quality of the available ground water.

LOCATION AND EXTENT

The combined western and southern parts of the Judith Basin form an irregular area of about 1,100 square miles that is bounded on the south by the Wheatland County line and on the southwest by the outcrop of the Madison Group on the flank of the Little Belt Mountains. The drainage divide between Otter and Arrow Creeks forms the western boundary of the study area. The foot of the Highwood Mountains and a part of the Judith Basin, Chouteau, and Fergus County lines form the northern boundary. On the east, the area is bounded by the east side of Range 12 East, an east-west reach of U. S. Highway 87, and the east side of Range 16 East. The location of the study area is shown on Figure 1.

SITE-NUMBERING SYSTEM

The wells, springs, and other sites noted in this report are numbered according to their location within the system of land subdivision used by the U. S. Bureau of Land Management. The first numeral of the site number denotes the township, the second the range, and the third the section in which the well is located. The lowercase letters a, b, c, and d after the section number show the location of the site within the section; the first letter indicates the quarter section and the second the quarter-quarter section. The lowercase letters are assigned in a counterclockwise direction, beginning in the northeast quarter. If two or more sites are located within the same quarter-quarter section, consecutive numbers follow the lowercase letters (Fig. 2).

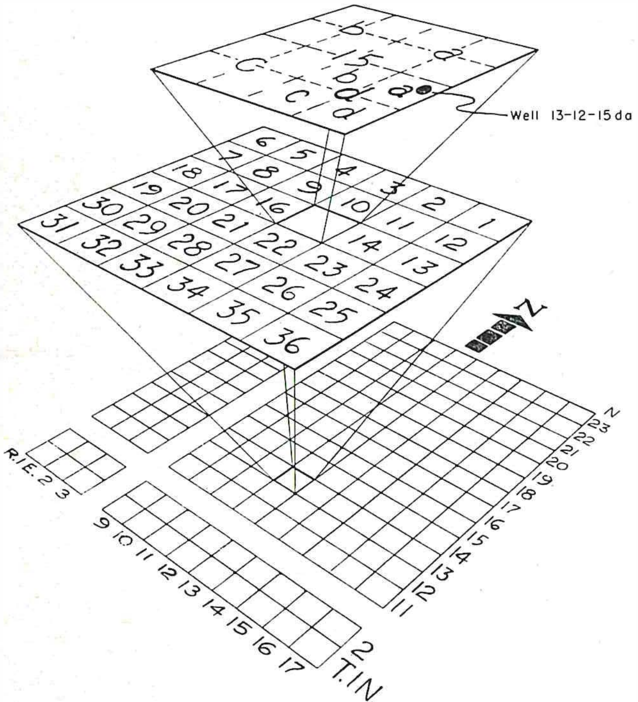

Figure 2.—Sketch showing system of numbering sites.

PREVIOUS INVESTIGATIONS

The earliest comprehensive descriptions of the geology and economic resources of the area covered by this report appeared in the Little Belt Mountains and Fort Benton Folios (Weed, 1899a and 1899b) and in a description of the geology of the Little Belt Mountains (Weed, 1900). Calvert (1909) reported on the geology and coal resources of the Lewistown coal field, and Fisher (1909) reported on the Great Falls coal field. Reeves (1929) discussed the faulting around the Highwood Mountains and its implications regarding oil possibilities. Reeves (1930) mapped the Big Snowy Mountains. Westgate (1921) investigated iron ore deposits on the flanks of the Little Belt Mountains. The rocks of the Highwood Mountains were studied by Larsen (1941). Perry (1932) prepared a reconnaissance report on the ground-water resources. Alden (1932) discussed the general physiographic history of the study area as a part of a much larger region. Vine and Hail (1950) prepared a preliminary map of the Hobson area, and Vine and Johnson (1954) prepared a map of the Stanford area. Vine (1956) prepared a report on the combined Stanford-Hobson areas. The Montana State Engineer (1963) summarized water rights in Judith Basin County.

METHODS OF INVESTIGATION

Geologic mapping by Vine (1956) was used for most of the study area. That part of the study area not covered by Vine's map was mapped on aerial photographs and later transferred to a base map.

An inventory was made and data compiled, when available, on the location, yield, depth to water, total depth, and formation(s) penetrated for most of the wells in the area, and on the location, yield, and source of many springs. The water levels in selected observation wells were measured periodically to determine the nature and amount of water-level fluctuations (Table 1, Part B). The altitudes at wells completed in the Kootenai Formation were determined by altimetry and instrumental leveling for use in the preparation of a piezometric map to show the general shape and slope of the piezometric surface for part of the study area.

The chemical characteristics of the water in the basin were determined from (1) chemical analyses by the Montana State Board of Health of water collected during this investigation from 25 well and stream sites, (2) chemical analyses that were made available from the files of the Montana State Board of Health and a private laboratory, Yapuncich, Sanderson, and Brown, and (3) data obtained on the conductivity of water from many wells and springs by the use of a portable conductivity meter.

The transmissibility of some of the water-bearing formations was determined in the field by means of seven flow-recovery and two pumping tests. Loss of flow from streams in the study area was investigated from records at two stream-gaging stations operated by the U. S. Geological Survey and from miscellaneous measurements of streamflow at several sites.

ACKNOWLEDGMENTS

The writer expresses his thanks to the many people and organizations whose courtesy and cooperation contributed to this study. Among these were O. C. Thatcher, a well driller who provided well logs; Yapuncich, Sanderson, and Brown Laboratory, who provided several water analyses; the Judith Basin Land Reclassification Committee, and the Agricultural Stabilization Service who aided in locating land corners on aerial photographs. Many residents aided the investigation by providing information on wells and springs and access to their land. The Judith Basin County Clerk and Recorder permitted access to well records filed in that office. The Soil Conservation Service provided data on well and spring developments done under its auspices. Local officials provided data pertaining to public water supplies.

GEOGRAPHY

TOPOGRAPHY AND DRAINAGE

The Judith Basin, a topographic depression in an unglaciated part of the northern Great Plains, is partly enclosed by the Little Belt, Highwood, Big Snowy, Moccasin, and Judith Mountains. The mountains rise to altitudes of approximately 9,000 feet. As the study extended only to the flanks of the mountains, however, the maximum altitude in the study area is about 6,000 feet.

The study area consists principally of rolling plains 3,500 to 6,000 feet above sea level. The plains are gently sloping, gravel-covered terraces dissected by commonly steep-sided stream valleys. Depth of stream valleys ranges from a few feet near the heads of small coulees to several hundred feet near stream mouths. Differential erosion of inclined strata near the mountains has resulted in moderate to strong local relief. The foothills thus formed are commonly capped by remnants of high-level terrace gravel. In the northern part of the study area, Arrow Creek and its tributaries have eroded deeply, producing a badlands topography.

Several other prominent topographic features are notable in the Judith Basin. Square Butte and the somewhat smaller Round Butte east of the Highwood Mountains are conspicuous landmarks that can be seen from most of the western part of the Judith Basin. They lie about 5 miles north of the study area. Wolf Butte is a prominent mountain peak, which stands out somewhat in front of the main mass of the Little Belt Mountains about 10 miles south of Geyser. A few miles west of Stanford, several gravel-capped remnants of a high terrace surface form the Stanford Buttes. Skull Butte, about 5 miles southwest of Stanford, is the topographic expression of a structural dome and towers nearly 1,000 feet above the adjacent plains.

Judith River, Arrow Creek, and their tributaries drain the Judith Basin. Principal tributaries to the Judith River in the study area are: Ross Fork, Sage Creek, Yogo Creek, Willow Creek, and Wolf Creek, the last formed by Running Wolf and Dry Wolf Creeks. Surprise Creek, Shannon Creek, McCarthy Creek, Old Geyser Creek, Braun Creek, and Cottonwood Creek are the major tributaries to Arrow Creek. None of the streams carry large perennial flows of water. The drainage pattern and major topographic features are shown on Figure 3.

CLIMATE

The climate of the western and southern parts of the Judith Basin is of the continental type and is semiarid in the basin to subhumid in the bordering mountains. The average annual precipitation ranges from 13 inches or less in parts of the study area to 28 inches in the Little Belt Mountains southwest of the area. About 70 percent of the precipitation in the basin falls during the growing season, which averages about 128 days at Stanford. Precipitation in the mountains bordering the study area is more evenly distributed throughout the year; about half falls as snow during the winter when it accumulates as a snowpack. Melting of the snowpack during the spring and early summer provides streamflow during this period.

The temperature in the study area is marked by extremes—most weather stations have recorded temperatures above 100°F and below -40°F. Summer temperatures are only moderately warm, maximum temperatures above 90°F occurring on about 15 days a year. During the winter months, arctic cold fronts move across the area, but severe cold usually lasts for only a few days and may be modified by warm "chinook" winds, which can raise the air temperature abruptly within a very short time.

Potential evaporation is between 50 and 55 inches in the study area—greatly in excess of precipitation. Evaporation is aided by low relative humidity during most of the year and by moderate to strong wind movement, especially in the fall and spring.

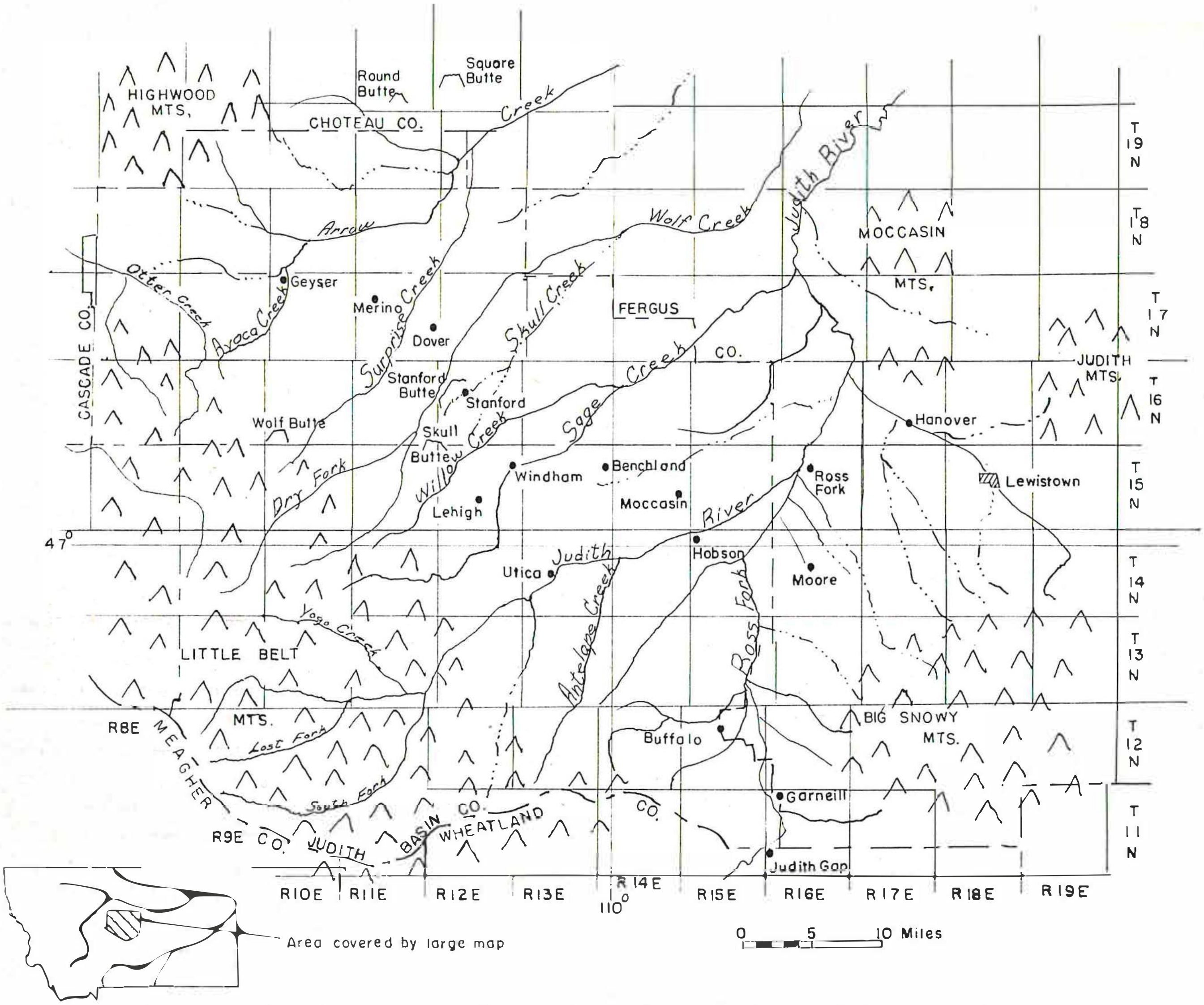

Figure 3.—Map of part of central Montana showing drainage and principal physiographic features.

GROUND WATER

The rocks that form the outer crust of the earth generally contain many openings termed voids or interstices. Below a certain level the voids are filled with water and form vast underground reservoirs for the storage and transmission of water. The water in these voids is termed ground water. Ground-water reservoirs are an integral part of the hydrologic cycle, which is the continuous cycle of water movement from the atmosphere to the land and oceans and back to the atmosphere by evaporation and plant transpiration (Fig. 4).

The water beneath the earth's surface can be divided into three zones, the zone of aeration, or vadose zone, the zone of saturation, and the zone of internal water. Within the zone of aeration are three belts, the belt of soil water, the intermediate belt, and the capillary fringe (Fig. 5). The belt of soil water is that part, directly below the surface, from which water may be discharged into the atmosphere by plants or by direct evaporation from the soil. The water in this belt is of great importance to agriculture, for it is near enough to the surface to be available to plant roots. The capillary fringe is a belt that overlies the zone of saturation and contains openings, some or all of which are filled with water that is continuous with the zone of saturation but is held above that zone by capillarity. Because capillarity can lift water to greater

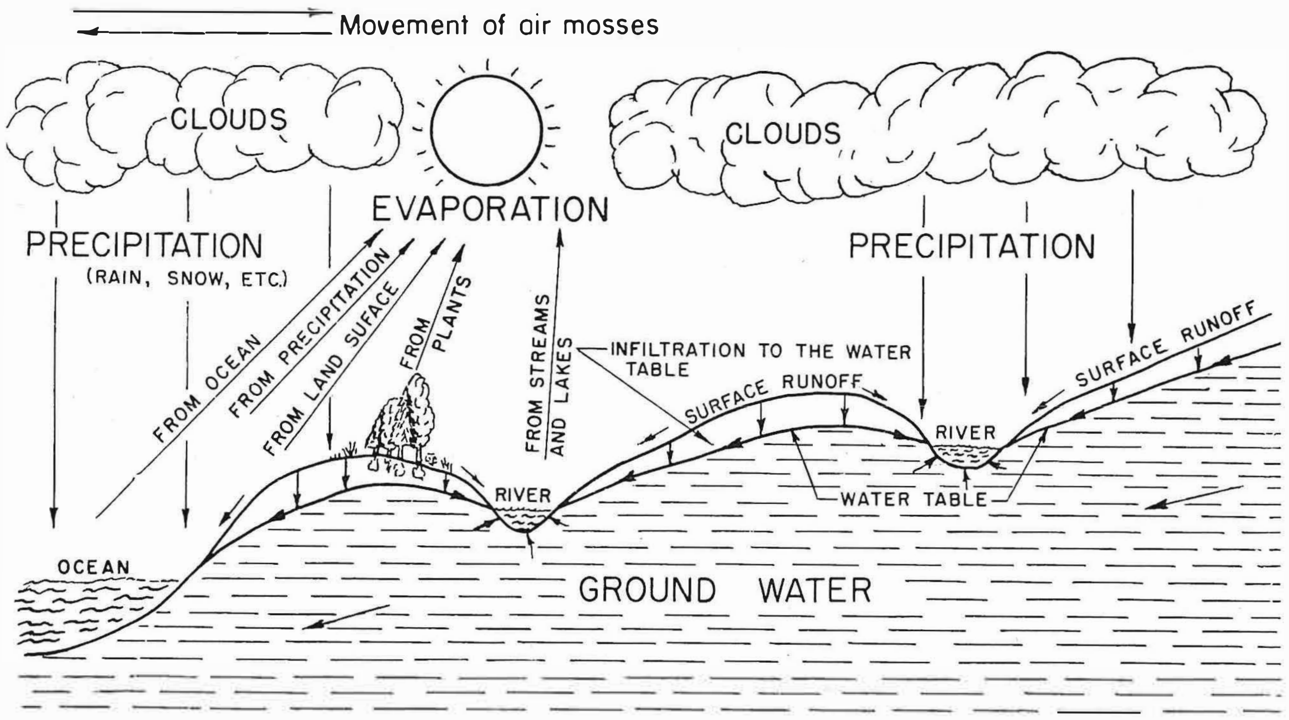

Figure 4.—Hydrologic cycle.

heights through very small openings than through large ones, the thickness of the capillary fringe depends on the texture of the rock or soil in which it occurs. The capillary fringe may be many feet thick in a fine-grained material, such as clay, or nonexistent in coarse material such as cobbles. The intermediate belt of the zone of aeration lies between the capillary fringe and the belt of soil water. Water that sinks into this belt is either drawn downward by gravity to the zone of saturation, is retained within the belt by molecular attraction as a coating on individual grains or a meniscus at grain contacts, or returns to the surface by capillary flow or as vapor. Water in the belt of soil moisture, the intermediate belt, or the capillary fringe is called vadose (hanging) water.

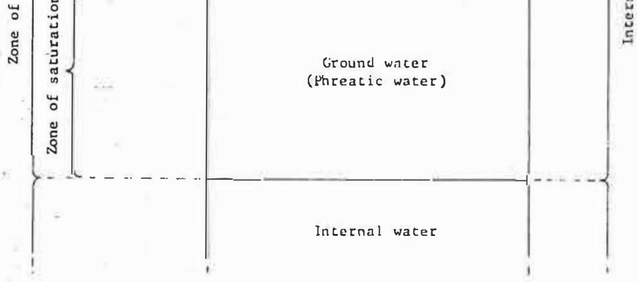

Figure 5.—Diagram showing divisions of subsurface water (From Meinzer, 1923b).

The water table marks the top of the zone of saturation in which all interconnected pore spaces are filled with water. Gravity is the predominant force acting on water in permeable material in this zone. Thus, water in this zone will flow to points of lower head such as in a pumping well or spring. The water in the zone of saturation is called ground water.

Below the zone of saturation is the zone of rock flowage in which temperature and pressure are so great that openings cannot exist in the rocks. In this zone water exists only in the molecular structure of the rocks and is called internal water. This zone is at tremendous depths and is, therefore, of little economic concern.

THE WATER TABLE

The water table, the upper surface of the zone of saturation in an unconfined aquifer, is the level at which water will stand in a well sunk into the zone of saturation. It is generally a sloping surface, having a gradiant in the direction of ground-water

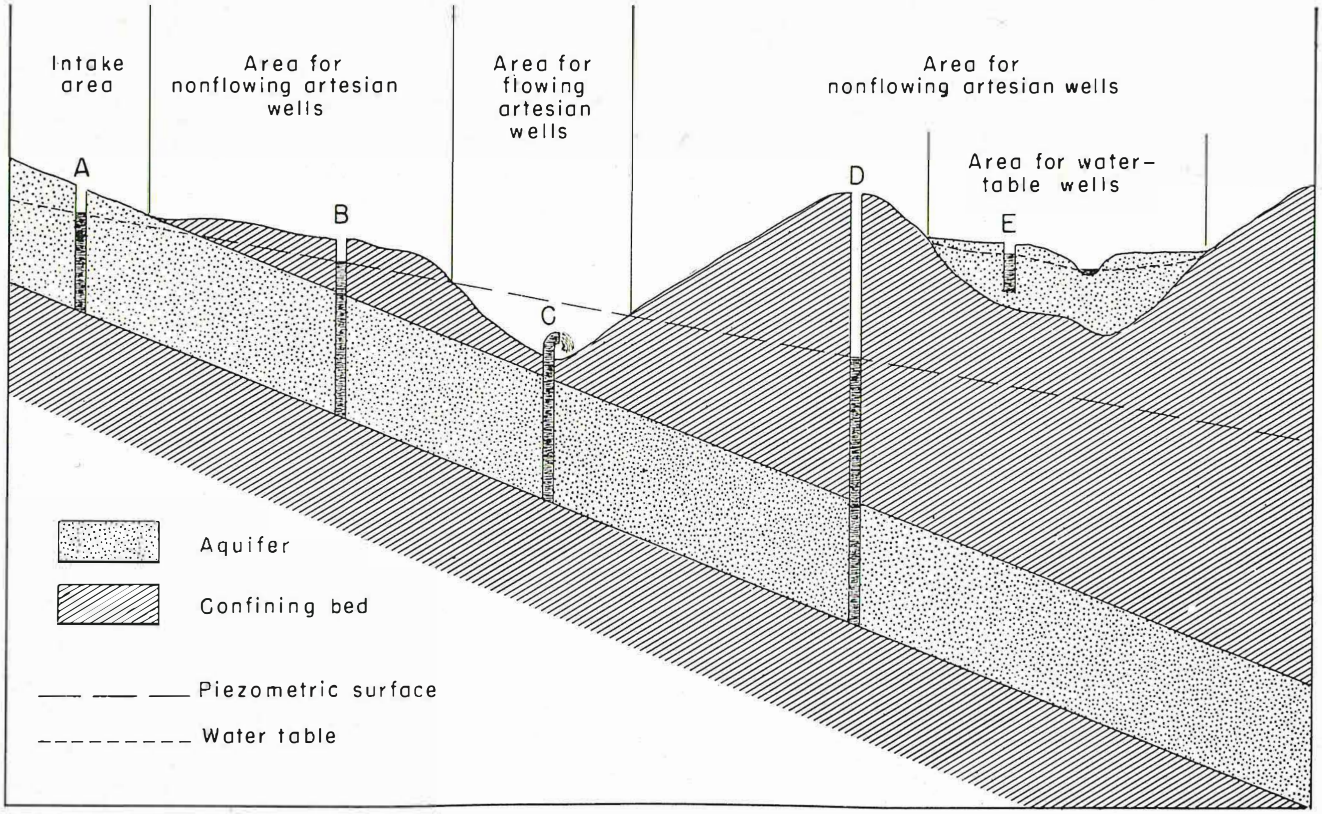

Figure 6.—Hypothetical section showing geologic environment necessary for unconfined (water-table) and confined (artesian) aquifiers.

movement. Movement is from points or areas of recharge, where water is added to the ground-water body, to points or areas of discharge. The slope of the water table depends upon the permeability (the ability of a rock to transmit water) of the rock materials below the water table and the amount of water transmitted. In materials of slight permeability the slope needed to move a given amount of water from a point of recharge to a point of discharge is greater than in materials of greater permeability. The presence of many different areas of recharge and discharge, and nonuniform permeability of the rock materials tend to make the water table an irregular surface.

In some places, an impermeable stratum may impede the downward movement of water through the vadose zone and cause it to form a zone of saturation. Water in such a zone is termed perched ground water. The water table of a perched ground-water body may be much higher than that of the underlying materials. Such perched water tables probably exist in the Judith Basin in places where igneous rocks intrude permeable sediments.

CONFINED GROUND WATER

Where ground water moves beneath a relatively impermeable stratum and is confined there under pressure, it is called confined or artesian water. If no saturated zone exists above the impermeable stratum, a water table does not exist there, and wells drilled will find no water until the impermeable stratum is penetrated. When the well is drilled into the underlying permeable material, the water rises in the well to a height corresponding to the pressure head on the artesian water.

The imaginary surface that represents the potential head to which artesian ground water will rise is called the piezometric surface. It is analogous to the water table. Each aquifer containing artesian ground water may have a different piezometric surface. In the Judith Basin, the exposures through which the artesian aquifers are mainly recharged are generally progressively higher for stratigraphically lower beds, and the piezometric surfaces of the older, deeper aquifers are usually higher than those of the shallower aquifers. Figure 6 illustrates the geologic environment necessary for water-table and artesian aquifers.

Contours (lines of equal altitude) on the approximate piezometric surface of water in the Kootenai Formation in the western part of the Judith Basin are shown on Plate 1 (in pocket).

RECHARGE

Recharge, the addition of water to the ground-water reservoir, may occur in several ways. The original source of almost all recharge is precipitation; however, in addition to direct infiltration of precipitation through the soil, ground water may be recharged by infiltration from influent streams (streams from which water sinks into the ground), by infiltration of irrigation water, by subsurface inflow of water from adjacent areas, or by interformational leakage from one bed to another.

In the Judith Basin, irrigation is practiced on only a small part of the land. Irrigation is unquestionably an important source of recharge to rocks underlying the irrigated land. Perennial streams are not abundant in the Judith Basin. Those that flow through the study area are commonly influent in at least part of their courses. Recharge from these streams is most conspicuous where they cross outcrops of the Madison Group, and it is discussed as part of the water-bearing characteristics of the Madison in this report. Direct infiltration of precipitation can take place throughout the study area, and it probably is the predominant form of recharge to most of the aquifers.

The amount of recharge by infiltration of precipitation is governed by many factors. Among these are: the amount of precipitation, the form of the precipitation (rain or snow), the intensity of the storms producing the precipitation, the slope of the land, and the vertical permeability of the soil and rocks. As the Judith Basin has a semiarid climate, direct recharge from precipitation probably is small. Many of the aquifers, however, crop out on the flanks of the mountains where precipitation is greater than it is in the center of the basin. The storms vary greatly in intensity but most are not so intense or long as to cause much runoff. The slope of the land also differs widely from place to place.. It is generally greatest near the mountains. The vertical permeability of the soils and rocks is generally much less than their lateral permeability.

In order for precipitation to recharge an aquifer it must first satisfy the demand for moisture in the belt of soil water. Soil moisture is being depleted almost constantly by evaporation and transpiration, hence is commonly deficient, and this zone absorbs and holds much of the moisture penetrating the land surface. When soil moisture requirements are met, the remaining water must penetrate the intermediate belt. The vadose water in this belt is also depleted by evaporation and capillarity so that part of the water is held in this zone. Thus, of the water absorbed into the ground, only a fraction may reach the zone of saturation. The amount of recharge derived from precipitation is probably only a small percent of the total precipitation, but when it is realized that 1 inch of recharge over an area of 1 square mile amounts to more than 17 million gallons, it is apparent that precipitation can be a source of much recharge. The actual amount of recharge from this source was not determined in this study, but it is probably the predominant source of recharge in the Judith Basin.

Subsurface flow from adjacent areas is probably insignificant in the ground-water regimen of the Judith Basin. Most of the adjacent land at higher altitudes than the study area is in the mountains and the outcropping formations generally dip under the basin to such great depths that the water in them is not economically recoverable under present conditions (1964).

Interformational leakage is difficult to evaluate. It is doubtless significant between shallow formations, such as between alluvium and underlying bedrock aquifers, but no means of measuring it were found in this investigation. Most of the bedrock aquifers are separated by thick beds of relatively impermeable rock. As the areas of these rocks in contact with one another are very great, and the difference in head may also be great, the amount of water transmitted through material of even very low permeability can be considerable.

DISCHARGE

Ground water may be discharged from the zone of saturation by evaporation, by transpiration from plants, by discharge into effluent streams (streams receiving ground water) or artificial drains, by springs, or by wells.

Water is discharged by evaporation in places where the zone of saturation is at the ground surface or so near that water can be lifted to the surface or to the zone of soil moisture by capillarity. This happens in some parts of the Judith Basin where the water table is forced to approach the surface because of a decrease in the ability of the water-bearing beds to transmit water, owing to reduced permeability of the material or to thinning of the beds. Most of these places are marked by marshy conditions and by the accumulation of various salts locally called "alkali".

Transpiration also is quantitatively important as a means of ground-water discharge, mainly in places where the zone of saturation is near the surface. The marshy places where ground water is at or near the surface are usually overgrown by water-loving plants such as cattails and sedges. The alluvium of many stream valleys supports thick growths of plants capable of sending roots down to the zone of saturation and using ground water from this zone. Such plants, called phreatophytes, include cottonwood and willow trees, chokecherry bushes, alfalfa (widely grown as a forage crop), wild roses, and greasewood. Greasewood is very tolerant of salinity but in the Judith Basin is almost restricted to the lower reaches of Surprise and Arrow Creeks, in the northern part of the study area. In some parts of the west, some phreatophytes are recognized as wasting economically significant quantities of water, and strenuous efforts are made to eradicate them. In the Judith Basin most of the wild phreatophytes are valued, at least to some extent, as cover for livestock and game and as forage for game. They are eradicated where they impede cultivation or livestock management but not ordinarily to conserve water.

Many of the streams in the Judith Basin are effluent in part of their courses and receive much of their flow during a part of the year from springs or seepage along their banks. Much of the streamflow attributable to ground-water inflow is counterbalanced by losses to ground-water reservoirs, hence determination of the total quantity of ground water discharged by streams would require very detailed measurements beyond the scope of this study.

Few artificial drains have been constructed in the study area. Discharge by this means is therefore small.

The locations of 1,378 wells and springs are shown on Plates 1 and 2 (in pocket). The discharge of wells and springs ranges from 1 gpm (gallon per minute) or less to about 1,000 gpm. The total discharge of ground water from wells in the study area is not known but is probably small in comparison to discharge by natural means.

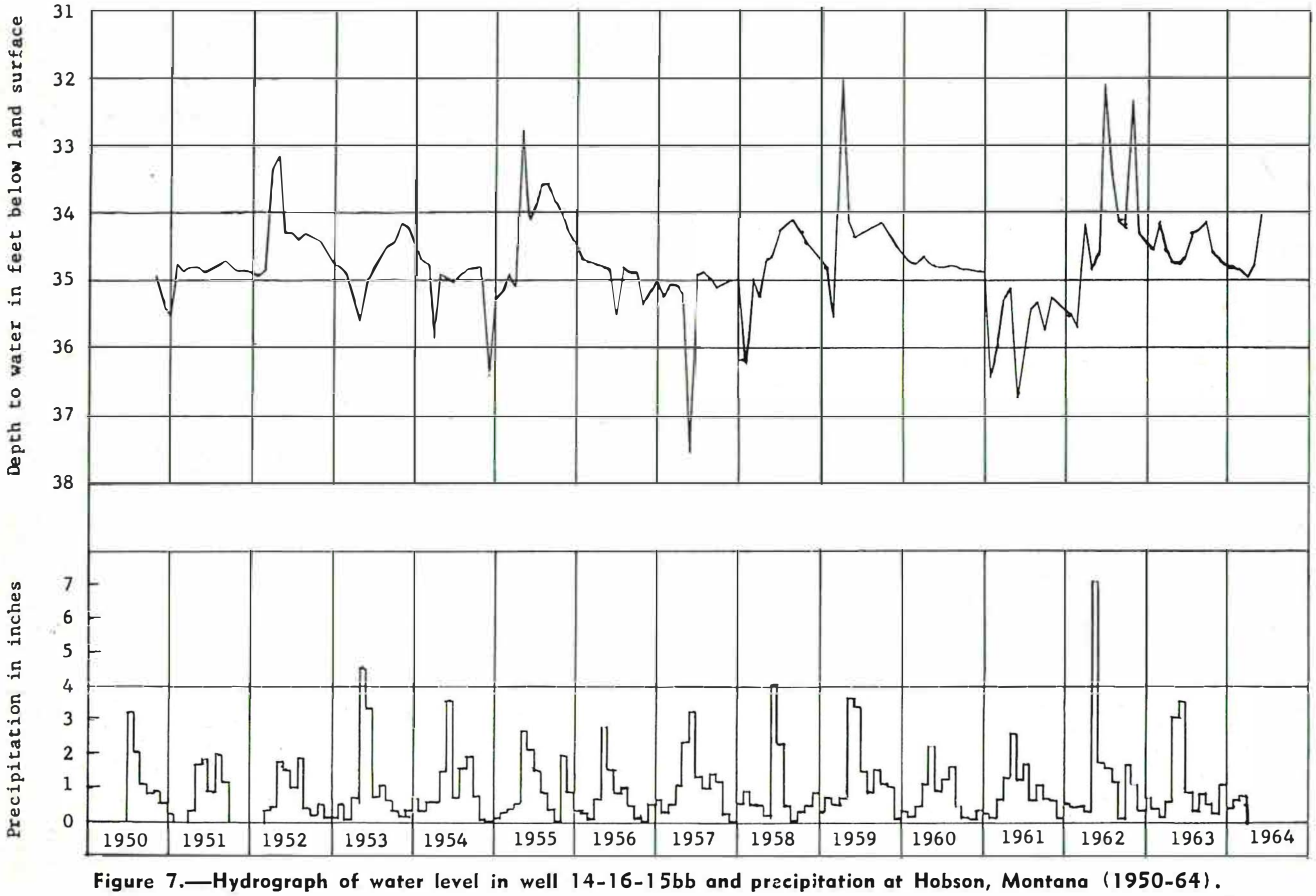

Figure 7.—Hydrograph of water level in well 14-16-15bb and precipitation at Hobson, Montana (1950-64).

WATER-LEVEL FLUCTUATIONS

The water table or piezometric surface in an aquifer is not a static surface but fluctuates because of changes in the amounts of recharge to and discharge from the aquifer. In any ground-water reservoir under natural conditions the amount of water discharged will, over a long period, be balanced by recharge. If discharge from the reservoir is increased, as by withdrawals from wells, without a corresponding increase in recharge, either the natural discharge from the reservoir will decrease or water will be withdrawn from storage, resulting in lowering of the water table or piezometric surface. Thus, changes in water levels reflect changes in ground-water storage. Water levels in 78 wells in the western and southern parts of the Judith Basin were measured periodically during this study to determine the short- and long-term fluctuation of the water table or piezometric surface of the principal aquifers. The records of these measurements are presented in Table 1 in Part B of this report. The hydrograph of well 14-16-15bb, shown in Figure 7, illustrates a general correlation between the water level in this well and precipitation at Hobson, Montana, the nearest Weather Bureau precipitation station. The hydrograph shows that the annual fluctuation is about 1 foot to 3 feet, but that the maximum range during the period of record was more than 5 feet.

HYDRAULICS OF WELLS AND SPRINGS

When a well begins discharging, the water level in the well and in the surrounding aquifer declines. The level to which it declines for a given rate of discharge is such that the water movement toward the well in the surrounding aquifer equals the rate of discharge. An increase in discharge necessitates additional lowering or drawdown of the water level.

Under water-table conditions, the zone of water-level lowering around a well takes the form of an inverted cone, the well being at the axis, and is called the "cone of depression". The size and shape of the cone of depression depend on the permeability of the aquifer, the rate at which the water is discharged, and the time since discharge began. To supply equal amounts of water to a well, water-bearing materials of greater permeability require less drawdown (flatter cone of depression) than do materials of lesser permeability (Fig. 8). The cone of depression around a pumped well will increase in depth and area until it intercepts enough recharge or reduces other discharge by an amount equal to the well discharge. The areal extent of the cone of depression is termed the "area of influence". The water level will be lowered in other wells within the area of influence of a well that is being pumped. This lowering is called "interference".

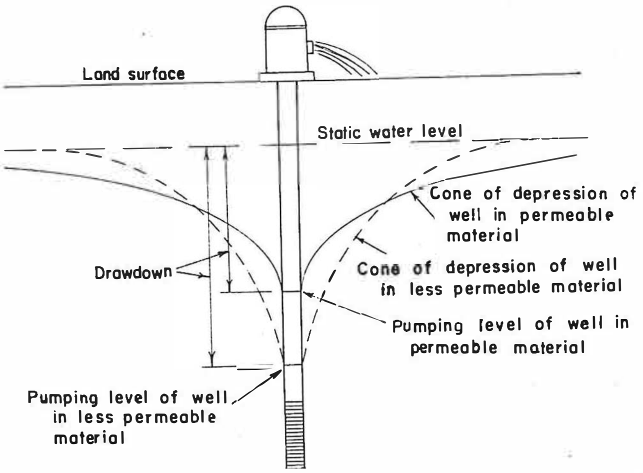

Figure 8.—Diagrammatic section through well being pumped showing drawdown, cone of depression, and differences in shape of cone of depression caused by difference in permeability of water-bearing material for same well yield.

The hydraulic characteristics discussed above apply also to artesian wells, except that the drawdown will be from the piezometric surface, which may be above the ground surface, and the cone of depression will be the lowering of the head or piezometric surface around the well.

The rate at which the area of influence of a pumping well expands depends upon the coefficient of storage and upon the permeability of the aquifer. The coefficient of storage is defined as the volume of water released from or taken into storage per unit surface area of the aquifer per unit change in the component of head normal to that surface.

The area of influence of a well in a very permeable aquifer will expand more rapidly than the area of influence of a well in less permeable materal if the storage coefficients of the two materials are equal. Thus, a well in more permeable material has a greater specific capacity (yield per unit of drawdown) than one in less permeable material, but its area of influence becomes large and interference with other wells and springs becomes appreciable much sooner than for a well in less permeable material.

The head in an artesian well is not ordinarily lowered below the top of the aquifer, and thus the aquifer material is not drained. The storage coefficient of an artesian aquifer represents the volume of water per unit of head change derived from compression of the aquifer material and the water. The storage coefficient of an artesian equifer is usually very small and may range from about 0.00001 to 0.001. Because of the low coefficients of storage, wells in artesian aquifers usually develop large areas of influence very rapidly.

The behavior of wells provides the basis for determining aquifer characteristics. These characteristics can also be determined by laboratory tests conducted on samples of the material, but unavoidable sampling errors and disturbance of the samples make these tests less reliable than field tests. The field tests, often called "pumping tests" but more properly called "aquifer tests", rely on precise observations of the changes in water levels in wells in response to accurately known discharge. Several basic simplifying assumptions must be made in order to interpret the observations obtained in an aquifer test. As the assumptions are rarely completely fulfilled in nature, some errors are introduced but they are generally small. Detailed discussion of aquifer-test analysis is beyond the scope of this report. The reader is referred to Ferris and others (1962) for a more thorough discussion.

Aquifer tests on seven flowing artesian wells were conducted during this study. The results of one of the tests were rejected because of a suspected casing leak. Two water-table aquifer tests were conducted. The artesian aquifer tests were made by allowing each well to flow at a known rate for a known period of time and then shutting off the flow and allowing the head to recover. From the measured changes in head during recovery, the field coefficient of transmissibility was computed by using the modified nonequilibrium formula (Ferris and others, 1962, p. 100). The field coefficient of transmissibility is the amount of water, in gallons per day, at the prevailing temperature, that is transmitted across each mile strip extending the saturated thickness of the aquifer under a hydraulic gradient of 1 foot per mile. The water-table aquifer tests were conducted by pumping a well and observing the drawdown in the pumped well and one observation well. The observations were evaluated by use of the Theis nonequilibrium formula (Theis, 1935). Neither of the tests was long enough to permit valid evaluation of the storage coefficient. The transmissibility coefficients obtained are discussed with the water-bearing characteristics of the formations in which the wells were constructed.

GEOLOGY

Sedimentary strata, ranging in age from Mississippian to Recent, are exposed in the study area. Older formations, including 850 to 1,300 feet of Cambrian rocks and 130 to 165 feet of Devonian rocks, were described by Weed (1899a) and Reeves (1930) in the Big Snowy and Little Belt Mountains. Igneous rocks intruded the sedimentary strata sometime between the end of the Cretaceous Period and deposition of Quaternary sediments, probably during early Tertiary time. At about the same time volcanoes at the present site of the Highwood Mountains ejected a large quantity of lava which, along with intrusive rocks, forms the Highwood Mountains. Table 1 shows the stratigraphic relations of the rocks of the Judith Basin. The distribution of the rock units is shown on the geologic maps (Pl. 3 and 4, in pocket).

STRUCTURE

The Judith Basin is a large northward-plunging syncline bordered by anticlinal folds, which form the Big Snowy, Little Belt, Judith, and Moccasin Mountains. The Highwood Mountains bound the syncline on the northwest, but their emplacement did not greatly disrupt the synclinal structure. Small faults and domes greatly modify the regional structure.

The structure of the study area affects the occurrence of ground water by its relation to artesian conditions and by limiting the distribution of water-bearing beds. Erosion has beveled and truncated the sediments on the flanks of the folds.

To predict the depth to a particular formation at a given site one must take the structure of the rocks into account. Structure contours drawn on the top of the Kootenai Formation are shown on Plates 3 and 4. By the use of these contours and the altitude of a particular site one can determine approximately how deep a well at the site must be to reach the Kootenai Formation. If a well is to be drilled to a deeper bedrock formation the approximate depth required can be computed by adding to this figure the thickness shown in Table 2 between the top of the Kootenai and the base of the other formation. For example, a well to be drilled at the southwest corner of sec. 15, T. 17 N., R. 12 E. (in the vicinity of Dry Wolf and Running Wolf Creeks on Table 2) would start at an altitude of about 4,000 feet. The structure contour (Pl. 3) indicates the altitude of the top of the Kootenai at this location to be about 3,000 feet. Therefore, a well 1,000

Table 1.—Stratigraphic units and their water-bearing characteristics in the study area.

System	Series	Formation		Maximum thickness (feet)	Lithologic character	Water supply
Quaternary	Recent	Alluvium		30(?)	Unconsolidated gravel, sand, silt, and clay.	A good source of water locally where materials are coarse, well sorted, and adequately recharged.
	? Pleistocene(?)	Terrace deposits		100	Unconsolidated sand and gravel. Very widespread.	Most used source of ground water in study area. Yields enough water for stock and domestic needs.
Tertiary		Extrusive rocks		?	Quartz latite and mafic phonolite.	Not important as an aquifer in the study area.
		Intrusive rocks		?	Dark quartz-free rocks forming dikes and sills; quartz-monzonite porphyry forming stocks and laccoliths.	Not an aquifer.
Cretaceous	Upper Cretaceous	Montana Group	Judith River Formation	400(?)	Gray to brown sandstone, sandy shale, clay, and some coal.	Yields water to some wells in the small outcrop area.
			Claggett Shale	500	Gray and brown shale. Poorly exposed.	Poor aquifer.
			Eagle Sandstone	300	Massive tan and white sandstone; contains some sandy shale beds.	A fair aquifer but present in only the southeastern part of the study area.
			Telegraph Creek Formation	160	Sandy shale beds; transitional between the Colorado Shale and the Eagle Sandstone. Mapped with the Colorado Shale.	Not a good aquifer.
		Colorado Shale		1,500	Fissile gray or black marine shale, thin local sandstone beds, numerous concretionary zones, and bentonite beds.	Generally not water bearing. A few sandy beds, particularly the First Cat Creek sand of drillers at the base, may yield moderate amounts of water.
	Lower Cretaceous	Kootenai Formation		500	Red shale and brown to gray sandstone. A few thin nodular fresh-water limestone beds.	Sandstone beds, particularly the basal bed, constitute a widely used aquifer. Flows obtainable in structurally favorable places.

Jurassic		Upper Jurassic		Morrison Formation	300	Variegated shale and siltstone; thin, nodular limestone; white and brown sandstone; black shale; and a bed of coal or carbonaceous shale near the top.	Not generally favorable as a source of ground water.
			Ellis Group	Swift Formation	165	Glauconitic sandstone and shale.	Can yield moderate amounts of water. Flows obtainable from this formation in lowermost parts of the basin.
				Rierdon Formation	200	Gray calcareous shale. Missing in outcrops through most of the area.	Not an aquifer in the study area.
		Middle Jurassic		Piper Formation	200	Red to varicolored shale and siltstone, thin limestone and shale beds. Absent in most outcrops in the study area.	Not an aquifer.
Carboniferous	Pennsylvanian	?		Amsden Formation	920	Lower unit of red shale; siltstone; and red, brown, or white sandstone and conglomerate. Upper unit of gray limestone, some red and gray shale partings.	Promising as an aquifer but thins and disappears basinward.
	Mississippian	Upper Mississippian	Big Snowy Group	Heath Shale	500	Fissile carbonaceous shale; dense black and gray limestone; sandstone in upper part.	Unfavorable as a source of water except for sandstone in upper part.
				Otter Formation	500	White to gray limestone and gray to green shale.	Not a good aquifer but yields water to springs.
				Kibbey Sandstone	300	Red, white, or yellow sandstone, siltstone, limestone, and gypsum.	Offers promise of good yields out in the basin, but the water table has not been found in the outcrop zone. Probably hydraulically interconnected with underlying Madison Group.
		Upper and Lower Mississippian		Madison Group	1,000	Massive gray dense limestone, greatly fractured and brecciated in upper part. Numerous solution openings.	Potentially a copious aquifer, but water is confined in discrete solution channels whose location cannot be determined on the ground surface. Water level seems to be generally very deep in the outcrop zone.
Pre-Mississippian				Older formations	?	Not studied but consist mainly of limestone and shale beds over crystalline Precambrian rocks.	Not economically important as aquifers in the study area.

Table 2.—Thicknesses of geologic units

Distance or depth in feet from top of the Kootenai Formation to the base of other units in indicated localities

Unit	Lone Tree Creek	Surprise Creek	Dry Wolf and Running Wolf Creeks	Willow Creek	Skull Butte	Antelope Creek and Wait Creek	Ross Fork
Kootenai Formation	450	450	450	450	550	350	375
Morrison Formation	700	700	700	700	715	700	650
Ellis Group	775	775	775	775	775	800	710
Amsden Formation	1,095	945	1,475	1,485	1,134	....	....
Otter Formation	1,920	1,770	2,365	2,465	....	1,800	1,800
Kibbey Sandstone	2,070	1,920	2,615	2,655	....	2,100	2,100
Madison Group (upper 200 feet)	2,270	2,120	2,815	2,855	....	2,300	2,300

(Modified from Bulletin 1027-J)

feet deep should penetrate approximately to the top of the Kootenai. If one wants to drill a well to fully penetrate the Kootenai, 450 feet should be added to the figure for the depth to the top of the Kootenai. Thus a well fully penetrating the Kootenai should be about 1,450 feet deep at the site. Similarly, the depth of a well drilled to the base of the Kibbey Sandstone would be the sum of the depth to the top of the Kootenai (1,000 feet at the described site) plus the distance from the top of the Kootenai to the base of the Kibbey (2,615 feet according to Table 2) or about 3,615 feet. Depths to the base of the Eagle Sandstone, the base of the Mowry Shale equivalent, and the base of the Muddy(?) Sandstone member of the Thermopolis Shale equivalent would be, respectively, about 1,920 feet, 720 feet, and 400 feet less than the indicated depth to the top of the Kootenai, in most of the area.

In places where erosion has removed part or all of the Kootenai, the structure contours indicate an altitude higher than that of the land surface. To estimate the depth needed to penetrate a deeper formation, the difference between the land-surface altitude and the structure contour must be subtracted from the depth of the deeper formation obtained from Table 2. For example, assume that a well penetrating the Amsden Formation is desired at site 16-11-35bd, in the vicinity of Dry Wolf and Running Wolf Creek (Table 2). This site is nearly on the 5,000-foot structure contour (Pl. 3), but the altitude of the site is about 4,650 feet. The difference of 350 feet between the altitude of the site and that of the top of the Kootenai must be subtracted from the depth of 1,475 feet to the base of the Amsden Formation as shown on Table 2. Thus the well should be about 1,125 feet deep.

GEOLOGIC UNITS AND THEIR WATER-BEARING PROPERTIES

OLDER ROCKS

The Madison Group is the oldest unit mapped during this study. Older formations, including 850 to 1,300 feet of Cambrian rocks and 130 to 165 feet of Devonian rocks, were described by Weed (1899a) and Reeves (1930) in the Big Snowy and Little Belt Mountains. These rocks have received little consideration in this study because they underlie the study area at great depth, and their hydraulic properties have not been tested by wells. An oil test hole drilled on the north flank of the Little Belt Mountains (oil test 12-14-19bd) penetrated some of the older rocks, and tests were made to determine the fluid content (drill-stem test). In the section from 2,809 to 2,886 feet, water was found in Devonian rocks and it rose 2,061 feet in the drill stem. The hydrostatic pressure recorded indicates that it would have risen to within about 60 feet of the top of the hole in time. As the hole was drilled at an altitude of 5,139 feet, this suggests that a flow of water could be obtained from Devonian rocks at altitudes lower than about 5,080 feet.

MADISON GROUP

Outcrops of the Madison Group along the flanks of the Little Belt and Big Snowy Mountains form part of the boundary of the study area. According to Weed (1899b, p. 2) the Madison is about 1,000 feet thick in the Little Belt Mountains. It crops out prominently, forming cliffs and hogbacks on the mountain flanks.

The Madison is made up of three formations: the Lodgepole, Mission Canyon, and Charles. The Charles is the uppermost and consists mainly of evaporite layers (gypsum, anhydrite, and perhaps salt) interbedded with limestone. It is not exposed, but beds totaling as much as 200 feet in thickness, noted in oil test holes drilled in the Judith Basin, have been assigned to the Charles Formation. Post-Madison erosion probably removed the beds in the present outcrop area of the Madison. I. J. Witkind (oral communication, 1963) suggested that part of the limestone breccia now included in the Mission Canyon Limestone may be derived from collapse of limestone beds of the Charles as more soluble evaporite minerals were leached out. The Charles would be unfavorable as a source of ground water.

Only the uppermost beds of the Mission Canyon Limestone of the Madison were studied during this investigation. These beds consist of gray dense limestone, which is greatly fractured and brecciated and contains numerous openings ranging from pinpoint vugs to caves. In some places erosion has removed part of the rock around caves and has exposed spectacular arches. A large part of the upper section of the Mission Canyon Limestone is breccia formed by collapse of old caves and cementation of the roof fragments.

The vuggy and cavernous openings in the Madison give rise to a high but erratically distributed permeability in some places. Few successful water wells have been drilled into the Madison in the study area and these few were drilled at sites where igneous intrustions have probably interrupted the flow of water through solution channels and caused a perched water table. Spring 14-11-6cb, though discharging from alluvium, probably rises at that particular point because a body of quartz monzonite porphyry has intruded the Madison, thus forming a ground-water dam that prevents passage of water in the limestone. Unsuccessful attempts to obtain water from the Madison in or near the outcrop zone indicate that the water table is very deep in many places. A nearly vertical hole in the Madison was uncovered by a road crew in the canyon of the South Fork of the Judith River (at approximately site 12-10-1ab). The writer tried to determine the presence of or depth to water in this hole in 1962 but found that it exceeded 250 feet—the length of the steel tape used. According to local reports, a man was lowered into this hole to an unreported depth. He did not reach the bottom nor see any water but reported that he could hear cascading water. This hole is only about 35 feet from the South Fork of the Judith River, which was flowing about 15 cfs of water at the time of the writer's visit.

Many oil companies have drilled into the Madison in central Montana in search of oil. Some of the oil test holes yielded water from the Madison in such copious amounts that it proved difficult to control. None of these oil test holes drilled in the study area are known to have produced much water from the Madison, but one drilled north of the study area (site 19-12-34aa) tested a section of the Madison at a depth of 3,030 to 3,080 feet. The test indicated that the water in the Madison was under sufficient pressure (final shut-in pressure) to rise to an altitude of about 3,880 feet above sea level at the site. The porosity of the upper part of the Madison in this oil test hole is as much as 16 percent.

The flow of many streams heading both in the Big Snowy and Little Belt Mountains decreases markedly where they cross outcrops of the Madison Group. The loss of water is due mainly to infiltration into the cavernous limestone. The loss of water is particularly conspicuous in Dry Wolf Creek. As the name implies, the creek is dry in its lower reaches most of the time. During moderately high stages, water flows to a sinkhole in the Madison at site 15-10-14cb. During this study, streamflow measurements were made at several sites to determine the amount of loss into the Madison. On May 29, 1964, a loss of approximately 10 cfs (cubic feet per second) was noted at the sinkhole. Between the Dry Wolf Ranger Station in sec. 31, T. 15 N., R. 10 E., and an irrigation diversion on the left bank at site 15-10-27ab Dry Wolf Creek was losing 90 cfs of water on May 28, 1964. Of the remaining water, approximately 42 cfs was diverted for irrigation and about 44 cfs remained in the creek. The 44 cfs in the creek and about 30 cfs of return flow from the irrigation diversion were lost within about 3 miles of the diversion. Seemingly, this water cannot be conducted out of the outcrop area as fast as it can be absorbed by the rocks, and part of it is held in storage. Measurements on June 3, 1964, indicated a loss of about 64 cfs between the Dry Wolf Ranger Station and the sinkhole at site 15-10-14cb. The sinkhole was accepting no water. The remaining 40 cfs of water flowed downstream from the Madison outcrop zone and was absorbed by alluvial gravel, but by the evening of June 8, 1964, this gravel was saturated and Dry Wolf Creek flowed all the way to its confluence with Running Wolf Creek. According to local reports, Dry Wolf Creek flows to the confluence only about once in seven years.

Other streams lose appreciable amounts of water as they cross the Madison outcrop. The Middle Fork of the Judith River was losing approximately 100 cfs between site 12-11-6aa and the Geological Survey gaging station at site 13-12-17ba on June 4, 1964. Yogo Creek was flowing 42 cfs at site 13-11-18db (approximately) on May 20, 1964, but was dry at a trail crossing about 3 miles downstream (site 13-

11-27ca). A measurement of the South Fork of the Judith River made on May 20, 1964, indicated a loss of 13 cfs between the measuring point (12-11-1ab) and a U.S.G.S. gaging station about 6 miles upstream (12-11-34ca). This was less than 10 percent of the flow measured, so measurement error was suspected. On April 28, 1964, however, the South Fork was dry for several miles upstream from its confluence with the Middle Fork; according to local report, water did not flow to the confluence until May 1. The gage at the upstream station recorded an average flow of 11 cfs on April 28, 17 cfs on April 29, and an increase thereafter. Therefore, 13 cfs seems a reasonable figure for the loss from the South Fork.

No accurate figures are available for the total loss of water into the Madison Group. As the Madison crops out in approximately 1,000 square miles in the Little Belt Mountains, the loss must be great. Most of the area underlain by the Madison is similar in climate and topography to that above the upper stream-gaging station on the South Fork of the Judith River. The land above the gaging station is mostly underlain by relatively impermeable rocks. During the period from October 1, 1958, to September 30, 1959, (arbitrarily chosen) the runoff from the land above the upper gaging station (above the Madison outcrop) was 3.44 inches per year, and that from the land above the lower gaging station (13-12-17ba) was only 1.65 inches per year. Thus, runoff from the mountains across the Madison outcrop is less than half that where it does not cross the Madison. Losses other than by infiltration, such as evaporation and transpiration by plants, should be comparable in both drainage areas. If infiltration losses are similar in all parts of the Little Belt Mountains underlain by the Madison, there should be a total loss of about 180,000 acre-feet per year (or about 250 cfs) from runoff from the mountains.

The points of discharge to which water absorbed by the Madison is conducted are not definitely known. Several large springs are known in central Montana, but they are not clearly related to the Madison. The largest of these springs is Giant Springs near Great Falls. It discharges about 600 cfs (Meinzer, 1927, p. 82). Others are the Lewistown Big Springs about 7 miles southeast of Lewistown, which discharge about 140 cfs (Calvert, 1909, p. 53), and Warm Spring about 12 miles north of Lewistown, which discharges between 180 and 190 cfs (Meinzer, 1927, p. 82). The Lewistown Big Spring is at an altitude of about 4,150 feet, which is higher than the 3,880-foot head noted in the oil test at site 19-12-34aa. Thus, it probably does not discharge much water from the Little Belt Mountains though it is well situated to discharge water from the Big Snowy Mountains. Warm Spring, at an altitude of 3,775 feet, probably discharges water that recharges rocks in the Judith and Moccasin Mountains.

Knowledge of the large losses of water from streams into the Madison Group has aroused interest among some Judith Basin residents in the possibility of recovering some of the water for irrigation use by deep wells. The high cost of drilling to the required depths discourages test drilling, and the fact that the location of the discrete openings through which water passes cannot be ascertained from surface indications makes the outcome of drilling at a given site impossible to predict. The Madison is a potential source of large amounts of ground water, but its development will require expensive and risky test drilling.

BIG SNOWY GROUP

The eroded upper surface of the Madison Group is overlain by three formations that constitute the Big Snowy Group. These formations of Late Mississippian age are, in ascending order, the Kibbey Sandstone, the Otter Formation, and the Heath Shale.

Kibbey Sandstone.—Red sandstone, siltstone, and shale constitute the lowermost 50 to 100 feet of the Kibbey Sandstone. The upper part consists of white or yellow fine- to medium-grained porous commonly friable clean sandstone overlain by interbedded siltstone, limestone, and gypsum, but the upper part is not well exposed. Thickness of the formation ranges from 150 to nearly 300 feet, according to Vine (1956, p. 421).

The friable porous sandstone beds in the upper part of the Kibbey should form a good aquifer. One dry hole was drilled into the Kibbey at site 13-12-3 near the outcrop. This disappointing result may be contrasted with a yield of about 2,200 gpm from the Kibbey Sandstone in an oil test hole near Hanover, Montana (Perry, 1932, p. 26) about 10 miles northwest of Lewistown. The comparison suggests that, although the formation can transmit water readily, the water table near the outcrop is at a considerable but unknown depth. Well 16-10-17cb flows about 5 gpm and is at a site where the Kibbey is at fairly shallow depth. Unfortunately, neither the depth of this well nor a log of formations penetrated was obtained. Well 16-10-18da, 100 feet deep, probably penetrated the top part of the Kibbey. Water stands at a depth of 30 feet in this well.

The porous permeable sandstone beds of the Kibbey overlie the irregular eroded surface of the cavernous upper part of the Madison, suggesting that the two formations may be hydraulically interconnected. Consequently, the water tables or piezometric surfaces of the formations may nearly coincide, and one may be recharged by the other through interformational leakage.

Otter Formation.—The Otter Formation overlies Kibbey Sandstone and consists of 400 to nearly 500 feet of limestone and green, gray, and purple shale (Vine, 1956, p. 422). Where exposed, the Otter may be easily recognized by the characteristic green shale, which constitutes much of the upper 200 to 300 feet.

Because the rocks of the Otter and the overlying Heath Shale are easily eroded, they are generally soil covered and exposures are few. Though limestone beds in the Otter give rise to some fairly large springs in the outcrop areas, the formation does not seem sufficiently promising as an aquifer to justify drilling into it.

Heath Shale.—The Heath Shale, composed mainly of fissile carbonaceous shale, includes some dense black and gray thin-bedded limestone and sandstone in the upper part. The carbonaceous shale is petroliferous in some places, and has yielded as much as 9 gallons of oil per ton of shale. Gypsum deposits are found in the Heath at some places, and gypsum was mined for a short time in sec. 16, T. 16 N., R. 10 E.

The shale weathers readily into soil that mantles most outcrops. It forms slopes that become slippery when wet, facilitating landslides of the Heath and overlying Amsden Formation. These landslides further obscure outcrops of the shale. Aragonite fragments on the weathereed soil commonly provide a clue to the presence of the Heath.

The Heath Shale is discontinuous in outcrop. Vine and Hail (1950) attributed this to either nondeposition or to lateral gradation of the black shale into green shale of the Otter Formation or red shale of the Amsden Formation. Walton (1946, p. 1304) thought the discontinuity was caused by erosion subsequent to deposition. Where exposed, the formation attains a maximum thickness of almost 500 feet.

Despite the fact that many springs issue from the Heath Shale, it is not a likely source of ground water. Local 10- to 20-foot sandstone beds near the top of the formation may yield water to wells, but these beds are discontinuous and thus hard to find in drilling. The yield from wells drilled into the Heath would probably be small.

AMSDEN FORMATION

The Amsden Formation, of Mississippian and Pennsylvanian age, is as much as 920 feet thick near the mountains but thins and disappears out in the basin, owing to post-depositional erosion (Vine, 1956, p. 430). Figure 9 is a thickness map of the Amsden showing this thinning and disappearance. The formation can be divided into several lithologic units. Two units are recognizable in most exposures—a lower clastic unit of red shale and siltstone and red, brown, or white massive sandstone and conglomerate, and an upper unit of gray limestone in thick beds separated by red and gray shale partings. In addition to the above-mentioned units, a shale bed overlies the upper unit near Wait Creek, and a marine limestone bed 30 feet thick underlies the lower clastic unit near Antelope Creek. West of Dry Wolf Creek, the upper unit thins and locally is missing.

The sandstone beds of the Amsden are commonly friable and, although fine grained, should be a fairly good source of ground water. The limestone beds of the upper part of the Amsden yield water to some springs in the area. Well 15-12-22cd was drilled as an oil-test hole but plugged back to the Amsden and converted to a water well. The owner reports a yield of 1,000 gpm. When the writer visited the well the casing and valve were corroded and the reported maximum yield was not verified. The well was flowing about 50 gpm through a partly opened valve and several leaks.

Thus, the Amsden seems to be a potential source of ground water but, because of thinning, its potential is great enough to justify drilling only south of U. S. Highway 87.

ELLIS GROUP

The Piper, Rierdon, and Swift Formations (in ascending order) constitute the Ellis Group of Middle and Late Jurassic age. These sedimentary rocks were deposited in a sea spread over an irregular surface on the Amsden Formation and older rocks. The Piper and Rierdon Formations were deposited only on the lowest parts of this surface in the northern part of the study area. Only the Swift Formation covered the higher parts of the erosional surface near the present Little Belt and Big Snowy Mountains. Thus, the Ellis Group differs greatly in thickness along the front of the Little Belt Mountains and is locally missing. Where penetrated by oil-test holes, the group is as much as 500 feet thick where all three formations are present. The Ellis Group seemingly becomes thicker northward at about the same

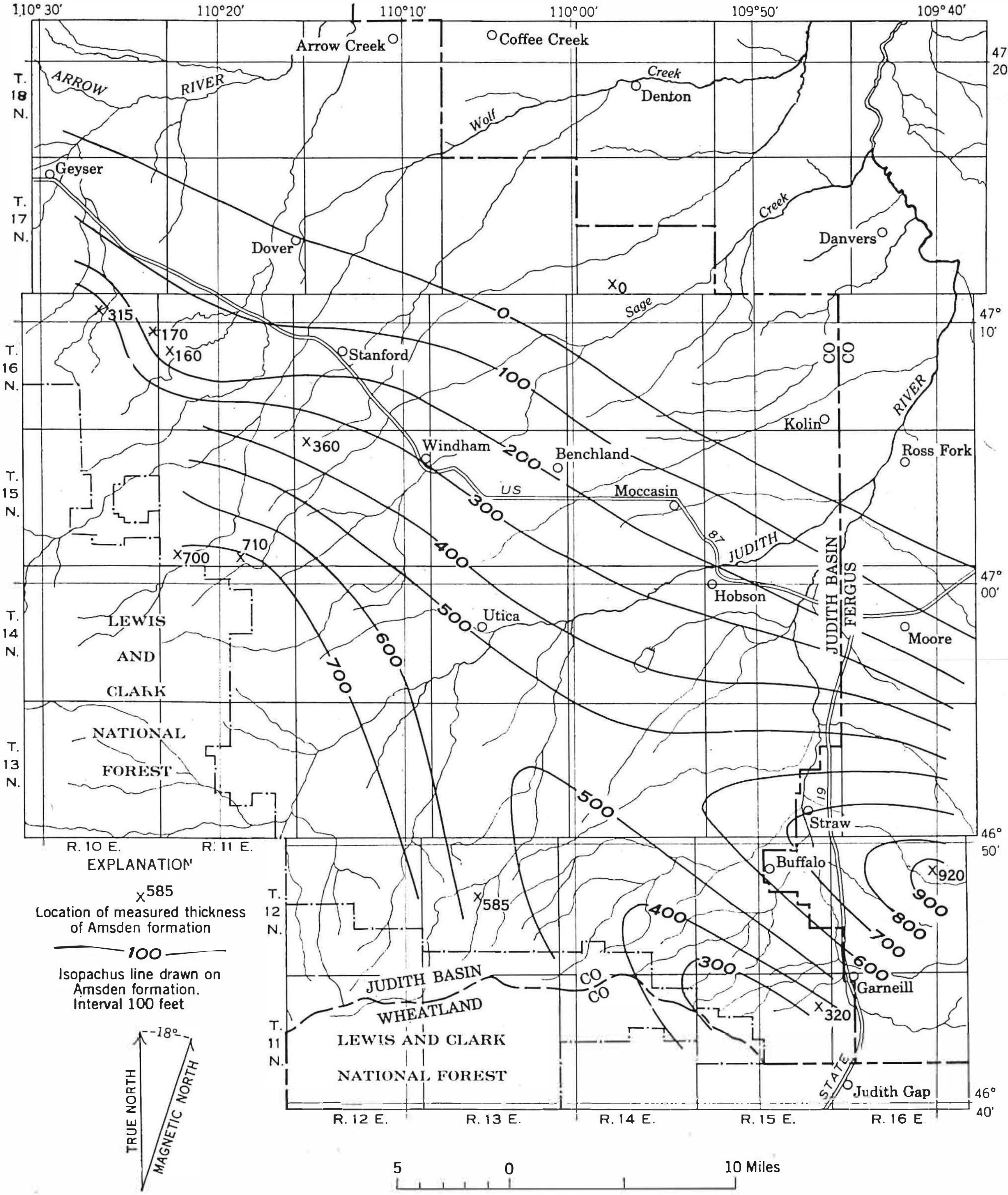

Figure 9.—Thickness map of Amsden Formation. (From Vine, 1956, p. 431.)

rate as the underlying Amsden Formation and Big Snowy Group become thinner, so that the combined thickness of these formations probably remains about 1,400 feet (Vine and Hail, 1950). The Ellis Group is shown as a single unit on the geologic maps (Pl. 3 and 4), but the formations composing the group are discussed separately in the section that follows.

Piper Formation.—The Piper Formation, of Middle Jurassic age, consists of red to varicolored shale and siltstone and thin limestone and gypsum beds. The formation is as much as 200 feet thick. Gypsum is mined from the Piper Formation at Heath and Hanover in the eastern part of the Judith Basin. The Piper is not a source of ground water in the study area.

Rierdon Formation.—The Rierdon Formation, of Late Jurassic age, consists principally of gray calcareous shale. The formation is commonly missing in outcrops in the study area but is as thick as 200 feet in outcrops around the South Moccasin Mountains (Miller, 1959, p. 16). The Rierdon is not a source of ground water in the study area.

Swift Formation.—The Swift Formation, of Late Jurassic age, ranges from a few feet to 165 feet in thickness, and overlies the Rierdon unconformably. The Swift consists principally of brown to orange medium- to coarse-grained glauconitic calcareous, ferruginous sandstone. Interbedded shale and sandstone compose a transitional zone at the base of the formation but are not present in all outcrops. Fossils, especially oyster shells, are numerous at some exposures.

The Swift is the source of water for many springs and several flowing wells in the foothills of the mountains but, because of its depth, it is poorly tested in the central part of the basin. Owing to a greater amount of interstitial cementing and thus a lower permeability, wells tapping the Swift probably will yield less water per unit of head than the more friable sandstone beds of other formations in the study area.

MORRISON FORMATION

The Morrison Formation, of Late Jurassic age, is composed of 50 to 300 feet of extremely lenticular continental beds that consist of variegated shale and siltstone, thin nodular limestone, white and brown sandstone, black shale, and a bed of coal or carbonaceous shale near the top. Dinosaur bones, gastroliths, and fresh-water clam shells are common in the Morrison, but the softness of the beds makes exposure rare. A coal bed near the top of the Morrison formerly supported an extensive mining industry in the study area, and is useful as a marker bed. Owing to an erosional unconformity between the Morrison and the overlying Kootenai Formation, the coal bed is locally absent.

The lenticularity of the beds composing the Morrison Formation makes it impractical to predict the sequence of beds that will be penetrated by a drill hole or whether the hole will produce water. Some abandoned coal mines in the study area discharge water. Some of the water may come from the Morrison coal but much of it is probably from sandstone of the overlying Kootenai Formation. The mine shaft at the site of the defunct town of Lehigh flows 50 gpm of water and is designated as well 15-12-21db in table 3 (Part B) of this report. Water pumped from this mine once was part of the municipal supply for Lehigh. Generally, the Morrison is not a good aquifer.

KOOTENAI FORMATION

The Kootenai Formation, of Early Cretaceous age, averages between 400 and 500 feet in thickness and is characterized by red shale and brown to gray sandstone. A few thin nodular fresh-water limestone beds also are present. A massive crossbedded gray to brown basal sandstone bed about 100 feet thick is one of the most prominently outcropping beds in the study area. This bed is known by drillers as the Third Cat Creek sand in the Cat Creek oil field of Petroleum County, Montana, and as the Cutbank sand in the Kevin-Sunburst field in Toole County, Montana.

The characteristic red shale of the Kootenai Formation is a good marker bed for drillers in the Judith Basin, contrasting sharply with the drab gray shale of the overlying Colorado Shale. The formation is one of the best aquifers in central Montana. The basal sandstone is the best water-bearing bed, but several stratigraphically higher sandstone beds also commonly yield water. Springs are numerous in the outcrop area of the Kootenai.

Many water wells have been drilled into the Kootenai in the study area and most of them have been successful as stock or domestic wells. Most of them were drilled at sites structurally favorable for artesian flow. The town of Stanford obtains part of its municipal supply from well 16-12-17da2 drilled into the Kootenai Formation at a depth of 1,050 feet. Well 17-10-6dd supplies water to the town of Geyser. The well, owned by Judith Basin County and administered by the Geyser Artesian Water Association, is 500 feet deep and finished in the Kootenai. Well 14-16-18ab is 1,375 feet deep and flowed a reported 60 gpm under a static head of about 254 feet above ground surface. This well supplies a cafe, bar, motel, service station, and trailer court at an establishment called Eddy's Corner.

The Utica Women's Club had well 14-13-16cc drilled into the Kootenai at a depth of 515 feet. The well flows about 15 gpm of water and the static head was 101 feet above land surface on October 17, 1963. Logs of the above wells are given in Table 2 in Part B of this report.

The static head of the water in several flowing wells producing from the Kootenai was measured during the fall of 1963. The results were used to draw the generalized contours on the piezometric surface shown on Plate 1. The contour map may be used to predict the likelihood that a well drilled into the Kootenai at a given site will flow. If the altitude of the site is lower than that of the piezometric surface, as indicated by the contours, the well should flow.

Five flow-recovery tests were made on wells in the Kootenai. From these tests the coefficient of transmissibility of the Kootenai near the well site was computed. These data are summarized in Table 3.

Table 3.—Summary of flow-recovery test data for wells in the Kootenai Formation.

Well number	Rate of flow (gpm at well head)	Length of test (minutes)	Maximum head (feet above land surface)	Coefficient af trans-missibility (gpd/foot)
15-12-23ba2	7.8	155	70.75	170
16-11-36ba	1.2	96	13.94	1,860
16-12-35ac	5.5	100	58.77	574
17-10-16bb1	3.3	35	93.05	870
17-10-33aa	4.5	60	20.71	1,380

The results of the tests are far from consistent. This may be the result of differences in well construction, penetration of the aquifer, or the aquifer itself. The Kootenai is of continental origin and, as is common in continental beds, is somewhat lenticular. Thus, the aquifer properties may differ sharply from one locality to another. None of the tests was long enough to test a very large part of the aquifer.

Selection of any one of the coefficients of transmissibility given in Table 3 or an average of these coefficients as typical of the formation is probably invalid. If the median figure (870 gpd/ft) is used, however, it can be estimated that about 8,700,000 gpd of water flows northeastward through the Kootenai in a 20-mile section of the formation measured along the 4,300-foot piezometric contour (Pl. 1) where the gradient is approximately 50 feet per mile.

COLORADO SHALE AND TELEGRAPH CREEK FORMATION

The Colorado Shale and Telegraph Creek Formation are poorly exposed in much of western and southern parts of the Judith Basin, and thus are shown as a unit on the geologic maps (Pl. 3 and 4). The character and water-bearing properties of the two formations are well enough known that they are discussed separately in the following sections.

Colorado Shale.—The Colorado Shale, of Early and Late Cretaceous age, comprises approximately 1,500 feet of fissile gray or black marine shale, thin local sandstone beds, many concretionary zones, and bentonite beds. The beds are generally poorly exposed, owing to deep weathering of the nonresistant shale and extensive mantling with alluvial deposits of Quaternary age.

Except for a few thin sandstone beds, the Colorado Shale is a poor aquifer. Because it underlies a large part of the area, many wells have been dug or drilled into the formation. Small amounts of water are obtained from weathered shale but wells in this material are usually undependable.

A sandstone bed at the base of the Colorado Shale, called the First Cat Creek sand by local drillers, can yield adequate amounts of water for domestic and stock use. A sandstone bed about 20 feet thick and about 300 feet above the base of the Colorado yields water to several flowing wells in the vicinity of Stanford. Local drillers call this bed the Muddy sandstone. It was not ascertained whether the bed is actually correlative with the Muddy Sandstone member of the Thermopolis Shale of southern Montana and Wyoming.

A flow-recovery test was made on well 16-12-24dd, which flows about 13 gpm of water from the Muddy sandstone of drillers. The coefficient of transmissibility determined in this test is 390 gpd per foot. The maximum head developed during the 120-minute test was 41.59 feet above land surface.

The artesian pressure of water in some beds of the Colorado Shale and the Kootenai Formation is sufficient that water has broken through the confining shale beds in several places in the western Judith Basin. Artesian springs have resulted. The water issuing from these springs is so admixed with mud derived from the confining shale or bentonite beds that a thick slurry is discharged (oozed). As the slurry is exposed to the air, part of the water is evaporated and the slurry becomes too thick to flow readily. The result is that a cone 1 or 2 feet above the land surface is formed. The springs (which discharge very little water) are called mud geysers by local residents, and the town of Geyser was named for them.

Personnel of the U. S. Geological Survey attempted to stimulate the flow of one of the "mud geysers" (17-10-1ab1) by pumping during the summer of 1962. The centrifugal pump used proved inadequate for handling the heavy slurry. Attempts to thin the slurry by adding water were hampered by lack of a nearby water supply. After much effort, the level of the bentonite slurry was lowered about 3 feet. The slurry required about 1 month to recover to its former level.

Telegraph Creek Formation.—The Telegraph Creek Formation, of Late Cretaceous age, is composed of 160 feet of yellow-weathering dark-gray to dark-brown shale and sandy shale beds. The formation is transitional between the underlying Colorado Shale and the overlying Eagle Sandstone. Like the Colorado Shale, the Telegraph Creek is poorly exposed in the study area and has therefore been mapped with the Colorado Shale, although faunal studies indicate that it is more nearly associated with the overlying Eagle Sandstone.

The Telegraph Creek Formation is a poor aquifer. As it crops out in only a small part of the area, it has been little tested, but the character of the rocks indicates that it would yield little water to wells.

EAGLE SANDSTONE

The Eagle Sandstone, of Late Cretaceous age, comprises about 250 feet of massive friable brown and white sandstone beds and interbedded gray shale and thin coal beds. The sandstone beds form cliffs at some places and, where so exposed, are commonly weathered into grotesque forms. Generally, however, the Eagle Sandstone is poorly exposed. The formation crops out only in a small area in the southern part of the Judith Basin, around the margin of the Highwood Mountains, and in small patches in the northern part of the area.

The Eagle Sandstone is a fair aquifer in the small part of the southern Judith Basin underlain by it. Near the Highwood Mountains the Eagle is untested. It underlies generally steep slopes where no wells have been attempted. Igneous rocks have so intruded it near the Highwood Mountains that its ability to transmit water has probably been greatly reduced.

The town of Judith Gap (south of the study area) obtains its municipal water supply from the Eagle. Well 11-16-20cd penetrated 225 feet of the Eagle, and yielded 70 gpm of water with 28 feet of drawdown (a specific capacity of 2.5 gpm per foot).

CLAGGETT SHALE

The Claggett Shale, of Late Cretaceous age, about 500 feet thick, is present in the southeastern part of the study area. It consists principally of gray and brown shale, which is easily eroded. Hence, broad valleys are commonly developed in the Claggett outcrop. Because the shale weathers deeply, exposures are poor. In the vicinity of Harlowton, about 15 miles south of the study area, the Claggett contains some lenticular beds of sandstone that yield water. Because of poor exposures it was not possible to determine whether the sandstone is present in the Claggett within the study area. Except for the possible presence of sandstone lenses, the Claggett is not a likely source of water.

JUDITH RIVER FORMATION

Only the lower part of the Judith River Formation, of Late Cretaceous age, is present in the study area. It is poorly exposed, and its thickness could not be determined. The formation consists of lenticular soft, friable gray sandstone beds, gray shale or claystone, and some thin coal beds. Sandstone beds of the Judith River Formation yield enough water for domestic and stock needs. Because of the small areal extent of the outcrop in the study area, few wells were found that obtain water from the Judith River Formation.

INTRUSIVE ROCKS

Two general types of igneous rocks have intruded the sedimentary rocks in the Judith Basin. The first type is made up of dark, deeply weathering rocks characterized by an abundance of ferromagnesian minerals (augite, biotite, olivine) and a high proportion of potash feldspar. W. T. Pecora (*In* Vine, 1956, p. 454) classed them as shonkinite and mafic alkalic monzonite. These rocks generally form dikes and sills in the sedimentary rocks. The dikes and sills are especially abundant near the Highwood Mountains. The Heath Shale, Otter Formation, and Colorado Shale seem to be especially subject to intrusion by these rocks. The dark igneous rocks are shown on the geologic maps (Pl. 3 and 4) as "other intrusive rocks" with no attempted differentiation.

The second general type of intrusive rocks in the Judith Basin is made up of light-colored quartz-bearing fine-grained porphyritic rocks (Pecora *in* Vine, 1956, p. 454). These rocks generally form massive laccoliths or stocks. Most of these mapped in the study area are along the mountain front. They are shown on the geologic maps as quartz monzonite porphyry.

The data do not permit determination of the time of emplacement of these rocks, but it was probably contemporaneous with the post-Cretaceous emplacement of many other intrusive bodies in central Montana.

The effect of these igneous rocks on ground-water conditions in the Judith Basin is difficult to assess. The numerous dikes near the Highwood Mountains probably impede the flow of ground water and account for the location of some springs. The baking effect of their emplacement altered some of the sedimentary rocks. In test hole 18-8-12ab the permeability of some of the shale seems to have been increased by this baking, which caused it to be more brittle and to form cracks. The cracks transmit water more freely than the shale would normally.

Springs are fairly abundant around the flanks of the quartz monzonite porphyry bodies. Water is probably transmitted readily through the pronounced cracks produced by weathering and by expansion and contraction in the rocks. These cracks probably do not extend very deep, however, so the ability of the rocks to store water and discharge it over a long period is small. Local ranchers report that springs around the intrusive bodies usually are not reliable.

An intrusive body of quartz monzonite porphyry along Running Wolf Creek probably forms a ground-water dam that causes water that would otherwise be lost into the Madison Group to issue as a spring. The spring (14-11-6cb) is the principal contributor to the perennial part of the flow of Running Wolf Creek.

EXTRUSIVE ROCKS

Volcanic rocks compose much of the Highwood Mountains, but they were not studied in this investigation. According to Larsen (1941, p. 1733) they are of two types—quartz latite and mafic phonolite. On the geologic maps accompanying this report the two types are undifferentiated.

As the volcanic rocks generally underlie steeply sloping, uninhabited land, their hydrologic importance to the study area is small. Many springs issue from the flanks of the Highwoods, and much of the water discharged is probably derived from volcanic rocks, but the topography generally prohibits drilling in places underlain by these rocks.

TERRACE DEPOSITS

Terrace gravel deposits, probably Pleistocene or younger in age, mantle much of the Judith Basin. The gravel consists chiefly of fragments of limestone and includes lesser amounts of sandstone in most of the study areas. Locally, particularly near the Highwood Mountains, fragments of igneous rocks compose much of the gravel. The gravel is generally coarser near the source of the rocks, in the mountains, than out in the basin. Most of the gravel is less than 6 inches in diameter, though boulders as large as 2 feet in diameter can be found near the mountains. The degree of rounding increases away from the mountains.

The gravel deposits underlie a series of terrace levels. Following the usage of Vine (1956), these terrace levels are designated on the geologic maps by numbers on the basis of their relative altitude. The highest and presumably the oldest known terrace is designated number 1 and progressively lower terraces are designated by larger numbers. Local variations of a terrace surface, some of which merge with the main surface, are designated by a plus (+) or minus (−) sign following the terrace number, according to whether they are higher or lower than the main surface.

The gravel deposits range in thickness from zero to as much as 100 feet but generally are no thicker than 50 feet. A thin gravelly loam soil is developed on most terraces. The soil tends to be thicker on lower, younger surfaces. This fact suggests that the soil is weathered from flood-plain deposits of sand and silt that covered the gravel deposits, rather than being weathered from the gravel itself. The finer material has been partly eroded away from the older terraces, leaving a stony soil.

The similarity of the terrace gravel deposits to modern alluvial gravel deposits suggests that the conditions of deposition were similar. It is not clear, however, whether the terraces are dissected remnants of much broader surfaces that once covered the entire basin or represent former valleys between bedrock hills that have since been eroded away (inverted topography). The mechanism causing the alternate deposition and erosion that formed the steplike terraces is also somewhat obscure. Pluvial conditions during glacial advances probably produced the torrential streams necessary to erode and transport the gravel; and periodic changes in erosional base level, due to disruptions of the course of the Missouri River by glacial ice, caused the alternating deposition by streams in this region.

Alden (1932, p. 13) correlated the highest terrace deposit (number 1) with the Tertiary (Miocene or Pliocene) Flaxville Formation of northeastern Montana. As this correlation is based only on physiographic relations, however, and because no evidence of age was found for this or any terrace, all terraces in the study area have been designated as Quaternary in age.

The terrace gravel deposits are of much economic importance in the study area. The soils developed on the terraces are among the best suited for agriculture in the basin. The surfaces are smooth enough to facilitate the use of heavy equipment and to make irrigation feasible, although only a small area on the terrace near Stanford and about 1,600 acres northeast of Ackley Lake are now irrigated (1964). The gravel underlying the terraces is generally permeable enough to provide good drainage of the overlying soil. The terrace gravel is partly saturated with water, and it yields water to many wells. The saturated thickness is not great, but many farms and ranches obtain enough water for domestic and stock use from wells or springs in the gravel. Moore obtains part of its municipal supply from a tunnel at the base of the gravel underlying the number 2 terrace deposit. This source yields approximately 50 gpm. Stanford pumps about 13 gpm from a well in gravel underlying the number 3 terrace. Most residents of Hobson obtain their domestic water from wells in gravel under the number 2 or 3 terrace deposits. Several excavations made to the base of the gravel for construction have required pumping 20 gpm or more to keep the gravel dewatered. Test hole 15-13-26aa (about 5 miles out of the study area) produced 70 gpm with about 4 feet of drawdown from about 8 feet of saturated gravel underlying number 2 terrace deposit.

Many contact-type springs issue from the base of the terrace gravel. Springs are especially common along the edge of terrace deposit number 2 bordering the south side of the Judith River Valley. Several businesses in Hobson obtain water from a hillside spring issuing from gravel underlying terrace deposit number 2. Excavation and street grading have been hampered in Hobson by shallow water uncovered during construction. On terrace deposit number 3 west of State Highway 19 are many marshy areas resulting from ground water being discharged at the surface because of thinning of the gravel.

The yield of wells in the terrace gravel deposits is restricted by lack of recharge and reduced permeability caused by cementation of the gravel. Some of the highest well yields are in the vicinity of Hobson where irrigation water from Ackley Lake recharges the gravel. The gravel is so well cemented that excavations and gullies in the deposits stand with vertical walls in many places. The base of many terrace gravels is cemented into a hard conglomerate, and a cemented zone 1 or 2 feet thick lies near the top of many of the deposits.

At Hobson, well diggers report a cemented zone at a depth of about 9 feet. This zone, which may represent the pre-irrigation water table, limits the depth of most dug wells. Ample water for domestic use is usually obtainable from gravel overlying the cemented zone, but contamination of ground water by a leaking gasoline storage tank forced some residents to have their wells drilled deeper. The cementing material, as well as much of the gravel, is calcareous, so acidizing as a means of developing wells may be useful. In this method, a strong acid (commonly hydrochloric) is used to dissolve part of the calcareous material that impedes the flow of water to the well.

Generally, the gravel deposits underlying the lower and most extensive terraces are favorable as a source of ground water. Best results can be expected near the middle of the terraces. Along the terrace edges the drawdown of the water table caused by discharge of springs is appreciable, and the gravel may be nearly dewatered. Even in the middle of the terraces no very large yields to wells can be expected, but quantities adequate for ordinary stock and domestic needs can be obtained.

ALLUVIUM

The streams of the western and southern parts of the Judith Basin flow through valleys partly filled by Recent alluvium. The alluvium shown on the geologic maps (Pl. 3 and 4) is restricted to areas of mappable size underlain by alluvium, but most stream valleys contain some alluvium. Thus, many wells shown in the table of wells and springs (table 3, Part B) produce water from alluvial deposits too small to show on the map.

The alluvium consists of unconsolidated sand, gravel, silt, and clay. The relative proportions of the constituents and the degree to which they are sorted differ considerably, but the alluvium of most major streams in the study area contains a fairly large proportion of gravel. That in the valley of the Ross Fork of the Judith River contains much slay, which is probably derived from tributaries draining land where the Colorado Shale is exposed. Generally, the coarseness of alluvial material decreases with increased distance from the mountains.

Like terrace gravel, the alluvium is a fairly reliable aquifer. It underlies a relatively small part of the area, but many shallow wells dug or (rarely) drilled into it produce adequate stock and domestic water supplies.

Information is scant as to the thickness of the alluvium. Shallow wells ordinarily suffice for domestic or stock water supplies, hence few well logs record the thickness of the alluvinm. Logs of test holes reveal a thickness of 36 feet for the alluvium in Wolf Creek valley at site 16-11-23dd1 and 15 feet at site 17-12-16da. Alluvium 32 feet thick was found at site 17-10-16db in the valley of Lone Tree Creek.

Torrential flows of water occasionally descend some of the streams and sink into alluvium in the lower reaches. The loss of water reduces the ability of the stream to transport material and causes deposition. This results in braided stream channels and wide valleys as the water leaves choked channels and creates new ones. This phenomenon is most noticeable in Wolf Creek and Dry Wolf Creek valleys. North of Stanford, Wolf Creek valley widens to as much as 3 miles. The usually dry channel is higher than much of the alluvial surface. Along the north edge of the valley the surface is so much lower than the channel that a small stream, Meadow Creek, originates from seepage from the alluvium.

Located as it is along stream courses, the alluvium is better situated to receive recharge than are the terrace gravel deposits. Along the Judith River and Wolf and Running Wolf Creeks, considerable alluvial land is irrigated by surface water, thus further recharging the underlying alluvium. The water table is within a few feet of the surface in many places; and in some places, the soil is moist during most of the year. Wet soil makes the use of machinery problematical at times, but the land is suitable for pasturing livestock. The good quality of the ground water and the permeability of the alluvium effectively prevent the accumulation of soil-damaging "alkali". Much land in these valleys is occupied by wild phreatophytes—cottonwood trees, willows, wild roses, chokecherries, and others.

Two aquifer tests were made using wells in the alluvium of the Wolf Creek valley. The test with well 16-11-35bd2 indicated a coefficient of transmissibility of 30,000 gallons per day per foot for the alluvium at this site. If the width of the alluvial deposits is taken as a quarter of a mile and the gradient as 85 feet per mile (measured between the tested well and well 16-12-7da2) an underflow of about 440 gpm is indicated. The test with well 17-of 4,600 gallons per day per foot. The decrease in 12-10ad1 indicated a coefficient of transmissibility of 4,600 gallons per day per foot. The decrease in transmissibility at this site is probably caused by a greater proportion of fine material in the alluvium and by thinning of the deposits.

Changes in the transmissibility of the alluvium in a stream valley due to changes in the thickness, width, or composition of the underlying alluvium can affect the behavior of the streams where they are hydraulically connected to the alluvium. Thus, Running Wolf Creek is a perennial stream from its emergence from the mountains to about sec. 15, T. 15 N., R. 11 E. There the valley widens and the cross sectional area of the alluvium increases. The water sinks into the alluvium at low creek stages and emerges from springs in sec. 2, T. 15 N., R. 11 E. From the springs the creek is perennial for about 7 miles, but there it again sinks into the alluvium in a widened stream valley.

Several ranchers in the Wolf Creek valley have expressed an interest in the use of ground water for irrigation. The surface water supply has been overappropriated, and in dry years shortages result. Seemingly, ground water is available in sufficient quantities from alluvium in the upper Wolf Creek valley for small-scale irrigation.

Some development of fairly large amounts of ground water from the alluvium in the Judith River valley seems possible, but there is little inducement at present (1964). Domestic and stock needs are met by small-capacity wells, and irrigation water is economically supplied by diversion of surface water.

COLLUVIUM AND LANDSLIDE DEPOSITS

Colluvium and landslide deposits occur in many places in the study area, but only where landslide deposits obscure sizable parts of the underlying rocks are they mapped. In the table of wells and springs (Table 3, Part B), colluvium is listed as the geologic source of water for some wells and springs. Colluvium is the nearly ubiquitous detritus resulting from small landslides, soil creep, and other processes. It ranges considerably in thickness, composition, and hydrologic properties. It is not an important aquifer in the study area.

CHEMICAL CHARACTER OF THE GROUND WATER

The extent to which water resources can be used depends on the quality as well as the quantity of the available water. As part of this study, efforts were made to determine the chemical quality of the water and to evaluate the suitability of the water for domestic, agricultural, and industrial uses. Twenty-five samples of water from various wells, springs, and streams were collected during the fall of 1963 and analyzed by the Montana State Board of Health. In addition to these, the Board of Health and owners provided eight additional analyses of water collected in past years. Yapuncich, Sanderson, and Brown Laboratories, of Billings, Montana, provided three analyses from their files. The analyses are summarized in Table 4. The location of sampling points is shown on Plates 1 and 2. The data pertain only to the dissolved mineral substances in the water and not to the sanitary condition of the water.

CHEMICAL CONSTITUENTS IN RELATION TO USE

DISSOLVED SOLIDS

The total mineralization of water is represented by the concentration of dissolved solids in ppm (parts per million). This concentration may be determined by evaporating the sample to dryness, or calculated by adding together the concentrations of the component ions. Analyses shown in Table 4 list calculated dissolved solids concentrations.

Generally, water with dissolved solids concentrations of less than 500 ppm is suitable for domestic use. Water with concentrations of more than 1,000 ppm is likely to contain enough of certain constituents to cause noticeable taste or otherwise make the water undesirable or unsuitable for use.

Residents of places where the dissolved solids concentration of drinking water exceeds these limits often develop a tolerance and, perhaps, a taste for the water. No apparent ill effects are generally noted. Water with dissolved solids concentrations exceeding 3,000 ppm is rarely acceptable as drinking water.

The dissolved solids concentration of the water samples collected in the western and southern parts of the Judith Basin ranged from 92 to 3,359 ppm. Of these, 22 had concentrations of 500 ppm or less, 9 had between 501 and 1,000 ppm, and 8 had concentrations ranging from 1,080 to 3,359 ppm.

SPECIFIC CONDUCTANCE

A portable conductance cell was used during this investigation to gather semiquantitative data on water quality. The cell measures the specific conductance of the water—that is, its ability to conduct electricity. The electric conductance of water varies with the concentration of dissolved solids in the water; thus, a high specific conductance indicates a high dissolved solids concentration. Though the specific conductance gives an idea of the concentration of dissolved solids it does not indicate what specific constituents may be present. For most water samples, the total dissolved solids, in parts per million, is numerically equal to 55 to 75 percent of the specific conductance in micromhos per centimeter at 25°C. Thus, water having a specific conductance of 1,000 micromhos could be expected to have a concentration of 550 to 750 parts per million total dissolved solids. Specific conductance was determined for water from several hundred wells and springs in the study area. The results are tabulated in Table 4 in Part B of this report. Specific conductance values determined in the study area ranged from 92 to 8,000 micromhos or more per centimeter at 25°C.

HARDNESS

The soap-consuming property of water is known as hardness. It may be caused by several different cations, present in variable proportions. It is a property of water and not a chemical constituent, but it is generally expressed as the amount of $CaCO_3$ in ppm required to produce the soap-consuming effect.

Classifying water as "hard" or "soft" is generally unsatisfactory because the previous experience of users will affect their judgment. Thus, in some places water having a hardness of 100 ppm would be regarded as "hard", and in other places such water would be regarded as "soft".

Hardness of water to be used for ordinary domestic purposes does not become particularly objectionable below the level of 100 ppm (Hem, 1959, p. 147). Above this level, hardness becomes increasingly troublesome as the concentration increases.

The hardness of 34 water samples was determined as part of this study (Table 4). Of these samples, 6 had a hardness of less than 100 ppm; 23 had hardness ranging from 110 to 693 ppm; and 5 had a hardness exceeding 1,000 ppm.

IRON

Ordinarily, iron occurs in ground water in low concentrations. It remains in solution and the water is clear until exposed to the oxygen in the air, whereupon the iron is oxidized to the ferric state and precipitated as the hydroxide ($Fe(OH)_3$) or oxide (Fe_2O_3). This precipitate causes brownish or reddish stains on porcelain fixtures, laundry, and other materials with which the water comes in contact.

The U. S. Public Health Service standard for water used on interstate carriers calls for iron concentrations of less than 0.3 part per million. Water

Table 4.—Chemical analyses of water from geologic units, rivers, and streams in Judith Basin.

[All analyses by Montana State Board of Health unless otherwise indicated. Figures are in parts per million.]

Location	Date of collection	Depth of well (feet)	Temperature (°F)	Silica (SiO_2)	Iron (Fe)	Calcium (Ca)	Magnesium (Mg)	Sodium and potassium (Na+K)	Bicarbonate (HCO_3)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F)	Nitrate (NO_3)	Dissolved solids	Hardness as $CaCO_3$	Conductivity (Micromhos per cm @ 25°C)
PRE-MISSISSIPPIAN ROCKS																
12-14-19bd*	----	2,809-2,866	----	----	----	124	60	----	230	333	15	----	----	645	----	----
MADISON GROUP																
13-11-14dd	11-13-63	140	44	----	0.14	37	22	0	195	10	4	0.1	8.1	200	182	350
13-11-35ab	11-13-63	----	45	----	.30	92	24	8	287	18	26	.1	62.0	372	330	610
15-10-21ad	11-12-63	Spring	44	----	.34	432	71	4	116	1,225	4	.4	0.0	1,884	1,370	1,700
16-15-30(?)*	8-12-60	4,472	----	----	----	650	180	100	145	2,339	19	----	----	3,359	----	----
OTTER FORMATION																
14-11-35bd	11-14-63	Spring	46	----	.14	40	11	2	152	21	1	.1	1.1	190	143	265
16-10-10bc	10- 1-63	Spring	45	----	.12	114	47	0	110	352	1	.6	1.6	576	479	840
AMSDEN FORMATION																
12-13-24ab*	----	----	----	----	----	509	267	53	110	2,288	10	----	----	3,181	2,369	----
15-11-27bb	11-12-63	Spring	49	----	1.38	46	50	0	232	108	4	.1	1.1	356	308	----
15-12-22cd	10- 2-63	650	48	----	.10	73	41	11	238	165	3	.4	0	390	352	710
16-11-18cc	10- 1-63	Spring	46	----	.12	275	107	110	189	1,160	4	1.0	0	1,906	1,128	1,780
ELLIS GROUP																
14-13-30bc	11-18-63	----	47	----	.10	26	27	90	275	120	4	.2	0	372	176	725
MORRISON FORMATION																
15-12-21db	10- 2-63	250	46	----	5.40	150	25	4	177	318	5	.9	.4	636	473	775
KOOTENAI FORMATION																
14-15-20dc	1936	----	----	5.6	----	4.7	1.6	131	291	31	6	----	----	353	----	----
14-15-30ad*	10-12-38	1,870	----	----	----	144	70	23	230	475	11	----	----	736	----	----
15-12-1cb	10- 2-63	----	46	----	.10	31	19	115	280	145	4	.5	0	432	154	780
16-11-36ba	10- 1-63	405	46	----	.30	57	11	0	6	55	2	.5	.4	198	187	285
16-12-17da2	10-21-53	1,060	----	----	Trace	52	15	23	204	57	7	----	----	256	187	----
16-12-35cc1	7- 2-51	720	----	0	15	59.5	26	2	192	92	3.5	----	0	284	----	470
17-10-6dd	10- 1-63	500	55	----	0	123	35	20	119	378	2	.6	0	636	451	925
17-12-21cd	11-18-63	1,563	42	----	.14	4	0	282	433	200	10	1.2	.4	720	11	----
18-11-22cc	11- 8-63	1,134	52	----	3.88	35	13	20	198	10	3	.5	.7	148	143	----

COLORADO SHALE

12-16-19cd2	6-23-54	80	----	--------	0	69	39	44	316	139	13	.7	0	500	334	--------
13-16-2bc	7-21-32	215	----	14.6	--------	0.8	2.1	134.4	216.5	2.6	5	------	--------	341	11	--------
16-12-16bb2	10-23-62	170	----	--------	4.0	18	6	--------	--------	93	9.0	------	--------	388	67	1,200
16-12-16bd2	10-24-62	193	----	--------	.44	18	0	--------	--------	181	10	------	--------	468	44	--------
16-12-24dd	11-12-63	320	44	--------	.20	26	11	355	311	570	8	.3	.9	1,080	110	1,700
17-10-1ab1	10- 1-63	Spring	55	--------	0	9	0	615	573	680	50	4.8	.2	1,706	22	--------

EAGLE SANDSTONE

12-16-32cc1	5-19-50	71.6	----	23	2	364	131	468	464	1,958	24	.4	Trace	3,260	1,447	--------

EXTRUSIVE ROCKS

19-10-15da	10- 1-63	Spring	47	--------	.06	9	3	30	88	19	2	.4	0	92	33	--------

TERRACE DEPOSITS

13-12-18ac	10- 2-63	Spring	46	--------	.14	95	28	15	162	236	3	.2	.7	504	352	780
14-16-15cb	4-13-59	50	----	--------	--------	72	34	0	220	40	5	------	0	375	319	--------
16-12-16bd1	10-24-62	46.5	----	--------	.6	91	53	--------	--------	185	24	------	--------	624	444	--------
18-12-2cb	10- 2-63	32.6	----	--------	0	136	86	240	284	815	45	1.2	74.3	1,596	693	2,000
18-12-3dc	--------------	23	----	2	1	290	216	178	222	1,734	131	------	40	2,985	1,610	--------

ALLUVIUM

14-11-6cb	11-12-63	Spring	44	--------	1.90	40	13	0	155	8	1	0	.2	136	154	--------
15-12-12da	11-18-63	9.9	48	--------	.14	84	89	35	262	387	15	.8	6.9	826	583	1,100
17-12-10ad1	11- 8-63	9.2	48	--------	.54	73	45	45	317	148	27	.2	12.7	486	369	900

JUDITH RIVER, MIDDLE FORK

13-11-35bd	11-13-63	Stream	----	--------	.14	35	11	0	146	9	2	0	.4	140	132	--------

WOLF CREEK

16-11-35bd	11-12-63	Stream	40	--------	.44	57	19	4	207	52	3	0	.5	240	220	--------

*Analysis provided by Yapuncich, Sanderson, and Brown Laboratories having a greater concentration is not injurious to health but will cause staining.

Some of the samples collected during this study precipitated part of the iron in solution on being aerated during collection. Thus the analyses may show a lower-than-true concentration of iron. Concentrations, as reported, range from 0 to 15 ppm, and 13 samples had concentrations of iron in excess of the 0.3 ppm limit recommended.

Most iron-bearing water can be treated so that it is satisfactory for domestic use. This is usually accomplished by aeration and filtration. Aeration exposes the water to the oxygen of the air, and most of the iron is precipitated. The water is then passed through a filter, commonly sand or charcoal, where the precipitate is removed.

SULFATE

Sulfate is one of the major ions in solution in most of the samples analyzed. Sulfate, in combination with sodium and magnesium ions, may impart an objectionable taste to water and has a purgative effect on humans and livestock. This purgative effect can, where due to only moderate concentrations, be tolerated but may cause distress to those unaccustomed to the water.

The U. S. Public Health Service recommends a limit of 250 ppm as the maximum concentration of sulfate in drinking water. Of the samples analyzed, 25 had concentrations of sulfate within this limit.

FLUORIDE

Fluoride is present in ground water only in small quantities, but a knowledge of the fluoride content of water is important. The use of water containing a concentration of fluoride in excess of 1.5 ppm by children during the formation of permanent teeth may cause mottling of the tooth enamel. If water containing lesser concentrations of fluoride is used during this formative period, however, the teeth tend to become resistant to decay. Of 27 fluoride determinations shown on Table 4, one sample had a fluoride concentration of 4.8 ppm. The others had concentrations ranging from 0 to 1.2 ppm. The sample containing the concentration of 4.8 ppm is from a "mud geyser", which produces very little water and is not a source of drinking water for humans.

NITRATE

Large concentrations of nitrate in water may cause cyanosis in infants if the water is used for drinking and in preparing formulas for feeding. The tentative limit is about 44 ppm (Hem, 1959, p. 239). Infant cyanosis is not fatal in most cases if diagnosed in time but may be fatal with continued use of the water. In addition to this hazard, the presence of nitrate in any but small concentrations (less than 10 ppm) in company with higher-than-usual concentrations of chloride may indicate that the water is contaminated. The presence of nitrate does not prove contamination, but it does suggest it.

Of 30 analyses of water samples from the study area listed on Table 4, only 6 had concentrations of more than 2 ppm nitrate. These 6 were analyses of water from the Madison Limestone, terrace deposits, and alluvium. Limestone, because of its comparatively large openings, is susceptible to contamination. The water in terrace deposits and alluvium generally is shallow and subject to contamination from barnyards, septic tanks, and privies.

DISSOLVED GASES

Many of the deeper wells and some of the springs in the study area yield water containing hydrogen sulfide (H_2S) gas or carbon dioxide (CO_2). Hydrogen sulfide has an offensive odor like that of rotten eggs. In great concentrations it is poisonous, but as a strong odor is imparted by less than 1 ppm of hydrogen sulfide in water, its odor usually suffices to warn of danger.

Carbon dioxide is an odorless gas, which has been reported in water from some of the deeper wells. It is generally harmless to people except in such strong concentrations that it displaces oxygen. It is commonly added to drinks to cause effervescence—as in soda pop.

Because the solubility of gases varies inversely with temperature and directly with pressure, these gases are most commonly associated with water from wells under considerable pressure head. The gases are rapidly evolved when the water is reduced to atmospheric pressure and warmed.

In solution, hydrogen sulfide and carbon dioxide gas form weak acids, which make the water corrosive. Reportedly, maintenance of Stanford's deep well (16-12-17da2) has been difficult and costly because of corrosion caused by carbon dioxide in the water. Several of the wells flowing under high static heads yield effervescent water. The odor presents an almost unmistakable clue to the presence of hydrogen sulfide, and its presence was noted during the inventory of many springs and wells.

Both carbon dioxide and hydrogen sulfide are products of the decay of organic matter. Appreciable amounts of carbon dioxide in water are noted mainly in water from the Kootenai Formation. Some of it may result from oxidation of some of the coal in the upper part of the Morrison Formation. Pyrite (iron sulfide) and other sulfides in the coal may be the source of part of the hydrogen sulfide. Hydrogen sulfide is also common in water from carbonaceous shale and from decaying modern organic matter.

The odor of hydrogen sulfide gas usually tends to make people regard water containing it as bad. The gas is easily removed by aeration and should therefore not present a serious deterrent to most uses of the water. The acidic condition it may give to the water and its action as a chemical reducing agent, however, make water containing hydrogen sulfide especially likely to contain objectionable amounts of iron and other dissolved solids.

Collection of samples for analysis of dissolved gases requires special techniques and equipment. For this reason, no analyses of dissolved gases are presented in this report.

CHEMICAL CONSTITUENTS IN RELATION TO IRRIGATION

The following discussion of the suitability of water for irrigation is adopted from Agriculture Handbook 60 of the U. S. Department of Agriculture (U. S. Salinity Laboratory Staff, 1954).

The development and maintenance of successful irrigation involve not only supplying irrigation water to the land but also control of salt and alkali in the soil. The quality of irrigation water, irrigation practices, and drainage conditions are involved in salinity and alkali control. Soil that was originally nonsaline and nonalkali may become unproductive if excessive soluble salts or exchangeable sodium are allowed to accumulate because of improper irrigation and soil-management practices or inadequate drainage.

In areas of sufficient rainfall and ideal soil conditions the soluble salts originally present in the soil or added to the soil with water are carried downward by the water and ultimately reach the water table. The process of solution and transportation of soluble salts by water moving through the soil is called leaching. If the amount of water applied to the soil is not more than the amount needed by plants, there will be no downward percolation of water below the root zone, and minerals will accumulate there. Impermeable soil zones can retard downward movement of water, resulting in waterlogging of the soil and deposition of salts. Unless drainage is adequate, leaching may fail, because leaching requires the free passage of water through and away from the root zone.

Four characteristics of water for irrigation seem to be most important in determining its quality. These characteristics are: (1) total concentration of soluble salts; (2) relative proportion of sodium to other principal cations (calcium, magnesium, and potassium); (3) concentration of boron or other elements that may be toxic to plants; and (4) under some conditions, the bicarbonate concentration as related to the concentration of calcium plus magnesium.

The specific conductance of water serves as a convenient expression of the total concentration of soluble salts for purposes of diagnosis and classification. It is easily measured by use of a portable conductance cell. In general, water having a specific conductance of less than 750 micromhos per centimeter is satisfactory for irrigation insofar as salt concentration is concerned. Crops differ greatly in their tolerance to salinity; thus, some crops may be adversely affected by irrigation water having a specific conductance in the range of 250 to 750 micromhos. Water having a specific conductance of 750 to 2,250 micromhos is widely used, but good management and favorable drainage conditions are necessary to prevent saline conditions.

Some criticism of the use of conductance as a criterion of water suitability has been voiced (Hem, 1959, p. 248) on the grounds that when the water contains fairly large concentrations of calcium and magnesium these may combine with carbonates to precipitate in harmless form in the soil. Sulfate may tend to precipitate as calcium sulfate.

The relative tolerance to salinity of crops that may be grown in central Montana is shown in Table 5.

When a soil containing exchangeable calcium (Ca^{++}) and magnesium (Mg^{++}) ions is irrigated with water in which sodium (Na^{+}) greatly outnumbers other cations, the calcium and magnesium of the soil will tend to be replaced with sodium, that is, the sodium is adsorbed. If the irrigation is long continued, the tilth and permeability of the soil may be impaired.

The U. S. Department of Agriculture (U. S. Salinity Lab. Staff, 1954) developed an empirical ratio to express the probable adsorption of sodium by soil to which a given water is added. This ratio, called the sodium-adsorption ratio or SAR, is determined

Table 5.—Relative tolerance of various crop plants to salinity

(Adopted from U. S. Dept. of Agriculture Handbook No. 60, 1954. In each column the first-named plants under each class are most sensitive and the last-named under that class the most tolerant.)

Sensitive	Moderately tolerant	Tolerant
	FRUIT CROPS	
Strawberry		
Plum		
Apple		
	VEGETABLE CROPS	
Green beans	Cucumber	Spinach
Celery	Squash	Asparagus
Radish	Peas	Kale
	Onion	Beets
	Carrot	
	Potatoes	
	Sweet corn	
	Lettuce	
	Cauliflower	
	Bell pepper	
	Cabbage	
	Broccoli	
	Tomato	
	FORAGE CROPS	
Ladino clover	Smooth brome	Barley (hay)
Red clover	Meadow fescue	Western wheat grass
Alsike clover	Blue gramma	Canada wild rye
	Orchard grass	Rescue grass
	Oats (hay)	Salt grass
	Wheat (hay)	Alkali sacaton
	Rye (hay)	
	Tall fescue	
	Alfalfa	
	Mountain brome	
	Perennial rye grass	
	Sweet clover	
	FIELD CROPS	
Field beans	Flax	Rape
	Corn	Sugar beet
	Oats	Barley
	Wheat	
	Rye	

by the following relation where ion concentrations are expressed in epm (equivalents per million):

$$\text{SAR} = \frac{\text{Na}^{+}}{\sqrt{\dfrac{\text{Ca}^{++} + \text{Mg}^{++}}{2}}}$$

The concentration of sodium (Na^{+}) in ppm can be converted to the concentration in epm by multiplying by 0.0435, the calcium (Ca^{++}) concentration in ppm can be multiplied by 0.0499, and the magnesium (Mg^{++}) in ppm can be multiplied by 0.0822.

The analyses with this report list combined sodium and potassium concentrations. The concentration of potassium is rarely very great in natural water, hence little error will be introduced if these combined concentrations are substituted for the sodium concentration.

As only the Madison Group, Amsden Formation, Kootenai Formation, and alluvium are regarded as potential sources of irrigation water in the study area, the SAR values for these formations only have been computed, and only samples for which the specific conductance is known are included. These SAR values are listed in Table 6.

Table 6.—Sodium-adsorption ratio (SAR) and specific conductance of selected samples of water

Well number	SAR	Specific conductance (micromhos per cm @ 25°C)
MADISON GROUP		
Well 13-11-14dd	0.0	350
Well 13-11-35ab	.2	610
Spring 15-10-21ad	.05	1,700
Oil test 16-15-30 ?	.9	3,000+
AMSDEN FORMATION		
Well 15-12-22cd	.3	710
Spring 16-11-18cc	1.4	1,780
KOOTENAI FORMATION		
Well 15-12-1cb	12.7	780
Well 16-11-36ba	.0	285
Well 16-12-35cc1	.05	470
Well 17-10-6dd	.4	925
ALLUVIUM		
Well 15-12-12da	.6	1,100
Well 17-12-10ad1	1.0	900

When the sodium-adsorption ratio and the specific conductance of a water are known, the classification of the water for irrigation can be determined by plotting these values on the diagram shown in Figure 10. Low-sodium water can be used for irrigation on almost all soils with little danger of developing harmful levels of exchangeable sodium. Medium-sodium water will present an appreciable sodium hazard in some fine-textured soils, especially poorly leached ones. Such water may be safely used on coarse-textured or organic soils having good permeability. High-sodium water may produce harmful levels of exchangeable sodium in most soils and will require special soil management such as good drainage and leaching and addition of organic matter. Very high sodium water is generally unsatisfactory for irrigation unless special action is taken, such as addition of gypsum to the soil.

Low-salinity water can be used for irrigation of most crops on most soils with little likelihood that soil salinity will develop. Medium-salinity water

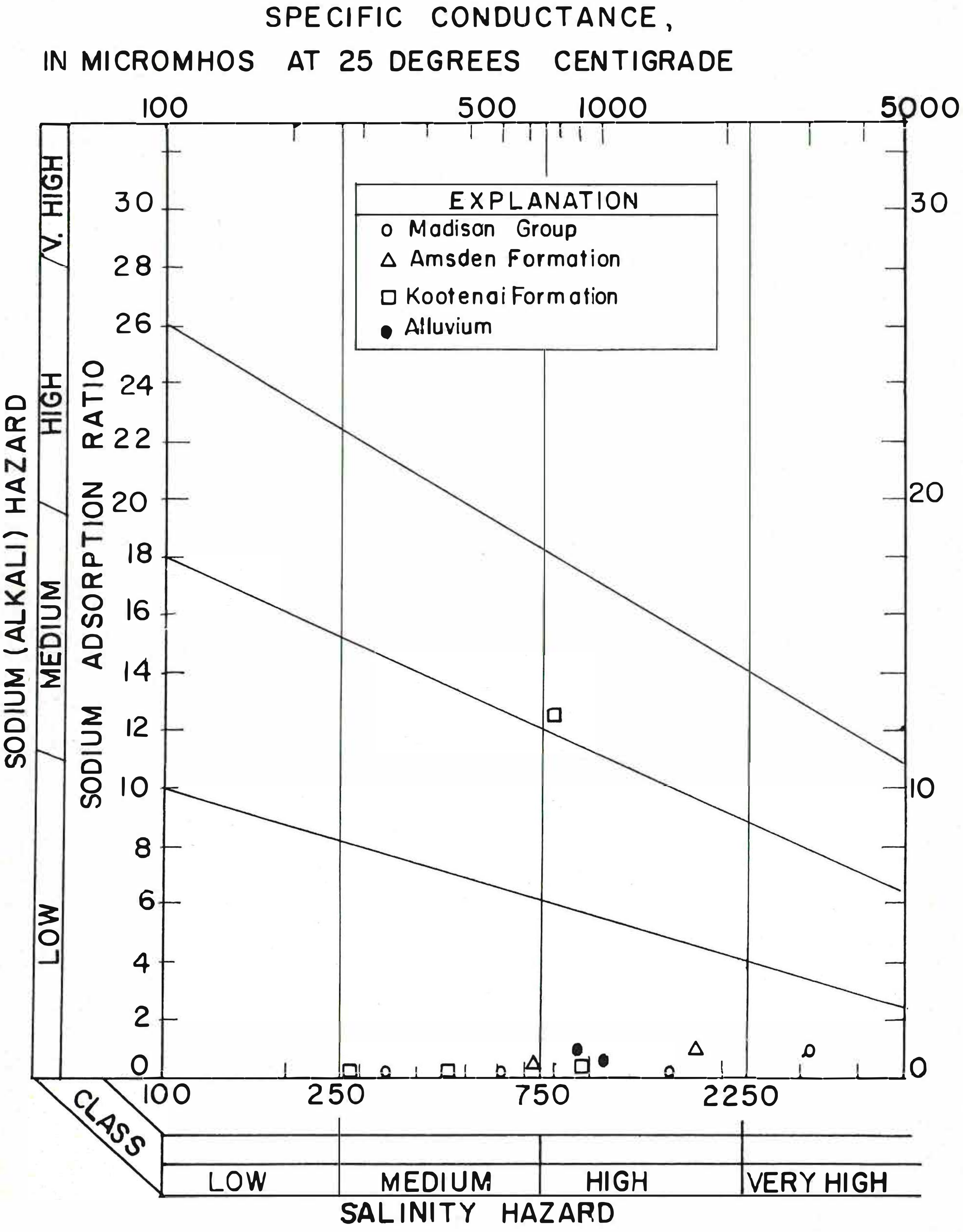

Figure 10.—Diagram showing suitability of water for irrigation.

can be used if a moderate amount of leaching occurs. Crops of moderate salt tolerance can be irrigated with medium-salinity water without special practices. High-salinity water should not be used on poorly drained soil. Very high salinity water is not ordinarily suitable for irrigation. It should be used only on crops that are very tolerant of salt and then only if special practices are followed, including leaching.

As may be noted on Figure 10, most of the water samples collected from aquifers that might be sources of irrigation water present a low sodium hazard. Many of them, however, present a high salinity hazard. Oil test 16-15-30 is located outside the study area but the analysis of water collected in a drill-stem test of the Madison Group is included as an indication of what water from great depths in the Madison may be like. This water presents a very high salinity hazard according to Figure 10. The predominant ions are calcium and sulfate. As these ions, in the form of gypsum, are sometimes applied to land as a soil conditioner, this water may not be harmful if applied as irrigation water.

Boron is essential to normal plant growth, but the quantity required is very small, and larger quantities are harmful. Boron was not determined in the analyses presented with this report.

In water having a high concentration of bicarbonate, calcium and magnesium tend to precipitate as the water in the soil becomes more concentrated as a result of evaporation and transpiration. The reaction does not ordinarily go to completion, but insofar as it does proceed, the concentration of calcium and magnesium is lowered, and the relative sodium concentration increases. The calcium and magnesium precipitate as carbonates, and any residual carbonate remains in solution as sodium carbonate. Thus, the SAR of the remaining solution is greatly increased.

Prediction of what may take place in the complex mineral assemblages of most soils on the basis of water analysis alone is very difficult. Conditions of the soil, drainage, rainfall, and many other factors should be considered as well.

CONCLUSIONS

Ground water is obtainable from the Madison Group, Kibbey Sandstone, Amsden Formation, Swift Formation of the Ellis Group, Kootenai Formation, some sandstone beds in the Colorado Shale, Eagle Sandstone, Judith River Formation, terrace deposits, and alluvium in the western and southern parts of the Judith Basin. Of these, the Kootenai Formation, terrace deposits, and alluvium now provide water to most wells. The Madison Group is almost untapped and its water-bearing potential is poorly known, but as it is known to accept a large volume of recharge from streams, it probably is capable of yielding a large amount of water. Only one well, in the Amsden Formation, had sufficient natural flow to irrigate a very large amount of land, but tests of aquifer properties indicate that alluvium in some stream valleys can yield enough water for tracts of as much as 40 acres.

Ground water in the western and southern parts of the Judith Basin is moderately to strongly mineralized but most of it is of good quality for stock and domestic use. Much of the water is hard. Water from the Madison, Amsden, and Kootenai Formations and the alluvium in Wolf Creek valley is of fair quality for irrigation. It generally presents a low alkalinity hazard but a medium to high salinity hazard.

SELECTED REFERENCES

ALDEN, W. C., 1932, Physiography and glacial geology of eastern Montana and adjacent areas: U. S. Geol. Survey Prof. Paper 174, 133 p., 51 pl.

CALVERT, W. R., 1909, Geology of the Lewistown coal field, Montana: U. S. Geeol. Survey Bull. 390, 83 p., 5 pl.

FERRIS, J. G., KNOWLES, D. B., BROWN, R. H., and STALLMAN, R. W., 1962, Theory of aquifer tests: U. S. Geol. Survey Water-Supply Paper 1536-E, p. 69-171, fig. 17-45.

FISHER, C. A., 1909, Geology of the Great Falls coal field, Montana: U. S. Geol. Survey Bull. 356, 85 p., 12 pl.

HEM, J. D., 1959, Study and interpretation of the chemical characteristics of natural water: U. S. Geol. Survey Water-Supply Paper 1473, 269 p., 2 pl., 40 fig.

LARSEN, E. S., 1941, Igneous rocks of the Highwood Mountains, pt. 2, The extrusive rocks: Geol. Soc. America Bull., v. 52, p. 1733-1752.

MEINZER, O. E., 1923a, The occurrence of ground water in the United States, with a disscussion of principles: U. S. Geol. Survey Water-Supply Paper 489, 321 p., 31 pl.

, 1923b, Outline of ground-water hydrology, with definitions: U. S. Geol. Survey Water-Supply Paper 494, 41 p.

, 1927, Large springs in the United States: U. S. Geol. Survey Water-Supply Paper 557, 94 p., 17 pl.

MILLER, R. N., 1959, Geology of the South Mocassin Mountains, Fergus County, Montana: Montana Bur. Mines and Geology Mem. 37, 44 p., maps.

MONTANA STATE ENGINEER, 1963, Water resources survey, Judith Basin County, Montana, Part I and II.

PERRY, E. S., 1932, Ground-water resources of Judith Basin, Montana: Montana Bur. Mines and Geology, Mem. 7, 30 p., maps.

REEVES, FRANK, 1929, Thrust faulting and oil possibilities in the plains adjacent to the Highwood Mountains, Montana: U. S. Geol. Survey Bull. 806-E, p. 155-195, pl. 44.

—, 1931, Geology of the Big Snowy Mountains, Montana: U. S. Geol. Survey Prof. Paper 165-D, p. 135-149, pl. 35-38.

THEIS, C. V., 1935, Relations between the lowering of the piezometric surface and the rate and duration of discharge of a well using ground-water storage: Am. Geophys. Union Trans., 16th Ann. Mtg., pt. 2, p. 519-524.

U. S. PUBLIC HEALTH SERVICE, 1962, Drinking water standards: U. S. Public Health Service Publication 956.

U. S. SALINITY LABORATORY STAFF, 1954, Diagnosis and improvement of saline and alkali soils: U. S. Dept. Agriculture, Agriculture Handb. 60.

VINE, J. D., 1956, Geology of the Stanford-Hobson area, central Montana: U. S. Geol. Survey Bull. 1027-J, p. 405-470, pl. 45-47, fig. 52-61.

, and HAIL, W. J., JR., 1950, Geologic map of the Hobson area, central Montana: U. S. Geol. Survey Oil and Gas Inv. Prelim. Map 108.

——— , and JOHNSON, W. D., JR., 1954, Geology of the Stanford area, Judith Basin and Fergus Counties, Montana: U. S. Geol. Survey Oil and Gas Inv. Map OM 139.

WALTON, P. T., 1946, Ellis, Amsden, and Big Snowy Group, Judith Basin, Montana: Am. Assoc. Petroleum Geologists Bull., v. 30, p. 1294-1304.

WEED, W. H., 1899a, Fort Benton, Montana: U. S. Geol. Survey Folio 55.

, 1899b, Little Belt Mountains, Montana: U. S. Geol. Survey Folio 56.

——, 1900, Geology of the Little Belt Mountains, Montana, with notes on the mineral deposits of the Neihart, Barker, Yogo, and other districts: U. S. Geol. Survey Ann. Rept. 20, pt. 3, p. 257-581, pl. 36-77.

WESTGATE, L. G., 1921, Deposits of iron ore near Stanford, Montana: U. S. Geol. Survey Bull. 715-F, p. 85-92.

ZIMMERMAN, E. A., 1962, Preliminary report on the geology and ground-water resources of southern Judith Basin, Montana: Montana Bur. Mines and Geology Bull. 32, 23 p., map.