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Moscow Basin Ground Water Studies

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MOSCOW BASIN GROUND WATER STUDIES

by

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Prepared in Cooperation with the Water Resources Research Institute, University of Idaho.

IDAHO BUREAU OF MINES AND GEOLOGY

MOSCOW, IDAHO

TABLE OF CONTENTS

TABLE OF CONTENTS (Continued)

TABLE OF CONTENTS (Continued)

LIST OF TABLES

LIST OF FIGURES

LIST OF FIGURES

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ABSTRACT

Moscow basin is in Latah County, Idaho, on the eastern edge of the Columbia Plateau physiographic province. The area of the basin is about 58 square miles. The principal water users, City of Moscow and University of Idaho, depend exclusively on ground water obtained from wells that reach three zones of artesian aquifers in the basalt flows and sedimentary interbeds of the Columbia River Group of Miocene age. The three artesian zones are designated the upper, middle, and lower artesian zones. The Columbia River Group is overlain by surficial sediments in which water generally occurs under water-table conditions. The Columbia River Group is underlain by a basement of crystalline rocks of pre-Tertiary age that also crops out beyond the limits of the basalt and forms the mountains that rim the basin on three sides. Where exposed at the surface, the crystalline basement contains water under water-table conditions. Neither the surficial sediments nor the crystalline basement will yield large amounts of water; the rocks of the Columbia River Group are the only source of water for public supplies. All ground water originates as precipitation that falls within the borders of Moscow basin; natural discharge of ground water is by underflow to the west.

Prior to 1960, the entire public supply was obtained from wells reaching the upper artesian zone. The quality of water was unsatisfactory because of excessive amounts of iron and moderate hardness. Between 1960 and 1965, wells were drilled into the middle and lower artesian zones, and by 1965, nearly all water pumped for public supplies was obtained from the middle and lower artesian zones. The waters from the middle and lower artesian zones contain only moderate amounts of iron and are relatively soft.

In 1896, water levels in wells in the upper artesian zone were at or slightly above land surface but declined continuously thereafter and, by 1960, static levels were nearly 120 feet below the surface in the vicinity of the City of Moscow wells. This decline led to suggestions that ground water recharge in Moscow basin was insufficient to balance pumpage. Following the phasing out of heavy pumpage of the upper artesian zone in 1960-1965, water levels rose and recovered to within 65 feet of the surface in 1969.

In our studies, two lines of investigation indicate that pumpage was not in excess of recharge during the period 1896-1960. We analyzed the pumpage and water level records of the public supply wells, using a mathematical model aquifer that utilizes the theory of image wells and assumes that there is no recharge to the basin. The results indicate that the decline of water levels in the upper artesian zone was actually much less than it would have been if pumpage was greatly in excess of recharge. We also studied the long-term records of water level fluctuations in observation wells in the basin. The water table in the surficial aquifers remained stable during the time that the water levels in the upper artesian zone declined. We attribute the decline of the water levels in the upper artesian zone to barrier boundary effects rather than to lack of recharge. The results of our studies support the views of previous workers who estimate ground water recharge to the basin by use of the equation of hydrologic equilibrium. All such estimates indicate that recharge is in excess of pumpage sufficient to meet the demands of the basin through the year 2000.

We also used our no-recharge mathematical model aquifers to estimate the total water in ground water storage in Moscow basin and to predict the decline in water levels that would occur by the year 2000. Although these figures are based on an assumption that we have rejected--no recharge to the basin--they do represent the smallest amount of water and the largest amount of drawdown to be expected. The study indicates that the middle and lower artesian zones would meet the anticipated 1965-2000 demand of 50.1 billion gallons and still have as much as 299.4 billion gallons remaining in storage in the year 2000. Water levels would be from 50 to 80 feet lower in 2000 than they were in 1965. This study indicates that ground water can supply the anticipated needs of Moscow basin well into the 21st century regardless of whether the water is derived from ground water storage or from recharge.

If need for water should exceed natural recharge at some time in the future, artificial recharge utilizing water from sources in and near Moscow basin could furnish more than 1 billion gallons of additional water annually. During a normal year, spring runoff from intermittent streams in the Palouse Range can provide about 900 million gallons over a 90 day period during February through May. The Moscow Waste Water Treatment Plant now discharges about 300 million gallons annually; the discharge should increase to about 1 billion gallons by the year 2000. The effluent could be further treated, then recycled by artificial recharge. Mathematical model studies show that the existing wells in the upper artesian zone can accept artifical recharge at rates of 1000 to 2000 gpm. for as many as 100 consecutive days without the cone of impression reaching the surface. Cost of artifical recharge probably is less than the cost of long-distance importation of water.

The waters of the surficial aquifers are relatively soft, averaging 87 ppm. hardness as $CaCO_3$, and are relatively low in dissolved solids, averaging 127 ppm. As the waters move into the upper artesian zone, average hardness increases to 135 ppm. and average dissolved solids increases to 190 ppm., probably as the result of solution of magnesium from magnesium-rich minerals in the basalts. As the waters move through the middle artesian zone and into the lower artesian zone, average hardness decreases to 84 ppm. but average dissolved solids increases to 286 ppm.; the decrease in hardness probably is the result of base exchange of sodium for calcium.

Calcium and bicarbonate are the dominant ions in most of the ground waters of Moscow basin, but calcium and sulfate or sodium plus calcium and bicarbonate are the dominate ions in a few of the waters. Excessive amounts of iron are moderately common in waters from the surficial aquifers and very common in waters from the upper artesian zone. The iron originates in high-iron clay deposits in the surficial aquifers on the outer margin of the recharge area of the artesian aquifers. The high-iron waters move laterally from the clay deposits into the upper artesian zone. Waters from the middle and lower artesian zones contained only moderate amounts of iron when the aquifers were first placed into use. Continued pumpage could induce the high-iron waters to move into the middle and lower artesian zones. Silica also is somewhat high for ground waters and the origin of the silica probably is related to the origin of the iron. Some of the waters in the surficial aquifers contain relatively high amounts of nitrates and chlorides that indicate possible contamination from septic tanks, fertilizers, and barn-yard wastes.

INTRODUCTION

During the period 1896-1960, the City of Moscow and the University of Idaho derived water from wells tapping relatively shallow artesian aquifers ("upper artesian zone") in the Columbia River Group. Water levels declined continuously through 1960 when wells tapping deeper artesian aquifers ("middle and lower artesian zones") began to be placed into service. By 1965, pumping had virtually ceased from the upper artesian zone and water levels recovered from 1960 through the end of observations in 1969. The waters from the upper artesian zone are moderately hard and contain excessive amounts of iron whereas the waters of the middle and lower artesian zones are softer and contain only moderate amounts of iron.

Purpose and Scope of Investigation

The long-continued decline of water levels in the upper artesian zone led to speculation that pumping is in excess of the natural recharge of ground water to Moscow basin. If recharge were inadquate through the 1960's, then recharge certainly would not meet the greater demands to be expected in the future. Therefore, the principal objective of this study is to examine the data on the 1896-1960 pumping of the upper artesian zone to determine if the data indicate that pumpage actually was in excess of recharge and then to predict the drawdowns in water levels in the middle and lower artesian zones in response to future pumpage. The secondary objective is to determine the origin and distribution of the iron in the waters of the upper artesian zone.

Location and Extent of the Area

Moscow basin is in west-central Latah County, in northern Idaho (Fig. 1). The amphitheater-like basin has an area of about 58 square miles. The boundaries are well defined by stream divides on the north, east, and south. The western boundary has been arbitrarily set at the Washington-Idaho state line for purposes of this report although Moscow basin is actually part of the larger Moscow-Pullman basin of Idaho and Washington. The poorly-defined southwest and northwest boundaries have been drawn to include only those tributary streams that join the main streams within Idaho. The area is approximately the same as that discussed by Stevens (1960) and Sokol (1966).

Previous Investigations

The geology and water resources of Moscow basin were first mentioned in a regional report by Russell (1897). Since then, approximately two dozen authors have written about the geology and hydrology of Moscow basin and nearby areas. The most important published reports include those by Laney and others (1923), Tullis (1944), Hosterman and others (1960), Stevens (1960), Foxworthy and Washburn (1963), Crosby and Chatters (1965), and Sokol (1966). Many of the other investigations are private reports by geologic and engineering consultants or unpublished results of research. These unpublished reports, through 1965, as well as published reports, are listed by Ross (1965, Table 1). The most significant unpublished report since 1965 is a Washington State University Master of Science in Geology Thesis (Chang-Lu, 1967).

Ross (1965) published an extensive preliminary report on our investigations to that time. Many of her conclusions and interpretations remain unchanged and are repeated below. Other publications that have resulted from our studies are a study of the possibility of artificial recharge in the basin (Jones, Ross, and Williams, 1968) and a preliminary report on the availability of ground water in the basin (Jones and Ross, 1969). Our original intention of publishing all of our basic data will not be possible. Basic data accumulated prior to 1965 were published by Ross (1965). Records of all wells and springs are on file with the Idaho Bureau of Mines and Geology in Moscow as are tabulations of data acquired since Ross' 1965 report (additional logs of wells, records of wells, all available water-level measurements, all available pumping data, and all available basic data from pumping tests.

Methods of Study

About 60 days were spent in field work during the summer and fall of 1964 and several weeks of additional field work were done in 1965 and 1966. Field work consisted of geologic mapping; of compiling an inventory of private wells and springs in the basin; of determining field chemical data such as electrical conductivity, pH, and iron; of collecting water samples for laboratory chemical analyses; of periodic measurements of water levels in observation wells; and of pumping tests of a few of the wells. Data on construction, water levels, and water chemistry were obtained for about 230 wells (estimated to be 80 percent of all wells in the basin as of 1965). Laboratory chemical analyses were run on waters from 38 of the wells and springs. Records of water-level fluctuations were obtained by periodic tape measurements or recording gage records in 11 observation wells beginning in 1963 and continuing, in some wells, into 1966. Since 1966, only one observation well has been measured periodically in Moscow basin. Three pumping tests were attempted, but the data obtained from the tests was not fully satisfactory.

Additional data, including records of water-level measurements, amounts of water pumped, and chemical analyses of water, were furnished by the City of Moscow, Physical Plant Division of the University of Idaho, and the U.S. Geological Survey. U.S. Weather Bureau precipitation records were taken from the literature.

The geologic and hydrologic data were integrated to determine the ground-water flow systems of Moscow basin. Then mathematical models of the aquifers were designed, utilizing theory of image wells. We used the models to compute theoretical drawdown in Moscow basin aquifers with various assumed aquifer properties. We then compared the theoretical drawdown with the actual drawdown in the real aquifers to determine which combinations of assumed aquifer properties produced drawdowns in model aquifers that most closely matched the drawdowns of the real aquifers. Best-fit model aquifer data were then used to predict the future drawdowns of the aquifers under natural conditions as well as the possible buildup of water levels in the aquifers under artificial recharge.

Chemical quality of water data were analyzed with the aid of maps of distribution of dissolved materials and with various graphic methods that are in general use in geochemical studies of ground water.

Well-numbering System

The well-numbering system used in this report is that formerly used by the U.S. Geological Survey in Idaho. The system indicates the location of wells within official rectangular subdivisions of the public lands, with reference to the Boise

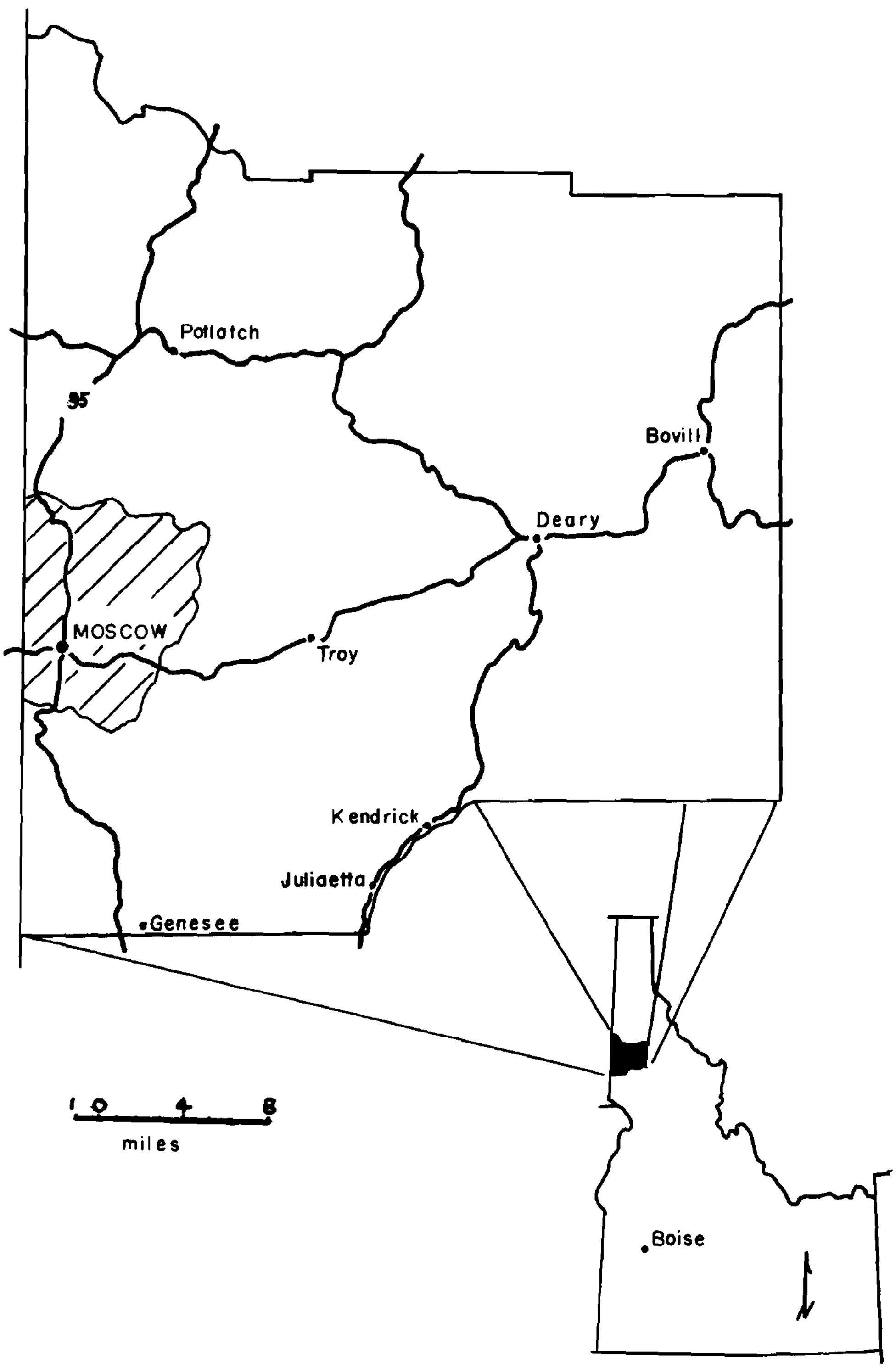

Figure 1. Location of Moscow basin, Latah County, Idaho.

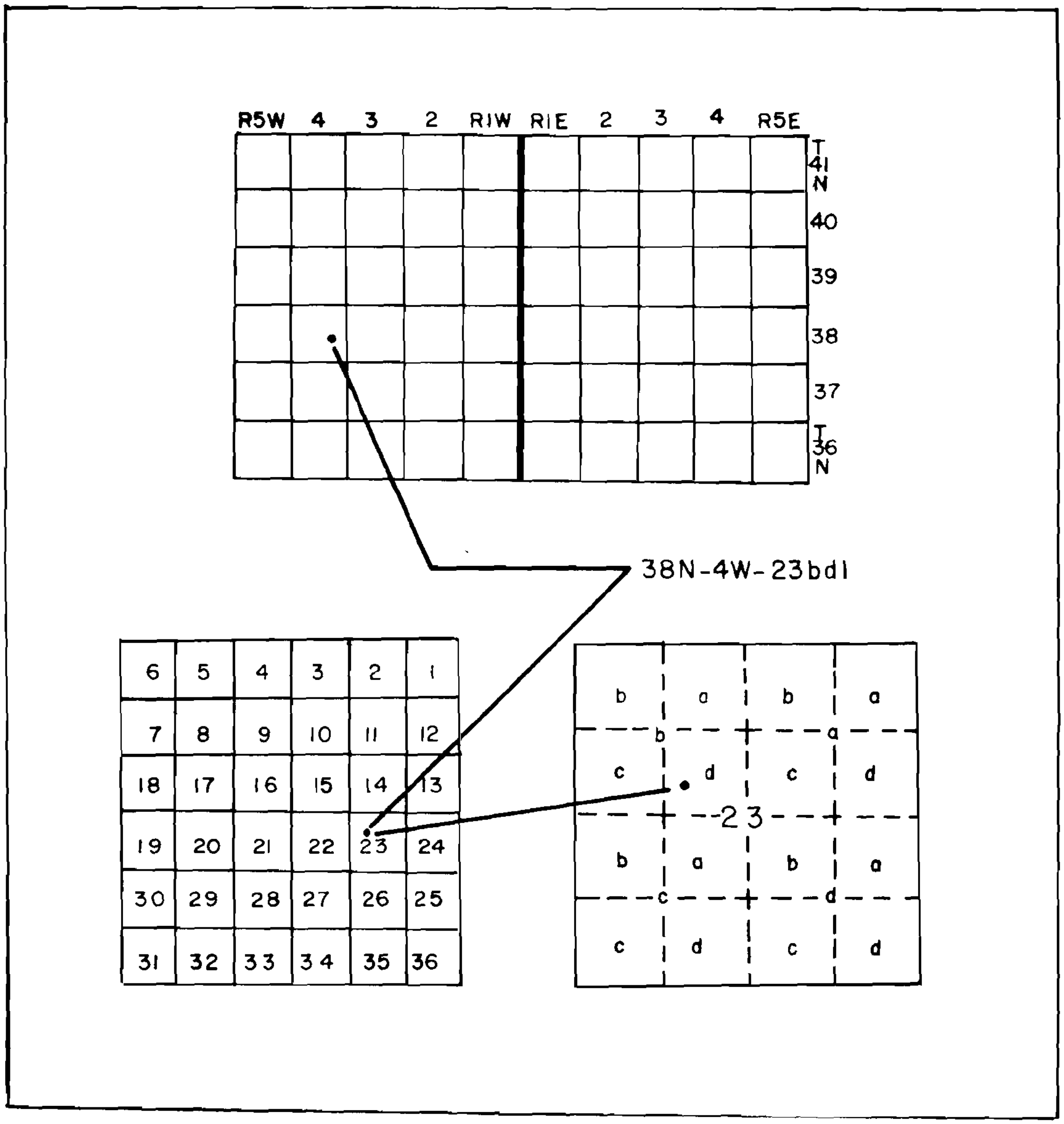

Figure 2. Sketch showing well-numbering system used in Idaho.

Table 1. Records of selected wells and springs in Moscow basin, Idaho.

Type of well: D, Dug; Dr, Drilled; Dv, Driven; B, Bored

Type of pump: P, piston; J, jet; SUB, submersible; T, turbine; C, centrifugal; R, reciprocating piston; S, suction; N, none

Use of well: D, domestic; S, stock; Irr, irrigation; Ind, industrial; PS, public supply; O, observation; U, unused; A, abandoned

Remarks: A, Chemical analysis on Table 10

Well No.	Owner or tenant	Type of Well	Year Constructed	Elev. of Land Surf.	Depth of Well (ft)	Dia. of Casing (in)	Depth to Water (ft)	Year Meas.	Kind of Pump	Use of Well	Quality of Water: Date of Collection	Temp (°F)	Iron (ppm)	Electrical Conductivity (micromhos/cm @25°C)	Remarks
39N-5W															
2bd1	Wilson Jasper	Dr	1920?	2655	212	6	157		P	D	8/6/64		0.0	230	A
5ab1	Mrs. Merle Stubbs	Dr		2710			80		J	D	11/17/65			195	Field pH 6.95, sample from tap
5ac1	Malcomb Furniss	Dr	1964	2690	110	8	50			D	6/29/65		0.0	215	Field pH 7.15, sample from tap
5ad1	Jack Marineau	Dr	1965	2680	405	8	200			D	11/17/65		9	380	A
5bb1	John Wallen	Dr	1940?	2700	230	8	220	1952	P	D	8/4/64	57	9.2	280	A
5da1	Norm Metzker	Dr	1964	2640	186	8	166	1964	J	D	9/7/65		250	1400	A
6dc1	John Ayers	Dr	1951	2680	376	6	120	1951	SUB	D	-	53	3-1/2	270	A
7ba1	Moscow; City 8	Dr	1965	2617	1442	20	212	1965	T	PS	11/16/65	75	0.6	450	A
7ba2	Moscow; City 7	Dr	1962	2614	666	20	218	1963	N	A					
7bc1	U. of I.; Univ. 3	Dr	1964	2565	1336	30	257	1963	T	PS	11/24/65	68	1.2	260	A
7cc1	U. of I.	Dr	1910?	2540		6	47	1965	P	A					
7cd1	U. of I.; Univ. 2	Dr	1951	2553	355	20	55	1951	T	PS					A
7da1	Moscow; City 3	Dr	1946	2560	245	18	81	1951	T	PS					
7da2	Moscow; City 2	Dr	1926	2560	240	15	84	1951	T	PS					A
7da3	Moscow; City 1	Dr	1882	2560	245	12	0	1895	T	PS					
7dd1	Garrett Fr'tlines	Dr		2561	231	6	51	1937	N	A					
8ab2	Moscow; City 5	Dr	1948	2660	373	24	240	1955	T	PS					
8ab3	Floyd McDonald	D		2610	46	36	15	1964	J	D	11/24/65	50	0.20	235	A
8ba1	Moscow; City 6	Dr	1959	2587	1305	28	287	1960	T	PS	11/16/65	77	0.6	420	A
8cc1	Louis Korter	Dr	1945	2560	120	6	103	1955	N	A					
8da1	Robert Jones	Dr	1964	2605	11	2	3	1964	N	O					
9ba1	A.A. Flack	Dr		2615	126	6	100		P	D	11/23/65		3.0	240	A
9bc1	Frank Eveland	D		2610	25	60	8	1964	J	D	8/12/65	54	Tr	550	A
9bc2	Frank Eveland	D		2610	13	60	8	1964	SUB	D	8/12/65		0.06	205	A
9bc4	Frank Randle	Dr	1963	2600	180	6	100	1963	SUB	D	8/12/65		Tr	250	A
9bd1	Frank Randle	D	1959	2605	27	36	18	1959	SUB	D	8/12/65		0.06	205	A
9bd2	Elmer Swanson	D	1957	2605	13	30	6	1957	J	D	8/12/65		4.0	240	A
9cc1	Chris Deesten	Dr	1950	2640	160		110	1962	P	D	11/23/65		3.6	440	A
10ac1	US Geol. Survey	Dr	1934	2630	22	1-1/2	16	1934							
10cb1	Roy Bell	Dr	1964	2670	276	6	40	1964	SUB	O	8/13/65		1-1/2	215	A
11da1	H.L. Wilcox	D		2820	21	36	7	1964	N	D	8/11/65	52	0.13	220	A
12ab1	Kenneth Nicols	Dr		2740	60	6			S	D,S	8/11/65	52	Tr	205	A
15ac1	Moscow Elks	Dr	1946	2600	252	8	60	1962	T	Irr	8/11/65		0.23	300	A

Table I. Continued.

Well No.	Owner or Tenant	Type of Well	Year Constructed	Elev. of Land Surf.	Depth of Well (ft)	Dia. of Casing (in)	Depth to Water (ft)	Year Meas.	Kind of Pump	Use of Well	Quality of Water: Date of Collection	Temp (°F)	Iron (ppm)	Electrical Conductivity (micromhos/cm @25°C)	Remarks
39N-5W (cont.)															
15bc1	U. of I; Parker Fm.	Dr	1957	2610	273	16	122	1957	T	Irr	11/16/65	53	3-1/2	240	A
15ca1	Bennet Lumber Prod.	Dr	1959	2590	252	10	90		T	Ind	8/11/65	57	6.0	320	A
16ac1	James Carrico	B		2630		18			C	D	8/13/65		0.16	770	A
16ad2	R.R. Reid	B		2640	58	12	12		J	D	8/13/65	55	Tr	200	A
16ad5	C.C. Warnick	Dr	1956	2700	246		220	1956	SUB	D	8/13/65		15.5	250	A
16ba1	George Wendt	Dr	1960?	2620	160	6			SUB	D	8/13/65	55	2.9	290	A
16bc1	Charles Jabbora	D		2595	18-1/2	60	9-1/2	1964	R	D	11/24/65	48	0.05	330	A
16bc2	Charles Jabbora	Dr	1945?	2595	40+				T	D	11/23/65	52	4.25	230	A
17cd1	Everett Hagen	Dr.	1958	2600	247	8	65		SUB	D	8/13/65		11.1	200	A
18ba1	U.Of I; Univ. 1	Dr	1920	2610	330	8	68	1921	T	PS					A
19ba3	Wayne Chestnut	Dr		2540	70	8	30	1964		D	11/24/65	54	0.4	480	A
19bb4	Eric Kirkland	Dr	1950?	2540	106	6	37	1965	J	D	8/13/65		Tr?	280	A
19da1	Jack Wren	Dr	1961	2575	45	6				D	11/23/65		<1/2	280	A
26bb1	Albert Oleson	Dr	1915?	2930	200	6	170	1964	P	D	8/13/65	54	0.26	490	A
40N-4W															
30ca1s	Fred Robison			2940					N	D	8/11/65		0.0	80	A
40N-5W															
18db2	Ray Kammeyer	Dr	1960?	3030	186	8	75	1964	SUB	D	8/11/65		1.85	140	A
18db3	Ray Kammeyer	D		3010	16	48	4-1/2	1965	S	D	11/18/65	54	0.20	80	A
20cb1	Elwood Widman	Dr	1960	2740	234	6	1	1960	SUB	D	8/11/65		0.50	250	A
29aa1	R.K. Bonnet	Dr		2740	130	8	50?		J	D	11/4/64		Tr	160	Sample from tap
29aa2	R.K. Bonnet	Dr		2720	150+	8	50.3	1964	N	A					
31ca2	N.T. Carson	D		2620	21	30	5.4	1964	N	A					
33bd1	A.E. Koster	Dr	1964	2670	90	8	12	1964	SUB	D	11/18/65	49	10.4	850	A

base line and meridian. The first two segments of the well number designate the township and range. The third segment is the section number, followed by two letters and a numeral, which indicate the quarter section, the 40-acre tract of land, and the serial number of the well within the tract. Quarter sections are lettered a, b, c, and d in counterclockwise order, starting from the northeast quarter of each section (Fig. 2). Within the quarter section, 40-acre tracts are lettered in the same manner. The serial number following the letters indicates the order in which the wells were first visited within the 40-acre tract. For example, well 38N-4W-23bdl is in the SE$\frac{1}{4}$ of the NW$\frac{1}{4}$ of Sec. 23, T. 38 N., R. 4 W., and is the well first visited in that tract.

The locations of all wells inventoried in Moscow basin are shown on Figure 3 (in pocket). Records of the wells that are discussed below are shown on Table 1 and the local designations commonly used for certain wells in Moscow basin are shown on Table 2. In general, wells will be referred to below by the local designation.

Table 2. Local Designations of Wells in Moscow Basin, Idaho

Local Designation	Idaho Bureau of Mines and Geology Well Number (Table 1)
University of Idaho	
University 1	39N-5W-18bal
University 2	39N-5W-7cdl
University 3	39N-5W-7bcl
Parker Farm	39N-5W-15bcl
City of Moscow	
City 1	39N-5W-7da3
City 2	39N-5W-7da2
City 3	39N-5W-7dal
City 4	39N-5W-8ddl
City 5	39N-5W-8ab2
City 6	39N-5W-8bal
City 7	39N-5W-7ba2
City 8	39N-5W-7bal
Bennett Lumber Products	39N-5W-15cal
Elks Golf Course	39N-5W-15acl

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The University of Idaho Water Resources Research Institute provided much of the financial support for this study. The Engineering and Water Deparments of the City of Moscow, the Physical Plant Division of the University of Idaho, the Boise office of the Water Resources Division of the U.S. Geological Survey, and the Moscow offices of the Soil Conservation Service provided data from their files. J.R. Simplot Company, local well drillers, and many well owners kindly supplied information.

The City of Moscow and the University of Idaho also aided in the installation and maintenance of automatic water-level recorders during the short-term water level studies. Some of the geologic and hydrologic information in this report has been condensed from material by Sokol (1966).

Roy E. Williams, Professor of Hydrogeology at the University of Idaho, provided much valuable advice and assistance during the later phases of the investigation, especially with respect to the model aquifer studies. He is coauthor of the preliminary report on artificial recharge in Moscow basin (Jones, Ross, and Williams, 1969).

GEOLOGIC AND HYDROLOGIC SETTING

Geology

Moscow basin consists of a rolling surface of low hills at the eastern margin of the Palouse Hills section of the Columbia River Intermontane province and of mountains at the western margin of the Northern Rocky Mountain province. The mountains rise 500 to 1700 feet above this surface on the northern, eastern, and southern borders of the basin. The mountains are underlain by granitic and metamorphic rocks. The granitic rock is mainly granodiorite of the Thatuna batholith of Cretaceous (?) age (Tullis, 1944), p. 143-174); p. 143-174); the metamorphic rock is primarily quartzite of the Belt Supergroup of Precambrian age (Tullis, 1944, p. 139-140). The geology of Moscow basin is shown on geologic maps that have been published elsewhere (Ross, 1965, Hosterman and others, 1960). Streams and other geographic features are shown on Figure 3. The stratigraphic column is shown on Figure 4.

The western part of the basin is underlain by a sequence of basalt and interbedded sedimentary material as much as 1400 feet thick. This sequence is part of the Columbia River Group that covers much of eastern Washington and northeastern Oregon as well as parts of western Idaho. The basalt generally is black and fine-grained with local zones of glassy, vesicular, or porphyritic texture. Much of it is fractured and broken. The interbedded sediments are primarily clays, silts, and fine-grained sands that commonly are designated as "Latah Formation". Some of the sedimentary beds resemble sediments deposited in lakes; other sediments are typical of stream deposits. The approximate position of the edge of the Columbia River Group is shown on Figure 5.

The Columbia River Group is overlain by reddish-brown loess (wind-blown silt) of the Palouse Formation.

Before Miocene time (about 25 million years ago), Moscow basin was part of a rugged mountain system with perhaps more than 4,000 feet of relief. The ancestral South Fork of the Palouse River, Paradise Creek, and Missouri Creek flowed in steep, narrow valleys. During the Miocene Epoch, flows of basaltic lava invaded the local drainage systems from the west. After each flow, or series of flows, surface drainage was blocked by the basalt so that ponded areas formed between the mountains east of the basin and the basalt surface to the west. Sediments were deposited in the ponded area before streams could cut new channels through the basalt dams to drain the ponded areas. Stream action dissected the lake beds and underlying basalt, and locally deposited sand and gravel in channels before the next invasion of lava. At least three major episodes of ponding occurred.

There is some indication that renewed uplift of the mountains occurred between the second and third episodes of volcanism (Cavin, 1964).

At a much later time, in the western part of the basin, wind-deposited, reworked, fine-grained sediments formed a loess mantle in which the present topography was developed by mass wasting and stream action.

Hydrology

A balance must exist between the quantity of water supplied the basin and the amount stored within or leaving the basin. A quantitative statement of this hydrologic equilibrium in its most general form is:

$$\left[\begin{array}{l}\text{Surface-water inflow + ground-water inflow + precipitation} \\ \text{+ imported water + decrease in surface storage + decrease} \\ \text{in ground-water storage}\end{array}\right]$$

$$= \left[\begin{array}{l}\text{Surface-water outflow + ground-water outflow +} \\ \text{evapotranspiration + exported water + increase in} \\ \text{surface storage + increase in ground-water storage.}\end{array}\right]$$

In Moscow basin, certain terms may be eliminated because they are not important in the basin's water balance. The modified equation becomes:

$$\text{precipitation} = \left[\begin{array}{l}\text{surface-water outflow + evapotranspiration} \\ \text{+ change in ground-water storage +} \\ \text{ground-water outflow.}\end{array}\right]$$

Ground-water recharge is equal to the increase in ground-water storage plus ground-water outflow. Therefore, the equation may be rewritten as:

$$\begin{array}{l}\text{ground-water} \\ \text{recharge}\end{array} = \left[\begin{array}{l}\text{precipitation - surface-water outflow} \\ \text{-evapotranspiration}\end{array}\right]$$

The amount of water available for recharge may be estimated if the other three factors can be measured or calculated.

Precipitation

All water in the Moscow basin enters as precipitation. Annual amounts at Moscow for the 30-year period, 1931-1960, averaged 22.2 inches. During this time, the minimum was 14.13 inches (1944) and the maximum was 34.01 (1948). Precipitation at most points in the basin is higher than at the gaging station at Moscow. Bloomsburg (1958) showed that precipitation on the Palouse Range above 4000 feet elevation is as much as twice the precipitation at Moscow. Because most of the basin is below 2800 feet elevation, but above the elevation at Moscow, the average precipitation on the basin probably approximates 1.25 times the precipitation at Moscow (Sokol, 1966, p. 7).

Precipitation is not distributed evenly throughout the year. More than two-thirds falls from October through March. Only part of the precipitation falls as rain; a relatively large amount falls as snow during the winter months. Although the snow pack at lower altitudes melts several times each winter, much of the snow at higher altitudes remains through the winter. During the spring, snow melts at progressively higher elevations, so that much snow at intermediate elevations is melted when most of the snow at high elevations remains on the ground.

Evapotranspiration

Perhaps the greatest problem in determining the hydrologic balance in

Moscow basin or elsewhere is the estimation of evapotranspiration. Stevens (1960, p. 342), using estimates by Criddle (1947) for adjacent areas, assumes that annual evapotranspiration for Moscow basin is 16.8 inches. Thus, Stevens calculates that approximately 10.5 inches of precipitation are available annually for runoff and ground-water recharge. However, he points out that his estimate may be in error as much as 25 per cent.

Sokol (1966, p. 7-8), using estimates for monthly potential evapotranspiration, calculates that approximately 12.5 inches of precipitation remain annually for runoff, infiltration, or ground-water recharge. Sokol further estimates that perhaps 9 of the 12.5 inches of excess precipitation infiltrates into the soil, mostly during the spring months. Much of this water probably is transpired or evaporated during the growing season. Some undoubtedly, however, does reach at least the shallow ground-water body.

Surface water

Few data on stream runoff from Moscow basin are available, and calculations are based on incomplete records or on extrapolations of records of four gaging stations that have been installed at various times on the drainage system. Two tributaries of the South Fork of the Palouse River, Crumarine and Gnat Creeks, were measured from 1955 to 1958. In addition, the station at Crumarine Creek was maintained intermittantly from 1958 to 1965. The South Fork of the Palouse River and Missouri Creek (also known as Missouri Flat Creek) were gaged near Pullman, Washington, during the 1930's.

Flow in Crumarine Creek (drainage area 2.4 square miles; all of the area above 2800 feet elevation) averaged almost 10 inches of water over the drainage area. Flow in Gnat Creek (drainage area 4.3 square miles; about one-half of the area above 2800 feet elevation) was equivalent to 5.1 inches of water over the area.

Average flow in the South Fork of the Palouse River and in Missouri Creek (which includes most of the Moscow basin drainage, plus additional drainage in Washington) was equivalent to 3 inches of rain over the area. Most of this area is at a lower elevation than is the area of Gnat and Crumarine Creek drainages.

On the basis of the above data, Sokol (1966, p. 10) summarized the surface water regimen in Moscow basin as follows:

> ....streams respond mainly to snow melt above an altitude of 2800 feet. At lower elevations, heavy rainfall is a more important source of stream flow. Where the streams flow on a granitic bedrock, they gain water through seepage from shallow ground water throughout the year. Where they flow over loess, the streams lose water through infiltration to ground water.

Ground Water

Ground water occurs in all the rock types in Moscow basin; pre-Tertiary crystalline rock, the Palouse Formation, alluvium, and the basalts and intercalated sediments of the Columbia River Group. The surficial sediments contain water generally under unconfined (water table) conditions, and confined (artesian) water occurs in crevices, brecciated zones, and vesicles in the basalt and in intercalated sand layers in the Columbia River Group.

HYDROSTRATIGRAPHIC UNITS

Well logs, chemical quality of water, and water-level records show that the Moscow basin ground-water system can be divided into: (1) surficial aquifers, (2) three artesian zones within the Columbia River Group, and (3) the buried basement complex. Relations of the hydrostratigraphic units and the terminology used are shown on Figure 4.

Subsurface maps of Moscow basin have been published by several previous workers. Ross (1965, pl. 3) shows structure contours on the top of the upper basalt; another interpretation is shown by Crosby and Cavin (1966, Fig. 2). Crosby and Cavin also published a structure contour map (op. cit., Fig. 3) on the top of the buried crystalline basement. This map showed a basement high centered beneath the city of Moscow pumping plant. The high was based on "quartzite" reported at depth of 553 feet in the original log of City Well 3. Crosby considered this high anomolous. At his suggestion, the City of Moscow had a core hole drilled to a depth of 615 feet adjacent to the site of City Well 3. The cores are basalt. Therefore, the buried high on the crystalline basement shown in NE$\frac{1}{4}$ SE$\frac{1}{4}$ of sec. 7 on Crosby and Cavin's map does not exist. Acting on Crosby's suggestion (oral communication), we have prepared a revised version of the basement map which is shown as Figure 5.

Chang-Lu (1967) published subsurface maps of several of the units in Moscow basin. However, he did not publish his basic data; therefore we are unable to evaluate the validity of his interpretations. The well logs available to us do not, in our opinion, provide sufficient data to justify interpretations in the sort of detail that were presented by Chang-Lu.

However, at least some of his interpretations seem to be correct. His Figure 29 shows a buried valley on the top of the pre-upper basalt surface which he considers to be a control in permeability of the upper basalt aquifer and therefore a control of the shape of the piezometric surface of the upper basalt aquifer, as shown in his Figure 30. We believe that his idea is essentially correct although we differ in our interpretation of the shape of the piezometric surface.

Surficial Aquifers

The surficial aquifers are in (1) surficial sediments and (2) exposed crystalline basement. The surficial sediments include the loess of the Palouse Formation and alluvial deposits as well as some rocks which may properly belong with the "Latah Formation". The rock units of the surficial aquifers generally are less than 200 feet thick and rest either on the Columbia River Group or on the crystalline basement. The aquifers in the exposed crystalline basement are permeable zones in the weathered rocks near the surface or are fractures in less-weathered rocks at greater depths.

Surficial Sediments

Large-diameter, shallow, dug wells yield moderate to small amounts of water from the loess of the Palouse Formation. Many of these shallow wells go dry, or nearly dry, in the summer months. Although the water table may stand near the ground surface, yield is small because the specific capacity (gallons per minute per foot of drawdown) is small.

The alluvial aquifers also yield small to moderate amounts of water to wells. Much of the alluvium is poorly sorted and clay lenses are numerous. However, some lenses of sand are present that will furnish sufficient water to a well to irrigate several acres. A well penetrating a clay lens may not yield enough water for household use; the distance between a productive well and a useless well may be as little as 15 feet. Some of the water in the alluvial aquifers is under artesian conditions, but the artesian systems are small and will not yield large quantities of water.

Exposed Crystalline Basement

Only small amounts of water can be obtained from the exposed metamorphic and plutonic basement rocks of pre-Tertiary age. Springs commonly are utilized for domestic or stock water in the parts of the basin underlain by these rocks. Some homes that have no well or spring use water that is trucked to a storage tank.

A few wells obtain small amounts of water from the near-surface weathered zone or from fractures in the deeper, less-weathered rocks. Only a few wells in Moscow basin obtain more than 2 or 3 gallons per minute from the crystalline basement. Experience elsewhere (Meinzer, 1923, p. 143-147) indicates that weathered or fractured crystalline rocks generally should produce enough water for domestic purposes within a depth of 200 feet from the surface, but cannot be depended upon to yield more than a few gallons per minute. In parts of Moscow basin, wells in the crystalline basement have yielded water from depths as great as 300 feet. Our experience in Moscow basin suggests that, if adequate water is not obtained in the first 300 feet of drilling in the crystalline basement, the well should be abandoned.

Columbia River Group

Several artesian aquifers are present in the various units of the Columbia River Group. These aquifers conform to the pattern described by Newcomb (1959) for the Columbia River Group; they consist of numerous individual aquifers that are separate tabular zones with poor hydraulic connection between zones. In the Moscow basin, some of the aquifers are in the basalt flows and some are in the sedimentary interbeds. The lateral extent of the different aquifers is variable; therefore, hydraulic connection between nearby wells of similar depths is variable. In some places, water levels in a well will show considerable effect of pumping from a nearby well of similar depth; in other places, nearby wells will have little effect and more distant wells may have noticeable effect.

Although the artesian aquifers of the Columbia River Group in Moscow basin are complex, they can be generalized into three zones of aquifers having similar artesian head and chemical quality of water (Fig. 4). Each zone corresponds to one of the three sequences of lava flows and interfingering and underlying sedimentary interbeds. The zones are designated as the upper, middle, and lower artesian zones.

The upper artesian zone extends from the surface--or the base of the surficial aquifers--to depths of about 700 feet. Until the early 1960's, almost all water pumped in the basin came from this zone. Two University of Idaho wells, four City of Moscow wells, and perhaps two dozen private wells derived water

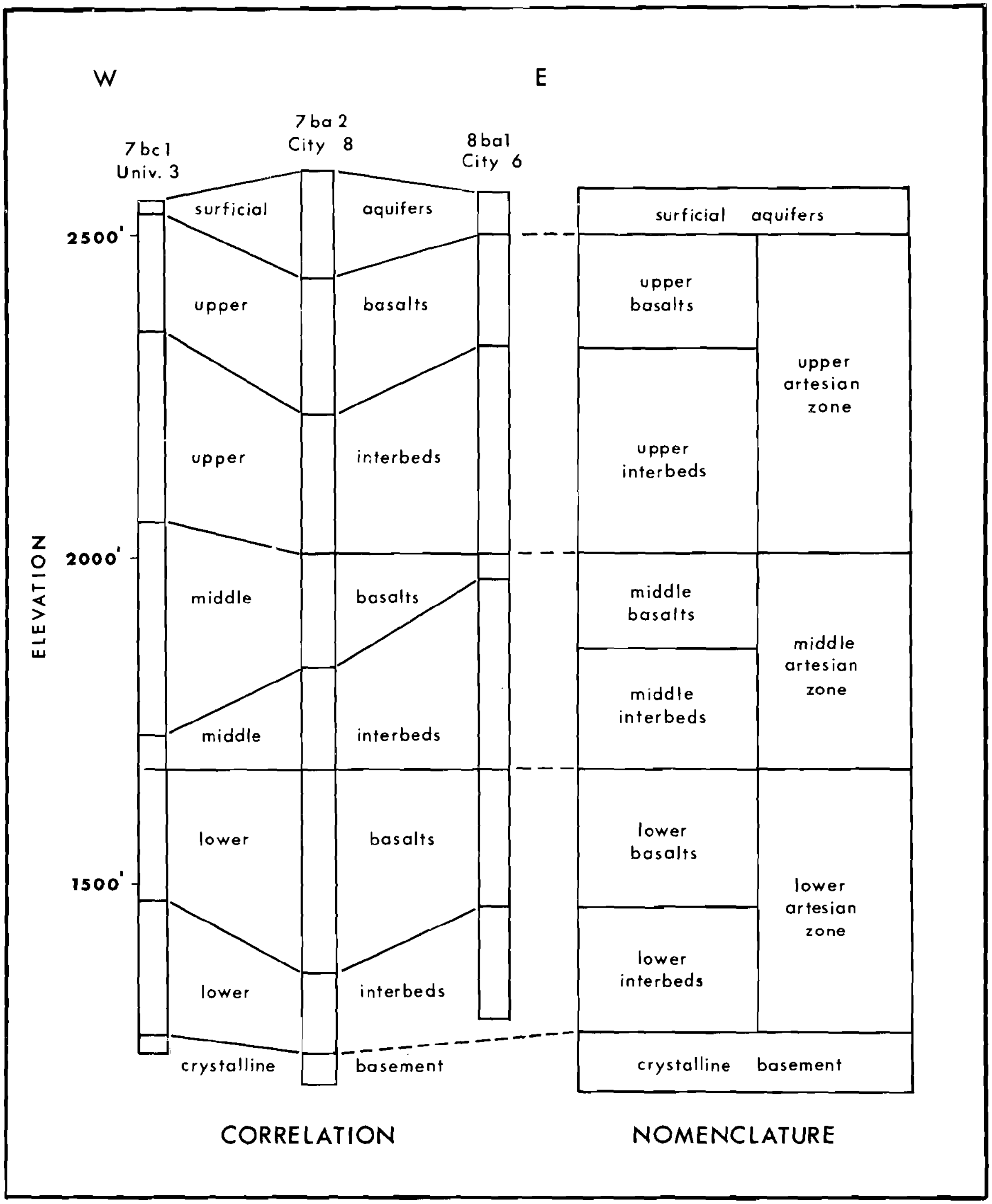


Figure 4. Nomenclature of hydrostratigraphic units and their correlation between three public-supply wells, Moscow basin, Idaho.

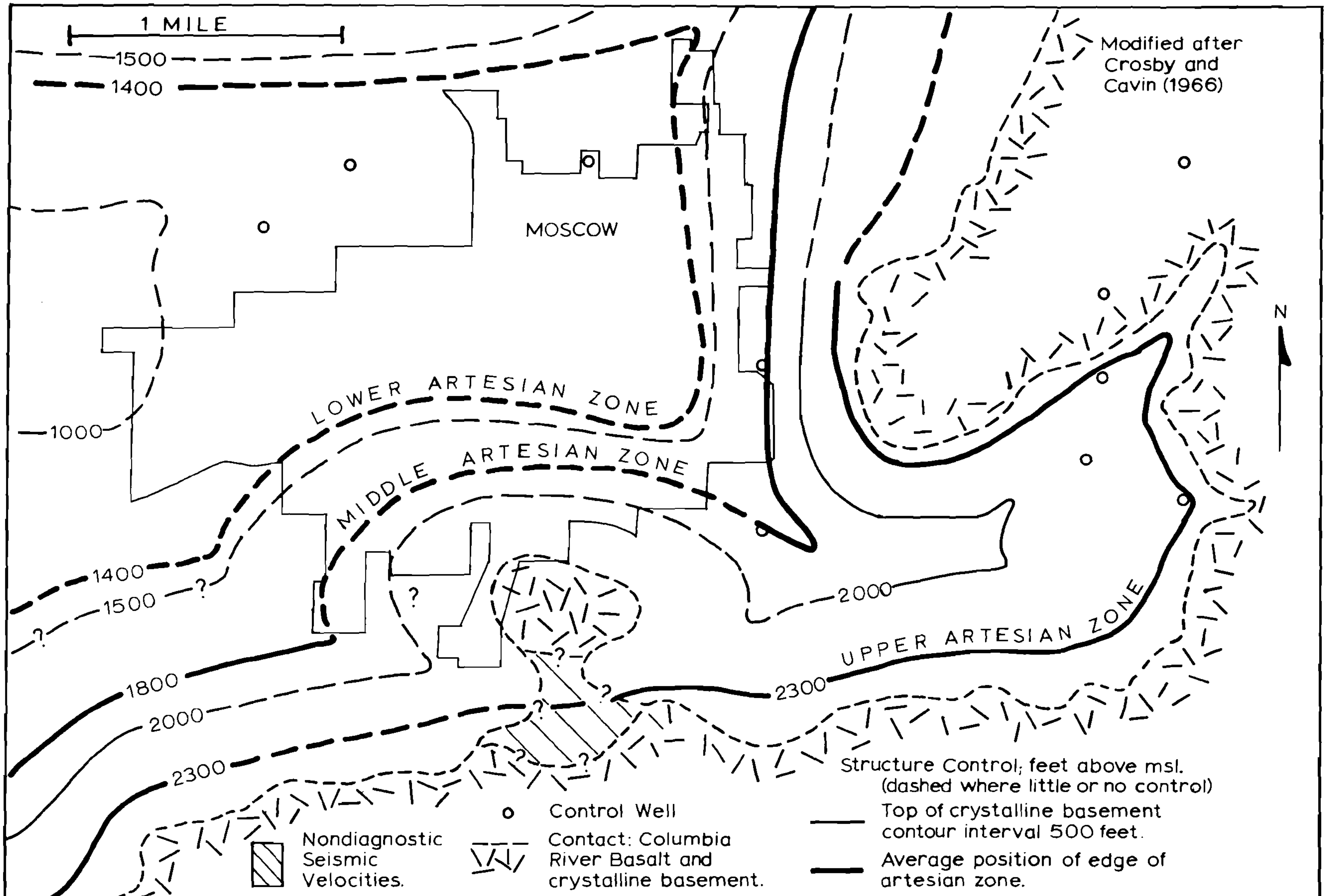


Figure 5. Structure contours on top of crystalline basement and on average position of edges of artesian zones, Moscow basin, Idaho. Control by well logs and seismic data.

either from permeable zones in the basalt or from lenses of sand immediately beneath the lowest layer of basalt. The University and City wells and one irrigation well pumped large amounts of water, mostly from sand lenses at about 2300 to 2400 feet. Beginning in the early 1960's the University and City put deeper wells into service and decreased pumpage from the upper artesian zone. By the mid-1960's, heavy pumping from the upper artesian zone ceased except during peak demand periods or during emergencies when the deeper wells could not be pumped because of difficulties with the pumps. Although the number of domestic wells drilled into the upper artesian zone has increased every year, the combined total pumpage of the domestic wells in the late 1960's probably was no more than 100 gpm.

The middle artesian zone lies at depths of 700 to 900 feet and the lower artesian zone, 900 to 1500 feet. Only a few public-supply wells penetrate these deeper zones; domestic wells generally obtain adequate water in the upper artesian zone. The only well that obtains large amounts of water from the middle artesian zone is University Well 3. Although the well was drilled through the lower artesian zone and into the crystalline basement, the lower artesian zone was not productive at University 3 and the casing was perforated between depth of 660 to 775 feet, (elevation 1905-1790) in the middle artesian zone. One well (City 4) that was drilled into the middle interbed was abandoned by the City of Moscow because of inadequate yield.

City 6 and City 8 derive water from the lower artesian zone--City 6, from both basalt and a sedimentary interbed, and City 8, from basalt. The productive zones are in the lower part of the lower artesian zone, at about depths of 1000 to 1400 feet (elevations 1600 to 1200 feet).

Buried Crystalline Basement

Crystalline rocks underly the surficial sediments and the Columbia River Group throughout Moscow basin. West of the outer margin of the basalt (Fig. 5), only 2 wells had been drilled into the buried crystalline basement by the late 1960's. In City 8, granitic rocks were reached at a depth of 1390 feet (Ross, 1965, p. 107) and were penetrated for 42 feet without any significant amount of water being discovered. In University 3, granitic rocks were reached at a depth of 1320 feet and were penetrated for 26 feet.

Experience elsewhere (Meinzer, 1923, p. 143-147) has shown that crystalline rocks seldom will yield more than a few tens of gallons per minute (gpm) to wells and almost never yield as much as 100 gpm. Because of the great depth to the buried crystalline basement, the expense of drilling to the basement is not justified in view of the small yields to be expected. For all practical purposes, the buried crystalline basement should not be considered an aquifer in the Moscow basin.

HYDRAULIC PROPERTIES OF THE AQUIFERS

Sizes and Shapes of the Aquifers

Surficial Aquifers

The individual aquifers in the surficial zones are small. The aquifers are lenticular sand bodies in the surficial sediments or are tabular to sheet-like fractures in the crystalline rocks. Minimum dimensions of most of the aquifers are as little as fractions of an inch in the fracture zones and a few feet in the sand bodies; maximum dimensions are no more than a few hundred feet laterally and a few feet vertically.

Artesian Aquifers

The artesian aquifers are bounded on north, east, and south by the relatively impermeable rocks of the crystalline basement. Much of the recharge to the artesian zones probably comes from downward percolation along this contact zone; some of the recharge may come from vertical percolation through the rocks overlying the artesian aquifers.

For the purposes of computations involving day-to-day operation of wells, the contacts on north, east, and south are regarded as barrier boundaries, and the artesian aquifers are regarded as having infinite extent to the west. Some sort of boundary must be present west of the Idaho-Washington state line, but the position and nature of the boundary is not known. The state line is arbitrarily used as the western border of Moscow basin; although this border has important political effects on the development of Moscow basin water, it is hydraulically neutral.

Much of the production from the upper artesian zone came from the upper 100 to 200 feet--elevation 2300 to 2400. For purposes of generalization, the 2300 foot contour on the crystalline basement (Fig. 5) was designated as the average position of the edge of the upper artesian zone. The upper artesian zone can be generalized as a rectangle about 20,000 feet across from north to south and 17,000 feet from east to west (Fig. 9). A centrally-located well on the state line would be the furthest from barrier boundaries in the basin--10,000 feet from the north and south barriers and 17,000 feet from the east barrier boundary. All existing wells in the upper artesian zone are closer than these distances to at least one barrier boundary.

Wells in the two deeper zones generally are closer to barrier boundaries than those in the upper artesian zone. The 1800 foot contour on the crystalline basement (Fig. 5) was designated as the average position of the edge of the middle artesian zone; the middle artesian zone can be generalized as a rectangle about 10,000 feet across from north to south and 15,000 feet from east to west (Fig. 9). The 1400 foot contour was designated as the average position of the edge of the lower artesian zone; the lower artesian zone can be generalized as a rectangle about 7500 feet from north to south and 15,000 feet from east to west (Fig. 9).

Ground-Water Circulation System

The unconfined aquifers in Moscow basin are recharged by direct infiltration of rainwater and snowmelt and, in the lower elevations, by stream-bed percolation. Discharge is to surface streams and by downward percolation to the artesian aquifers. In some reaches of streams flowing across the exposed crystalline basement, considerable exchange takes place back and forth between ground water and surface water (Chang-Lu, 1967, p. 61-69). The ground-water circulation system of Moscow basin is shown on Figure 6.

The artesian aquifers are recharged mainly by water moving downward through the discontinuous zone of unconsolidated sediments and weathered crystalline rock lying between the edge of the basalts and the crystalline basement. Vertical recharge may come from water moving downward from overlying water table or artesian aquifers, either directly downward through underlying units or laterally to the edge of the aquifer, then downwards.

The first wells drilled in the upper artesian zone, in the late 1890's, were flowing wells (Russell, 1901). At that time, the piezometric surface of the upper artesian zone was higher than the water table of the overlying surficial aquifers, at least in the lower parts of the basin. Under these conditions, some of the water from the upper artesian zone could have discharged by upward movement into the overlying water table bodies. However, most of the discharge probably took place by underflow to the west, out of Moscow basin. The relations between water levels were reversed when pumping drew the piezometric surface of the upper artesian zone below the water table. In response to the reversal of gradient, the water from the water table aquifers could move downward into the artesian zones.

The piezometric surfaces of the middle and lower artesian zones apparently always have been lower than that of the upper artesian zone. In late 1966, the elevation of the piezometric surface of the upper artesian zone in the center of Moscow basin was at about 2490 feet (Fig. 10), whereas the elevations of the piezometric surfaces of the middle and lower artesian zones were both about 2300 feet.

This head differential of nearly 200 feet is a driving force that could cause vertical recharge by water moving downward to the middle and lower artesian zones from the upper artesian zone. The head differential would have to move the water from the elevation 2300 (productive levels of the upper artesian zone) to elevation 1905-1760 and 1600-1200 (productive levels of the middle and lower artesian zones). The water would have to move vertically through the upper interbed, which is dominately silty sediments, then through the upper part of the middle basalt to the productive level in the middle basalt, then through the middle interbed and upper part of the lower basalt to the productive level in the lower basalt.

In the absence of data on the vertical permeability of the beds, analysis of actual vertical recharge is not possible. Even theoretical analysis would serve little purpose because of the difficulty of estimating permeabilities. Data in Wenzel (1942, p. 13) show that permeabilities of silty sediments can range through six orders of magnitude. Vertical recharge probably does occur in Moscow basin, but the amount cannot be estimated with the data available.

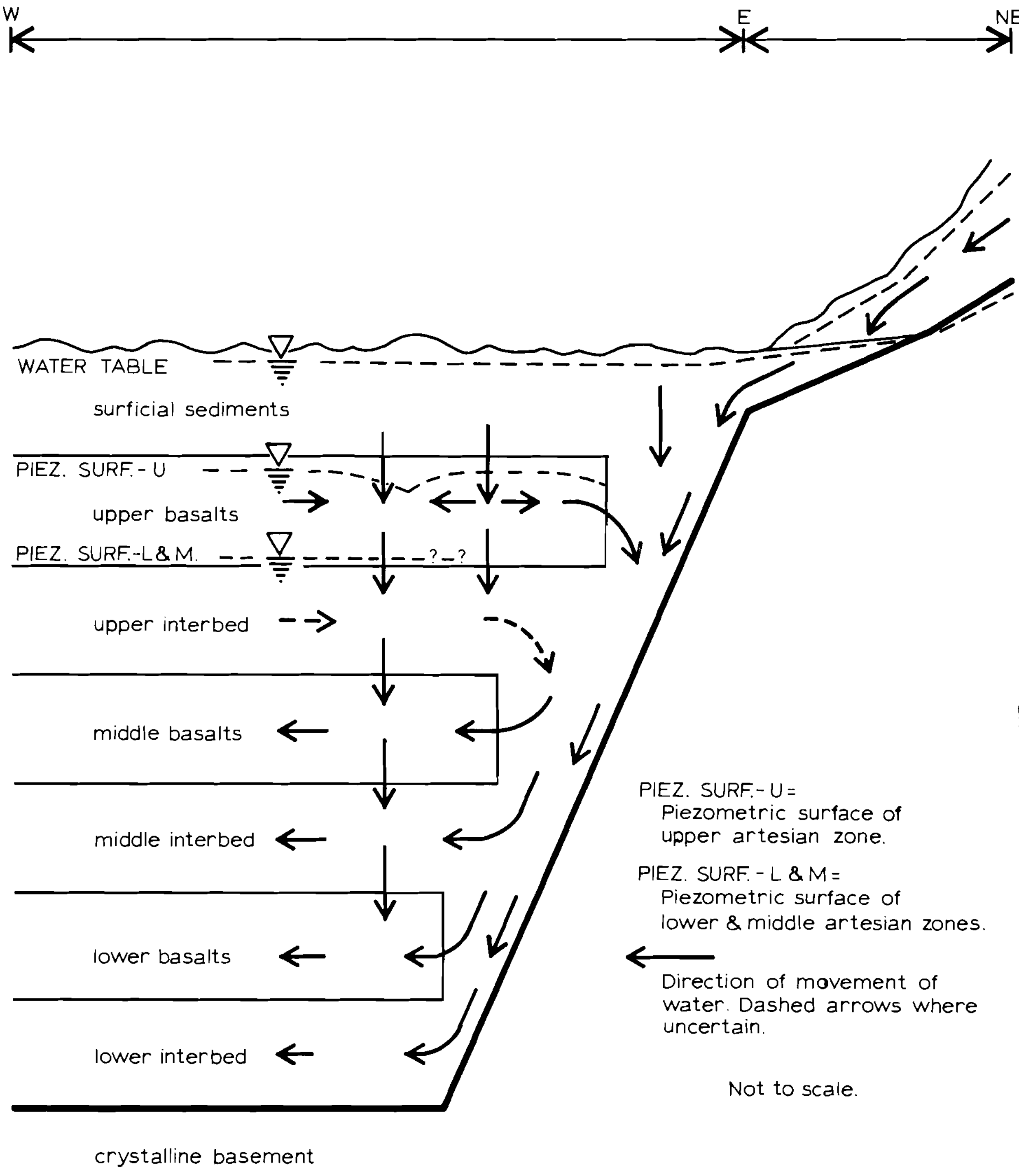


Figure 6. Schematic representation, 1966 ground-water circulation system, Moscow basin, Idaho. For discussion, see text.

The head differential between the upper artesian zone and the deeper zones also could cause water to move laterally through the upper artesian zone into the consolidated materials at the margin of the basin, then downwards into the deeper artesian zones. Ross (1965, p. 51 and pl. 8) mapped a trough in the piezometric surface of the upper artesian zone that coincides with the edge of the zone. This trough probably represents the decline in head as the laterally moving waters pass out of the upper artesian zone into the aquifers in the unconsolidated materials.

Factors in Aquifer Analysis

The hydraulic properties of aquifers usually are computed by analysis of data obtained through special types of field pumping tests. Field methods are of limited value in Moscow basin because:

1. Too few wells pump from the deeper artesian zones.

2. Aquifer boundaries are close and have so great an effect on water levels as to make analysis of pump-test data very difficult.

3. Drawdowns are small for high pumping rates; recovery is so rapid that recovery tests are not satisfactory.

One pumping test was run on a well in the surficial sediments and one on a well in the crystalline basement. The City of Moscow arranged for one short test using wells in the lower artesian zone. Production pumping tests were run on many wells in Moscow basin at the completion of drilling; however, water level measurements from production tests generally are not sufficiently frequent and precise, and pumping rates seldom are sufficiently closely controlled, to provide data from which hydraulic properties of aquifers can be computed.

Most of the aquifer properties were estimated. Values were estimated from production pumping test data using methods in the literature, or from values reported from similar aquifers elsewhere, or from studies of mathematical models of the Moscow basin aquifers.

The two most important hydraulic properties of aquifers are transmission of water and storage of water. These properties are expressed quantitatively as the two "aquifer constants", defined as (Ferris and others, 1962, p. 72-78):

Coefficient of Transmissibility = "T"

> Rate of flow of water, at the prevailing water temperature, in gallons per day, through a vertical strip of the aquifer 1 foot wide extending the full saturated height of the aquifer under a hydraulic gradient of 100 percent. Expressed as gallons per day per foot; gal/day/ft.

Coefficient of Storage

> 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. Expressed as a dimensionless decimal fraction.

Where proper data are available, coefficient of transmissibility and coefficient of storage generally are computed by some variation of the Theis nonequilibrium equation.

The Nonequilibrium Equation

Although the nonequilibrium equation of Theis (1935) is based on a number of assumptions that should limit its usefulness, it has proved reasonably accurate in many ground-water flow systems that do not meet all of the restrictions of the equation.

We used the modified nonequilibrium equation of Jacob (1950) in order to simplify calculations and permit the use of straight-line solutions. The form that we used is from Ferris and others (1962, p. 92):

$$s = \frac{264Q \log 10}{T} \frac{(0.3\ Tt)}{r^2 S}$$

The factors in the equation are:

s = drawdown, feet.

Q = pumping rate, gallons per minute.

T = coefficient of transmissibility, gallons per day per foot.

t = time, days

r = distance, from pumping well to point where drawdown is being calculated, in feet.

S = coefficient of storage, dimensionsless fraction.

One of the assumptions of the nonequilibrium equation is that the aquifer is of infinite areal extent. Although this assumption is not valid in Moscow basin, the finite real aquifers can be transformed into imaginary infinite aquifers by the use of the theory of image wells.

Image Well Theory

The following discussion, taken verbatum from Ferris and others (1962, p. 144-168), is presented here to provide an understanding for the following sections in which image well theory is used to analyze the effects of the aquifer boundaries of Moscow basin. Illustrations were taken from Walton (1962, Fig. 9 and 11).

> "The development of the equilibrium and nonequilibrium formulas discussed in the preceding sections was predicted in part on the assumption of infinite areal extent of the aquifer, although it is recognized that few if any aquifers completely satisfy this assumption. In many instances the existence of boundaries serves to limit the continuity of the aquifer, in one of more directions, to distances ranging from a few hundred feet to as much as tens of miles. Thus when an

aquifer is recognized as having finite dimensions, direct analysis of the test data by the equations previously given is often precluded. It is often possible, however, to circumvent the analytical difficulties posed by the aquifer boundary. The method of images, widely used in the theory of heat conduction in solids, provides a convenient tool for the solution of boundary problems in ground-water flow. Imaginary wells or streams, usually referred to as images, can sometimes be used at strategic locations to duplicate hydraulically the effects on the flow regime caused by the known physical boundary. Use of the image thus is equivalent to removing a physical entity and substituting a hydraulic entity. The finite flow system is thereby transformed by substitution into one involving an aquifer of infinite areal extent, in which several real and imaginary wells or streams can be studied by means of the formulas already given. Such substitution often results in simplifying the problem of analysis to one of adding effects of imaginary and real hydraulic systems in an infinite aquifer.

An aquifer boundary formed by an impermeable barrier, such as a tight fault or the impermeable wall of a buried stream valley that cuts off or prevents ground-water flow, is sometimes termed a "negative boundary". Use of this term is discouraged, however, in favor of the more meaningful and descriptive term "impermeable barrier". A line at or along which the water levels in the aquifer are controlled by a surface body of water such as a stream, or by an adjacent segment of aquifer having a comparatively large transmissibility or water-storage capacity, is sometimes termed a "positive boundary". Again, however, use of the term is discouraged in favor of the more precise terms "line source" or "line sink", as may be appropriate.

Although most geologic boundaries do not occur as abrupt discontinuities, it is often possible to treat them as such. When conditions permit this practical idealization, it is convenient for the purpose of analysis to substitute a hypothetical image system for the boundary conditions of the real system.

In this section, where the analysis of pumping-test data is considered, several examples are given of image systems required to duplicate, hydraulically, the boundaries of certain types of areally restricted aquifers. It should be apparent that similar methods can be used to analyze flow to streams or drains through areally limited aquifers.

An idealized section through a discharging well in an aquifer bounded on one side by an impermeable barrier is shown in Figure 7. It is assumed that the irregularly sloping boundary can, for practical purposes, be replaced by a vertical boundary, occupying the position shown by the vertical dashed line, without sensibly changing the nature of the problem. The hydraulic condition imposed by the vertical boundary is that there can be no ground-water flow across it, for the impermeable material cannot contribute water to the pumped well. The image system that satisfies this condition and permits a solution of the real problem by the Theis equation is shown in Figure 7c. An imaginary discharging well has been placed at the same distance as the real well from the boundary but on the opposite side, and both wells are on a common line perpendicular to the boundary. At the boundary the drawdown produced by the image well is equal to the drawdown caused by the real well. Evidently, therefore, the drawdown cones for the real and the image wells will be symmetrical and will produce a ground-water divide at every point along the boundary line. Because there can be no flow across a divide, the image system satisfies the boundary condition of the real problem and analysis is simplified to consideration of two discharging wells in an infinite aquifer. The resultant drawdown at any point on the cone of depression in the real region is the algebraic sum of the drawdowns produced at that point by the real well and its image. The resultant profile of the cone of depression, shown in Figure 7d is flatter on the side of the well toward the boundary and steeper on the opposite side away from the boundary than it would be if no boundary were present.

An idealized section through a discharging well in an aquifer hydraulically controlled by a perennial stream is shown in Figure 8. For thin aquifers the effects of vertical flow components are small at relatively short distances from the stream, and if the stream stage is not lowered by the flow to the real well there is established the boundary condition that there shall be no drawdown along the stream position. Therefore, for most field situations it can be assumed for practical purposes that the stream is fully penetrating and equivalent to a line source at constant head. An image system that satisfies the foregoing boundary condition, as shown in Figure 8b allows a solution of the real problem through use, in this example, of the Theis nonequilibrium formula. Note in Figure 8b that an imaginary recharging well has been placed at the same distance as the real well from the line source but on the opposite side. Both wells are situated on a common line perpendicular to the line source. The imaginary recharge well operates simultaneously with the real well and returns water to the aquifer at the same rate that it is withdrawn by the real well. It can be seen that this image well produces a buildup of head everywhere along the position of the line source that is equal to and cancels the drawdown caused by the real well which satisfies the boundary condition of the problem. The resultant drawdown at any point on the cone of depression in the real region is the algebraic sum of the drawdown caused by the real well and the buildup produced by its image. The resultant profile of the cone of depression, shown in Figure 8b is flatter on the landward side of the well and steeper on the riverward side, as compared with the shape it would have if no boundary were present."

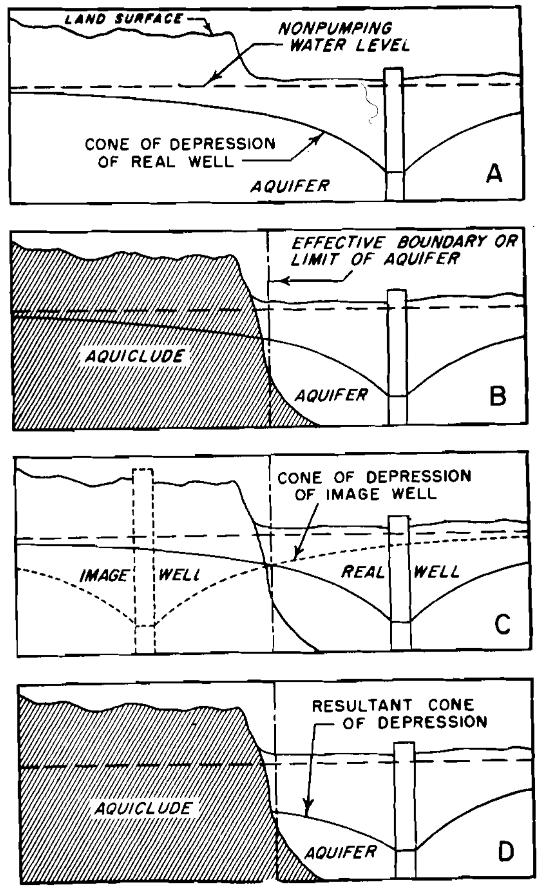


Figure 7. Diagrammatic representation of the image well theory as applied to a barrier boundary. (Walton, 1962, Fig. 9)

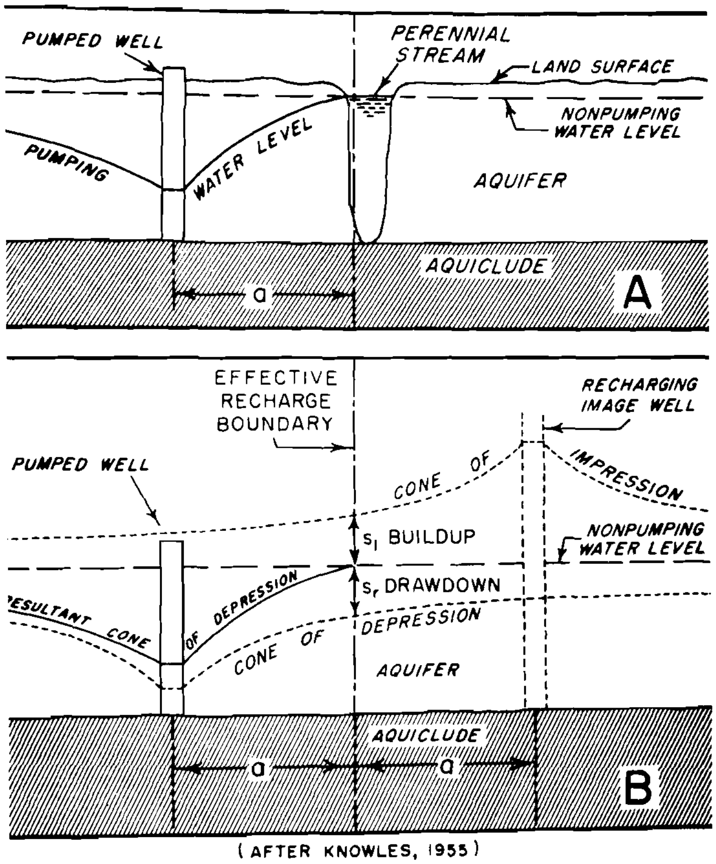


Figure 8. Diagrammatic representation of the image well theory as applied to a recharge boundary. (Walton, 1962, Fig. 11)

Image Well Array for the Moscow Basin Artesian Aquifers

The irregular boundaries of the three artesian zones (Fig. 9) can be generalized as three straight lines interesecting at right angles, an image well array that is discussed by Ferris and others (1962, p. 156-159). Preliminary studies of this aquifer model showed that little error is introduced by utilizing only one well that is equidistant from the north and south boundaries, rather than using several wells at the actual locations of real wells. Using the single central well greatly reduces the number of image wells necessary and thus greatly reduces the number of calculations necessary for each solution of the model.

Figure 9 shows the models for the three artesian zones. Dimensions of the models are shown on Table 3. The image well arrays in all three models continue outward to infinity. The number of wells that will affect the central well depends on values selected for the length of time of pumping, coefficient of storage, and coefficient of transmissibility in the model under study. In some of our model aquifer studies, several hundred image wells were needed before the effect of additional wells became small enogh to be neglected.

Table 3. Dimensions of the models of the artesian zones, Moscow basin, Idaho

Artesian Zone	Width N-S ft.	Distance from Central Well to Eastern Boundary ft.	Spacing Between Image Wells ft.
Upper	20,000	8,500	20,000
Middle	12,000	10,000	12,000
Lower	7,500	6,000	7,500

Variation in Specific Capacity of a Well

The specific capacity of a well is the number of gallons pumped per minute, per foot of drawdown. Specific capacity is not a constant for an individual well, but tends to decrease with increase in discharge owing to increased friction and turbulence in the well and pump ("well loss"). Specific capacity also decreases if the cone of influence of the well migrates outward far enough to reach one or more barrier boundaries. Decrease in specific capacity because of barrier boundary effect can be detected only if the well is pumped at a constant rate.

Specific capacity may increase with time of pumping. Generally, this happens during the production test immediately following the drilling of the well and indicates that the well was not fully developed prior to the test. This effect is most common in screened or perforated wells in unconsolidated sediments.

Hydraulic Properties of the Surficial Aquifers

Water Table

Ross (1965, pl. 5) has published a map of the water table in the surficial aquifers of Moscow basin. With minor exceptions that are of no importance here, the map shows the usual "subdued reflection" of the surface topography.

Aquifer Constants

Ross (1965, p. 42-48) discussed the results of a pumping test on well 39N-5W-9bc2, a dug, water-table well in surficial sediments, probably alluvium. The results of the test are:

Coefficient of Transmissibility ("T") = 1.2×10^3 gal/day/ft.

Coefficient of Storage ("S") = 9.0×10^{-2}

The T is rather low, but probably is typical of the fine-grained surficial sediments. The S is in the lower part of the range of values that is typical of water-table wells (Ferris and others, 1962, p. 78).

We ran a pumping test of well 40N-5W-29aal with well 29aa2 as an observation well. The wells are in the exposed crystalline basement. Well 29aa2 is a project observation well on which a recording gage operated for about ten months. The recording gage record showed no evidence of water level changes in response to changes in barometric pressure. Therefore, 29aa2 probably is a water-table well. The long-term hydrograph from the recording gage record on well 29aa2 is shown in Figure 14. During the test, the presence of several barrier boundaries became evident when sharp inflections showed up in the time-drawdown plots of the data. These barriers probably are the ends of fracture zones from which the water was derived. The results of the test (prior to barriers coming into effect) are:

Coefficient of Transmissibility ("T") = 7.0×10^3 gal/day/ft.

Coefficient of Storage ("S") = 2.0×10^{-4}

The effect of the barrier boundaries was to reduce the apparent T to about 2.0×10^3 or 3.0×10^3 gal/day/ft. Such reductions are to be expected as various fractures are depleted of water. The S is unusually low for a water-table well.

Hydraulic Properties of the Artesian Aquifers

Piezometric Surfaces

We cannot map the piezometric surfaces of the middle and lower artesian zones because too few wells reach them, however, numerous wells penetrate the upper artesian zone.

Maps of the piezometric surface to the upper artesian zone have been published by several workers. Laney, Kirkham and Piper (1923) show a simple pattern, open to the west, which is based on water levels in about 20 wells. Their original data were not published. Ross (1965, pl. 8) shows a somewhat more complex pattern, based on about 17 wells. A later version based on about 40 wells is reproduced, in a simplified version, as Figure 10 (Jones, Ross, and Williams, 1968, p. 280). Basic data to support the interpretation are in

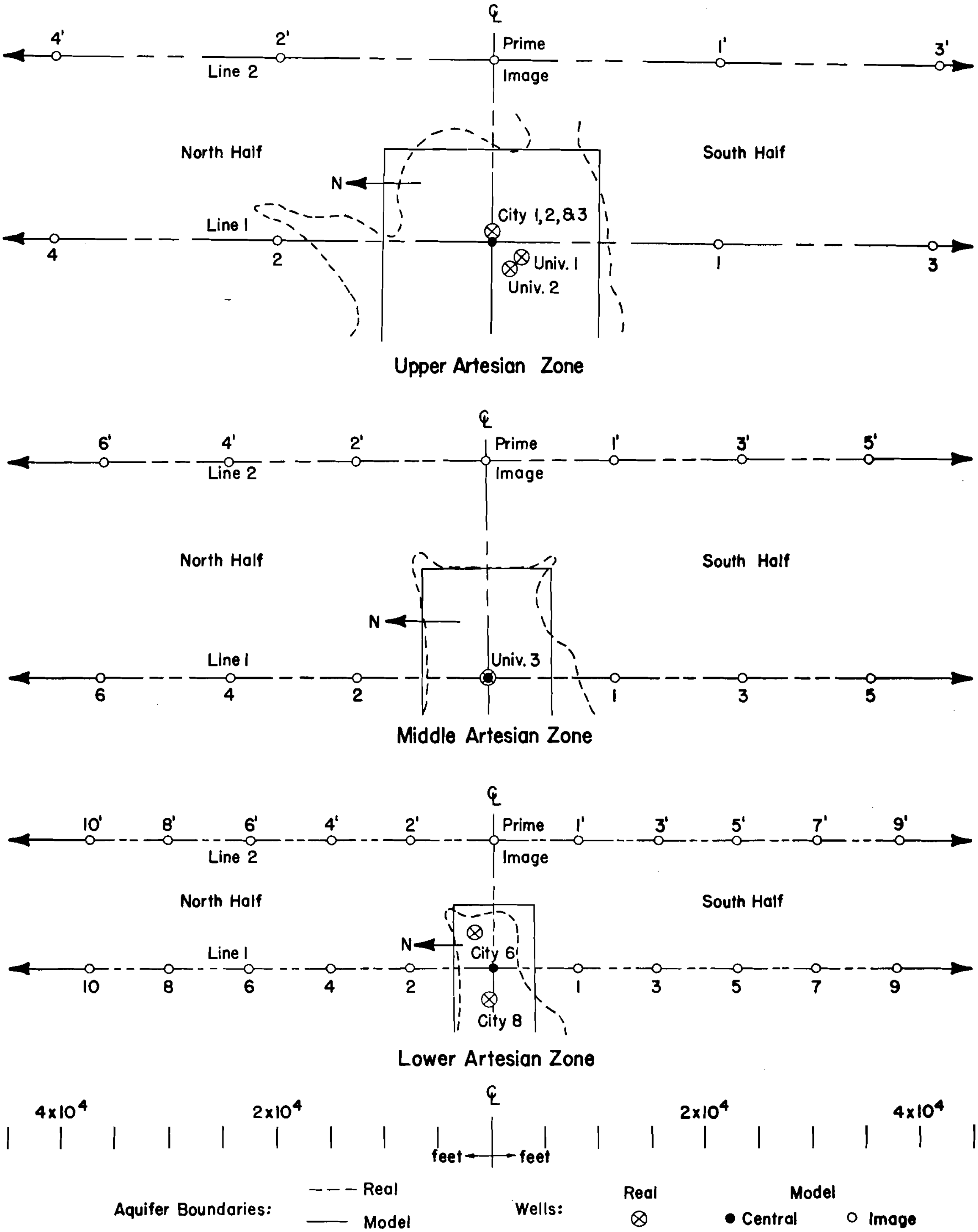


Figure 9. Comparison of real and model artesian aquifers, Moscow basin, Idaho.

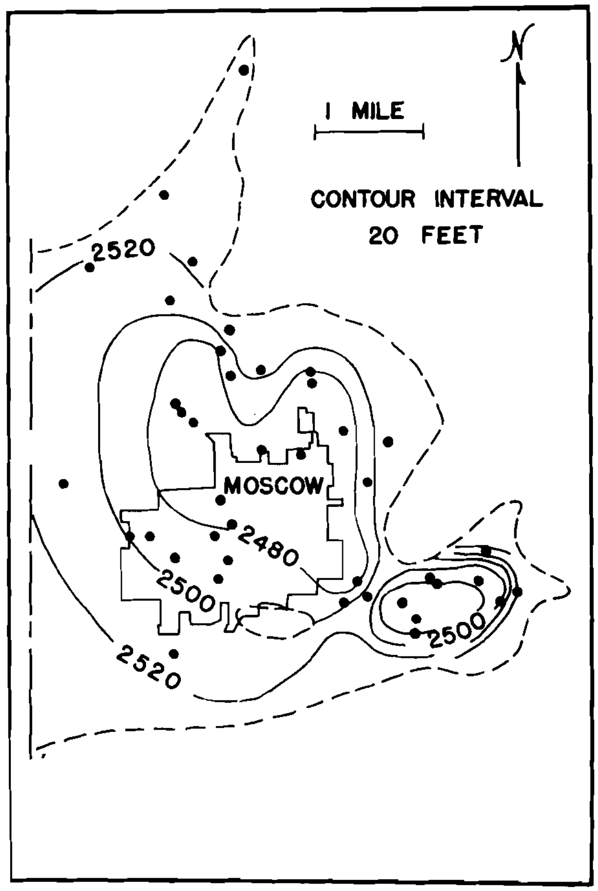


Figure 10. Piezometric surface of the upper artesian zone, Moscow basin, Idaho, as of 1966.

Ross (1965, Table XV) or are on file with the Idaho Bureau of Mines and Geology.

Figure 10 presents our interpretation that, as of 1966, the piezometric surface of the upper artesian zone is dominated by two cones of depression. The smaller cone, in the sub-basin to the southeast, is the result of heavy pumping of two irrigation wells and one industrial well (Parker farm, Elks Golf Course, and Bennett Lumber Products wells).

The larger cone, in the main basin, is shown as closed and is interpreted as a residual cone of depression dating from the period of heavy pumping of the upper artesian zone by the City and the University.

Chang-Lu (1967, Fig. 30) presents a different interpretation of the piezometric surface of the upper artesian zone (his "upper basalt aquifer") that is based on water levels in about 50 wells--many of which must be the same wells as we used in our interpretation. Chang-Lu shows the piezometric surface as open to the west and discusses the control of the shape of the surface by the shape of a buried channel at the base of the "upper basalt aquifer" (loc. cit., p. 70-75). Unfortunately, Chang-Lu did not publish his basic data; therefore, we are unable to determine why his interpretation differs from ours. In our opinion, examination of his map suggests that, if he had continued his 2460, 2480, and 2500 foot contours, they would have closed at about the same general area that we close ours.

In both our map and Chang-Lu's map, the apex of the cone of depression is considerably to the north of the former center of pumpage. Chang-Lu relates this position to the buried channel.

When the upper artesian aquifer was first developed in the 1890's, wells flowed in the vicinity of the Moscow pumping plant. Therefore, the elevation of the piezometric surface originally was at least 2620 feet in the central part of the basin. We have no other data on which to construct a pre-development piezometric map and therefore have no idea what the influence of the buried channel may have been on the pre-1890 piezometric surface. The displacement of the apex of the 1966 cone of depression north of the center of pumpage suggests that the buried channel may exert an influence on the 1966 surface--if so, then it probably also influenced the pre-1890 surface. Our conclusion is that the 1966 residual cone of depression is superimposed on a trough controlled by the buried valley.

Theoretical Effect of Barrier Boundaries on Specific Capacities

Moscow basin artesian aquifers are small and barrier boundary effects should be expected to show up during constant-rate periods in pumping tests, if the rates of expansion of cones of influence are large. Data on the rate of expansion of the cone of influence are available for the lower artesian zone. During the test of March 31, 1965, pumping of City Well 6 caused drawdown to begin in City Well 8, 6,000 feet away, in less than 2 minutes. Rapidity of rate of expansion of the cone of influence of City Well 6 is confirmed from recording instrument records. A recording water-level gage was operated on City 8 from March 15 to April 19, 1965 while a recording voltmeter was operated on the pump controls for City Well 6. A hydrograph of this period was published by Sokol (1966, Fig. 15). Although the records are not as precise as those of the pumping test, they do indicate a time of transmission of cone of influence of 5 minutes or less.

In the model of the lower artesian zone, (Fig. 9), City Well 6 is 2,000 feet from the north barrier boundary, 2,500 feet from the east barrier boundary, and 5,000 feet from the south barrier boundary. City Well 8 is 7,500 feet from both the north and south barrier boundaries and 8,500 feet from the east barrier boundary. If the cone of influence of a well in the lower artesian zone can migrate 6,000 feet in less than 2 minutes, the cone will reach all three barrier boundaries within a very few minutes for any well in the lower artesian zone.

When the rate of expansion of the cone of influence in an aquifer is known, the time after pumping begins at which image wells will influence water levels in an observation well can be predicted by the Law of Times (Walton, 1962, p. 16):

$$\frac{t_1}{r_1^2} = \frac{t_n}{r_n^2}$$

t_1 = time after pumping begins for cone of influence to migrate from pumping well to observation well; in minutes.

r_1 = distance from pumping well to observation well, in feet.

t_n = time after pumping begins for cone of influence of a pumping image well to arrive at observation well; in minutes.

r_n = distance from a pumping image well to observation well, in feet.

Using the data of the March 31, 1965, pumping test in the model of the lower artesian zone, the times after pumping begins for the cones of influence of image wells to arrive at the central well are:

Minutes	Cumulative Number of Wells	Minutes	Cumulative Number of Wells
3.5	2	445.5	47
7.9	3	522.8	51
12.4	7	606.4	55
27.8	11	696.1	59
49.5	15	792.0	63
77.3	19	894.0	67
111.4	23	1002.4	71
151.6	27	1116.8	75
198.0	31	1237.5	79
250.6	35	1364.3	83
309.4	39	1497.4	87
374.3	43		

These data apply to a pumping rate of 900 gpm, but serve to give an idea of the times of arrival of image well effects for other pumping rates. The amount of drawdown caused by each image well at any given time is a function of pumping rate, coefficient of transmissibility, and coefficient of storage. Computations of effects are not particularly useful in the absence of data on the aquifer constants. However, the data do indicate that water levels in wells in the lower artesian zone are under the influence of barrier boundary effects within a few minutes of the beginning of pumping and that the number of image wells affecting the water levels increases at a greater rate in the earlier part of the pumping cycle--within the first 4 to 5 hours after pumping begins. Because the image wells in the model extend to infinity, image well effects should continue to increase as long as the well is pumped continuously.

The middle and upper artesian zones are larger than the lower artesian zone. No data on rate of spread of cone of influence are available for the middle artesian zone, but the performance of University Well 3 suggests rapid spread.

In the upper artesian zone, very little evidence of interference from other wells is present in the records of a recording water-level gage that was operated on University Well 2 between June 25 and July 14, 1964 (Sokol, 1966, Fig. 3). Either rates of expansion of cones of influence in the upper artesian zone are much slower than those of the lower artesian zone, or the aquifer supplying University Well 2 is poorly connected with the aquifers supplying the other wells in the upper artesian zone.

However, it is possible that the rates of expansion of cones of influence in the middle and upper artesian zones are as great as that of the lower artesian zone. The data for the lower artesian zone are regarded as the upper limit of the range of probable values for the rate of expansion of cones of influence in the middle and upper artesian zones.

Sources of Specific Capacity Data

Data on production pumping tests and on daily operation of wells in Moscow basin are on file in the offices of the Moscow City Engineer and of the Physical Plant Division of the University of Idaho. Data on specific capacity are analyzed below for 8 wells; 5 are in the upper artesian zone, 1 is in the middle artesian zone and 2 are in the lower artesian zone. Most of the data are production pumping tests in which drawdowns were determined by airline and pumping rates were measured with an orifice gage.

In most of the tests, pumping rates were varied. Generally, the tests began at rather low rates, then the rate was increased at various times. Very few of the tests were at constant rate in the early part of the test when barrier boundary effects should have been least pronounced. However, the earliest reported values of specific capacities will give the best available indications of the ideal performance of the aquifer under minimum barrier boundary effect.

Barrier boundary effects were not apparent in production pumping tests in the upper and middle artesian zones. Barrier boundary effects are present in one well in the lower artesian zone and may be present in the other.

Specific Capacities of Wells in the Upper Artesian Zone

Detailed production test data are available for University Well 2 and City Well 6 (at its original depth of 250 feet, upper artesian zone). Partial data are available for a special pumping test in City Well 2 and for the production test of the University of Idaho Parker Farm Well. Data are available for daily operation of City Well 3. No barrier boundary effects are evident in the records of tests of these wells. Specific capacity decreases in each well with increase in pumping rate.

University Well 2 obtains most of its water from sands in the upper interbed. During the pumping test of February 24-25, 1951, the well was pumped at 420 gpm. for the first half hour with a drawdown of less than 1 foot. Well loss must have been very low at this pumping rate; the specific capacity should be a very good indication of the ideal performance of the aquifer. In later parts of the test, pumping rates were increased, drawdowns were larger, and specific capacity declined throughout the test. However, at the end of the test, the well was found to have collapsed and filled in to 114 feet above the original well bottom. The decrease in specific capacity during the test could have been related to the well collapse and should not be used to deduce any properties of the aquifer.

City Well 6 was originally drilled to a depth of 250 feet and was tested on January 3-4, 1956. The earliest reported specific capacity was 18.8 gal/ft after 30 minutes of pumping 300 gpm. Pumping rates were increased in steps to 1140 gpm. during the first 5 hours and 32 minutes of the test; specific capacity decreased from 18.8 to 16.5 gal/ft. From 5 hours and 37 minutes to 27 hours, the pumping rate was fairly constant at 1515 to 1585 gpm. (4-1/2 percent range), but the specific capacity increased from 18.0 to 19.8 gal/ft. The increase may have been caused by development of the well. The log of the well indicates that the productive zone (229 to 233 feet) is in basalt. This is the only well in basalt in Moscow basin that shows an increase in specific capacity throughout the production pumping test.

A special pumping test was run on City Well 2 on February 19-20, 1958. The complete record was in the files of the Moscow City Engineer in the spring of 1969, but was missing in June of 1969. Partial data copied from the record show an earliest reported specific capacity of 100 gal/ft after 10 minutes of pumping 300 gpm. At the end of 1 hour of pumping, specific capacity was 55 gal/ft at a pumping rate of 550 gpm. At 8 hours, the pumping rate was 1160 gpm. and, at 24 hours, it was 1120 gpm. Specific capacity at the end of 8 hours and 24 hours was 58 gal/ft. Specific capacity apparently remained constant from 8 to 24 hours of pumping, indicating an absence of barrier boundary effects during this part of the test.

Detailed records are not available for the pumping test of the University of Idaho's Parker Farm well (15bc1). The log of the well in the University Engineer's office states that drawdown at the end of 24 hours of pumping 1000 gpm. was 21 feet; the specific capacity at that time was 47 gal/ft.

Daily operation records of City Well 3 in the files of the Moscow City Engineer's office indicate drawdowns of 8 to 12 feet at the end of normal days of intermittent pumping 1350 gpm. Specific capacity is placed at 135 gal/ft after 8 hours of pumping.

Specific Capacity of a Well in the Middle Artesian Zone

University Well 3 was tested September 5-6, 1963. The earliest reported specific capacity is 500 gal/ft after 30 minutes of pumping 1000 gpm. At the end of 1 hour of pumping 1000 gpm, specific capacity was still 500 gal/ft.

Pumping rate was increased at about 3 hour intervals throughout the test. The performance of the well is remarkable; drawdown was only 2 feet after $6\frac{1}{2}$ hours of pumping 1000 to 1200 gpm. and only 6 feet at the end of the 24 hour test with the last $4\frac{1}{2}$ hours being at a pumping rate of 2185 to 2197 gpm. Specific capacity at the end of 24 hours was 365 gal/ft.

Apparent initial increase of specific capacity after each increase of pumping rate probably was the result of use of airline to determine water level measurements. The airline apparently lagged somewhat behind the change in water level. Shortly after each increase in pumping rate, specific capacity declined, but then remained constant for the remainder to the 3-hour constant rate period. Decrease in specific capacity seems to be related only to increase in discharge. Barrier boundary effect would be very difficult to detect in a well with such a high specific capacity when an airline is used to measure water levels; airlines are accurate only to about 1 foot.

Specific Capacities of Wells in the Lower Artesian Zone

City Well 6 was tested several times when it was deepened to the lower artesian zone. The final test was on April 27-28, 1960.

The record of the test does not state the exact time that pumping began. The first useful figures on specific capacity are 29.3 gal/ft at 1532 gpm. and 42.5 gal/ft at 1660 gpm. From 11:25 p.m. on April 27 until 10:00 a.m. on April 28, the pumping rate was relatively constant, fluctuating between 1660 and 1710 gpm. (6 percent). During this period of 10 hours and 35 minutes, specific capacity decreased from 42.5 gal/ft at the beginning to 34.6 gal/ft at the end. During the next 9 hours of the test, the pumping rate was nearly constant at 1620 gpm. and specific capacity was constant at 35.2 gal/ft. Because the decrease in specific capacity took place during a constant-rate period that must have been fairly early in the test, it may have been partly the result of barrier boundary effect. The long period of constant specific capacity later in the test could have been because fewer image wells become effective with increasing time.

On March 29-31, 1965, we ran a pumping test using City Well 6 as the pumping well and City Well 8 as the observation well. In preparation for the test, City Well 6 was not operated from 7:30 p.m. on March 30 until 6:27 a.m. on March 31. After the test began, the specific capacity of City Well 6 was 53.5 gal/ft after 8 minutes of pumping 940 gpm. and 50.5 gal/ft after 13 minutes of pumping, the rate having reduced to 910 gpm. Because the pump was operated intermittently in later parts of the test, the rest of the specific capacity data are not useful.

City Well 8 was tested December 11-12, 1964. The earliest reported figure on specific capacity is 36.4 gal/ft after 30 minutes of pumping 1090 gpm. At the end of 1 hour of pumping 1090 gpm, specific capacity had decreased to 34.1 gal/ft. The pumping rate of 1090 gpm. was maintained for the first

7 hours of the test. Specific capacity decreased steadily for the first $4\frac{1}{2}$ hours, reaching 31.2 gal/ft; during the remaining $2\frac{1}{2}$ hours, specific capacity was constant. From 7 to 13 hours, pumping rate was 1300 gpm; specific capacity was essentially constant at 25.1 gal/ft. Pumping rate was reduced to 1200 gpm from 13 to 17 hours; specific capacity was essentially constant at 26.1 gal/ft. Pumping rate was reduced to 1093 gpm. from 17 to 21 hours; specific capacity increased from 29.6 gal/ft at $17\frac{1}{2}$ hours to 31.3 gal/ft at $18\frac{1}{2}$ hours, but then remained constant for the remainder of the period. Pumping rate was then reduced to 1000 gpm. from 21 to 24 hours; specific capacity increased from 33.3 gal/ft at 21 hours and 5 minutes to 34.5 gal/ft at $21\frac{1}{2}$ hours, then remained constant for the rest of the test. The decrease in specific capacity during the first $4\frac{1}{2}$ hours of constant-rate pumping probably is the result of barrier boundary effect. The increases in specific capacity following some of the reductions of pumping rate in later parts of the test may also be the result of decreases in barrier boundary effects.

Aquifer Constants

In the absence of suitable pump test data, the properties of the artesian aquifers must be estimated. Estimates are based on methods of indirect calculation from the literature, on data reported from similar aquifers elsewhere, and on studies of mathematical models of the Moscow basin aquifers.

Coefficient of Storage

According to Ferris and others (1962, p. 76), the coefficient of storage of artesian aquifers generally ranges from 1.0×10^{-3} to 1.0×10^{-5}. Other authors report higher values from basalt aquifers of the Pacific Northwest. Values for Snake Plain basalt range from 1.0×10^{-1} to 1.0×10^{-3} (Walton and Steward, 1959; Mundorff, Crosthwaite, and Kilburn, 1964). However, Snake Plains basalts differ from Columbia River basalts in that they originated from central eruptions rather than fissure eruptions and are characterized by pahoehoe lavas and lava tubes that are absent in the Columbia River basalts. Because primary permeability is a function of origin in volcanic rocks, Columbia River basalts may not have the same sort of aquifer characteristics as Snake River basalts.

Nevertheless, high values of coefficient of storage have been reported from Columbia River basalts. Eddy (1969) reports values of 1.0×10^{-2} to 1.0×10^{-3} for the basalts in the central Columbia Basin Project of Washington. According to D.A. Myers (oral communication), U.S. Geological Survey personnel working in the Columbia River basalt consider 2.3×10^{-3} to be the best average value of coefficient of storage. Price (1961) reported a value of 2.0×10^{-4} for a well in Columbia River basalt near Walla Walla, Washington.

We have no direct data on coefficient of storage of Columbia River basalt in the Moscow basin artesian aquifers. Our studies of mathematical models that are discussed below show that results that most closely match the actual performance of the real aquifers are obtained when the values used are:

Upper artesian zone	2.3×10^{-3} to 3.0×10^{-4}
Middle artesian zone	2.3×10^{-3}
Lower artesian zone	1.0×10^{-3}

Although these values are high for artesian aquifers, they are in the range that has been reported from basalt aquifers elsewhere. The aquifer models assume that no recharge is taking place; if recharge is actually taking place in the real aquifer, one of the effects would be model coefficients of storage that are higher than the real value.

Coefficient of Transmissibility

Several graphs for estimating coefficient of transmissibility from specific capacity are in the literature. We used the graph of Walton (1962, p. 12-13) because specific capacities of some Moscow basin wells are so great as to be off scale on the graphs of Myer (in Bentall, 1963, p. 338-340) and Hurr (1966). Data are shown on Table 4. Where results could be checked between the several graphs, Walton and Myer corresponded closely whereas Hurr tended to give values about 50 percent lower. The method of Hurr seems to be more applicable to water-table aquifers than to artesian aquifers.

Because of possible image-well effects and because of the common practice of increasing pumping rate during production pumping tests in Moscow basin, we used the earliest reported specific capacities whenever possible. We could not avoid using several 1-hour figures as well as an 8-hour and a 24-hour figure. Most of the data probably are influenced by barrier boundary effect and well loss and should be regarded as minimum values. The data indicate a considerable range of values--a relation that is common in basalt aquifers.

The value of 1.0×10^6 gal/day/ft for University Well 2 is considered a very good figure because of the low pumping rate (and low well loss) and because of its early time in the test. The 24-hour specific capacity of 47 gal/ft for the Parker Farm well does not differ much from the 24-hour specific capacity of 58 gal/ft for City Well 2.

The effect of full penetration of Moscow basin aquifers is shown by the difference between the data for City Well 6 at 968 feet and at 1305 feet. At their final depths, City Well 6 is nearly fully penetrating and City Well 8 does penetrate fully; specific capacities and transmissibilities of the two wells are very similar.

A wide range of values of coefficient of transmissibility has been reported from basalt aquifers elsewhere. Typical upper values for Snake Plain basalt range from 1.0×10^4 to 1.0×10^7 gal/day/ft (Walton and Steward, 1959; Mundorff, Crosthwaite, and Kilburn, 1964). Upper values for Columbia River basalts are less well known. Data on permeabilities from Eddy (1969), when recalculated as transmissibilities for conditions similar to Moscow basin, indicate a value of 1.0×10^4 gal/day/ft, a figure that seems small for Moscow basin. Foxworthy and Bryant (1967) report a coefficient of transmissibility of 3.2×10^5 gal/day/ft as determined by recovery methods for a well in Columbia River basalt near The Dalles, Oregon. Specific capacity of the well indicated a value of 1.0×10^6 gal/day/ft. Price (1961) reported a coefficient of transmissibility of 3.7-4.0×10^5 gal/day/ft and a coefficient of storage of 2.0×10^{-4} for a well in Columbia River basalt near Walla Walla, Washington.

In our studies of mathematical models of Moscow basin aquifers that are discussed below, coefficients of transmissibility that produced results that are closest to the actual performances of the real aquifers are:

Upper artesian zone	1.0×10^6 gal/day/ft
Middle artesian zone	1.0×10^6 gal/day/ft
Lower artesian zone	4.0×10^7 gal/day/ft

These values are all very large, but are within the upper range of indicated values from specific capacity for the upper and middle artesian zones. The value for the lower artesian zone is much higher than the specific capacity value and is higher than any value we have seen reported in the literature for basalt. The model aquifers assume that no recharge takes place; the effect of actual recharge to the real aquifer would show up as values of model aquifer constants that are higher than the real value.

Table 4. Coefficients of Transmissibility Estimated from Specific Capacities of Wells, Upper Artesian Zone, Moscow Basin, Idaho

Well	Time Since Pumping Began Hours	Minutes	Specific Capacity gal/ft	Coefficient of Transmissibility gal/day/ft
Upper artesian zone				
University 2	0	15	420	1.0×10^6
Parker Farm	24	0	47	1.0×10^5
City 2	0	10	300	2.0×10^5
City 3	8	00	135	4.0×10^5
City 6 1/	1	0	17	3.0×10^4
Middle artesian zone				
University 3	1	0	500	1.0×10^6
Lower artesian zone				
City 6 2/	0	10	11	2.0×10^4
City 6 3/	0	10	53	1.0×10^5
City 8	1	0	34	8.0×10^4

1/ Original well, 250 feet deep.
2/ After deepening to 968 feet.
3/ Final depth to 1305 feet.

Relations of Coefficient of Transmissibility to Coefficient of Storage in the Lower Artesian Aquifer

The expansion of the cone of influence of City Well 6 through 6000 feet to City Well 8 in less than 2 minutes can be placed into a set of curves prepared by Theis (1952). The graph shows that expansion at this rate will take place when the ratio $S/T = 2.5 \times 10^{-11}$. Unfortunately, the data from the test do not permit the use of the graph to determine S and T directly.

However, the applicability of various values of S and T can be checked by substitution into the ratio. Substitution of the "normal" range of coefficient of storage for artesian aquifers results in:

S	T
1.0×10^{-3}	4.0×10^{7}
1.0×10^{-5}	4.0×10^{5}

Substitution of the range of values of coefficient of storage reported for Columbia River basalts results in:

S	T
1.0×10^{-2}	4.0×10^{8}
1.0×10^{-3}	4.0×10^{7}

Substitution of the coefficient of transmissibility estimated from specific capacity results in:

S	T
2.3×10^{-6}	8.0×10^{4}

Substitution of some of the coefficients of transmissibility reported from Columbia River and Snake River basalts results in:

S	T
2.5×10^{-4}	1.0×10^{7}
2.0×10^{-5}	4.0×10^{5}
2.5×10^{-7}	1.0×10^{4}

The values $S = 1.0 \times 10^{-3}$ and $T = 4.0 \times 10^{7}$ appear in two of these computations, and are the values that produced results in the model aquifers that most closely matched the actual performance of the real lower artesian zone. These values seem to fit the lower artesian zone very well, but only under the condition that no recharge takes place. The value of coefficient of transmissibility is higher than any we have seen reported in the literature.

MATHEMATICAL MODEL AQUIFER STUDIES

We designed mathematical models of the artesian aquifers of Moscow basin for the purposes of:

1. Testing the hypothesis that recharge to Moscow basin is very small (Crosby and Chatters, 1965).

2. Predicting the depths to water in the year 2000 assuming that the aquifers would continue to perform in the future as they did in the past.

3. Predicting the seasonal increase in water levels of the upper artesian zone under artificial recharge.

Factors involved in mathematical model aquifer studies include size and shape of aquifer, recharge, discharge, time, changes in water level, and the aquifer constants (coefficient of transmissibility and coefficient of storage). Commonly, model studies are used to predict future water level trends in aquifers of known size and shape when recharge, discharge, and aquifer constants are known or can be estimated reliably. In Moscow basin, size and shape of aquifers are fairly well known, but reliable data are lacking on recharge and only partial data are available on discharge and on aquifer constants. Our approach was to postulate that no recharge takes place, then to assume a range of values of aquifer constants and test the assumed values in models that were pumped at the same rates and for the same lengths of time as the actual aquifers were pumped. We assumed that models that most closely reproduced the actual drawdown of the real aquifer have aquifer constants that are close to the values for the real aquifer. We then used the best-fit values of aquifer constants to predict future drawdowns, to estimate water in storage, and to predict the effects of artificial recharge.

We chose to study a no-recharge model because it should give results that are reasonably close to those of a model that assumes recharge which is small compared with pumpage. Previous investigators have suggested that recharge in Moscow basin is less than pumpage; Crosby and Chatters (1965, p. 16) estimate the recharge to be only one-tenth of the 1965 pumpage. Therefore, the no-recharge model provides information on the probable performance of the aquifers under one set of conditions that may apply to Moscow basin. Model aquifers that assume recharge are much more complex than the no-recharge models; we have not attempted any studies of models that include recharge.

A preliminary report has been published on the results of the no-recharge model studies (Jones and Ross, 1969). The complete study and some revisions of the results are given below. A preliminary study also has been published on the artificial recharge studies (Jones, Ross, and Williams, 1968); however, this study was completed prior to the pumping model study. New knowledge on the aquifer constants that were obtained in the pumping model study require that we revise our estimates of buildup of the piezometric surface under artificial recharge. The revisions are given below.

Descriptions of Moscow Basin Model Aquifers

The three artesian zones of Moscow basin were generalized into two-dimensional models by using the contact of the zone with the crystalline basement as the aquifer boundary (Fig. 9). Because the production from the upper artesian zone comes largely from the upper portion of the zone, elevation 2300 feet is considered the average position of the boundary. Although current production from the middle and lower artesian zones does come from the lower part of each zone, we do not yet know enough about the zones to state that only the lower parts are productive. Therefore, elevation of the contact of the middle of each of these zones is used for the average position of the boundary: 1800 feet for the middle artesian zone and 1400 feet for the lower artesian zone.

The two-dimensional models were further generalized by treating the contact with the crystalline basement as a barrier boundary and generalizing the boundary as three straight lines intersecting at right angles. This model is discussed by Ferris and others (1962, p. 156-159). The model differs from the real aquifer in that the model has no western boundary. In a later section, we will explain why it is not necessary to study a four boundary model of Moscow basin.

To simplify calculations, all pumpage in the model of each zone was assigned to a single central well which is equidistant from the north and south boundaries. Distance of the central well from the east boundary depends on the positions of the real wells in each zone. The central well is about midway between the real well positions in the upper and lower artesian zones and is at the actual position of the real well in the middle artesian zone. The image well arrays are constructed for the central well and continue outwards to infinity. In many calculations, image wells as much as several million feet from the central pumped well were used.

Preliminary calculations indicated that little error was introduced by using a single central well rather than using several pumping wells at the actual positions of the real wells. Use of a single well reduces the number of image wells and the number of calculations needed, as well as the difficulties of interpretating the model. The models used are shown in Figure 9.

Calculations Used in Model Studies

We assumed a range of values of model aquifer constants that includes coefficients of transmissibility estimated from specific capacities of Moscow basin wells (Tab. 4) and values of coefficients of transmissibility and coefficient of storage that have been reported from basalt aquifers elsewhere. We tested the values in model aquifers that were pumped for the same lengths of time and at the same rates as the real aquifers had been pumped.

Average annual pumpage for each aquifer was recalculated as continuous pumpage, in gallons per minute, of the single central well. Length of time of the pumping periods was calculated in days.

Using the modified nonequilibrium equation, the following were computed for the end of each period of pumping:

1. Infinite-aquifer drawdown at a distance of 1 foot from the pumped well (time-drawdown equation);

2. Slope of the cone of influence (Δ s; distance-drawdown equation);

3. Radius of the cone of influence (r_0). A sample calculation is shown in Table 5.

Theory of image wells shows that the effects of barrier boundaries will be the equivalent of the sum of the additional drawdowns caused by the interference of every pumping image well within the distance r_0 at the end of the pumping period. The theoretical drawdown of the central well at the end of the pumping period is the drawdown of the central well itself, computed as though it were in an infinite aquifer, plus the sum of the drawdowns in the central well caused by interference from the image wells, well losses being neglected.

The amount of interference caused by each image well was computed from a plot of depth to water--depth to the cone of influence--against log of distance from the central well. On semilog paper, a cone of influence plots as a straight line that has its origin at r_0 and has a slope of Δs per log cycle. When positions of the image wells with respect to the central well are plotted, the graph shows the drawdown in each image well caused by pumping of the central well. Each image well has the same effect on the central well as the central well has on each image well; therefore, the drawdown caused in the image well by the central well is the same as the drawdown caused in the central well by the image well. A sample plot is shown in Figure 11.

In all three model aquifers, image well drawdowns were calculated only for the north half of line 1 (Fig. 9; the line that passes through the central well). Differences between the effects of the image wells on line 1 and the effects of the corresponding wells on line 2 (Fig. 9; the line east of the east barrier boundary) are so small that they can be neglected, especially for wells more than 1.0×10^4 feet from the central well. Accordingly, the model may be considered to consist of 4 half-lines of image wells, all of which have the same effect on the central well, plus the prime image well on line 2 directly east of the central well, plus the central well itself. Total drawdown at the end of the pumping period is the sum of the effect of the north half of line 1 multiplied by 4 plus the drawdowns of the prime image well and the central well.

Drawdowns of individual image wells were computed only for the wells within 1.0×10^5 feet of the central well. Beyond this distance, differences between drawdowns of the individual wells in each log cycle of distance are small. Computations were shortened by taking the average drawdown over each log cycle of distance and multiplying it by the number of image wells in the log cycle. Some of the log cycles contain as many as 50 image wells; therefore the procedure saved considerable time. In some computations, drawdowns were computed as the average drawdowns over one-tenth of a log cycle. Image well effects were computed out to distance r_0 except that the last partial log cycle commonly was omitted when the effect of the wells in the interval would be very small.

Use of Model Aquifer Data to Predict Drawdown or Buildup of Real Aquifers

The equations used in the calculations (Tab. 5) show that, in a series of model studies that use the same length of time and the same pumping rate, the time-drawdown of the central well is related to:

$$\frac{\log_{10} T:S}{T}$$

and the distance to the edge of the cone of influence is related to:

$$\sqrt{T:s}$$

In our computations, more than one set of values of T and S produced model drawdowns that matched real aquifer drawdowns during the same pumping period. The results of the computations probably depend more on the ratio T:S than they do the actual values of T and S. Assuming that the aquifer will continue to perform in the future as it has in the past, any set of values of T and S that produced the same drawdown in the model aquifer as in the real aquifer can be used to predict future drawdowns in the real aquifer.

The values of aquifer constants obtained from the model studies were used to predict the water levels in the three artesian zones at the year 2000 at a pumping rate that would meet the anticipated demands. Total water in usable storage as of 1965 was estimated from model studies in which pumping rates were increased until the water levels were drawn down to critical depths. Critical depth was defined as the depth at which dewatering of the aquifer is about to begin and was calculated as depth to the top of the producing zone minus an allowance for well loss, interference between wells and intermittent pumping of wells. The water in storage was calculated by multiplying the pumping rate by the time necessary to bring the water level down to the critical depth.

Mathematical models were also used to predict the buildup of piezometric surface that would result from artificial recharge of the upper artesian zone. Values of aquifer constants for the pumping model studies of the upper artesian zone that produced the closest match of real aquifer drawdown were used in the computations as well as other values of aquifer constants that include coefficient of transmissibility estimated from specific capacities of wells and coefficients of storage selected from the "normal" range for artesian aquifers. The models were used to predict buildup at the City Pumping Plant at the end of artificially recharging for 100 days at the rates of 1000 and 2000 gallons per minute.

Pumping of the Upper Artesian Zone

The longest period of record of pumpage and water levels is for the upper artesian zone. In 1948, the City of Moscow began to keep daily records of water pumped and of static and pumping water levels. The University of Idaho began to keep similar records in 1954. A few data are available prior to 1948. Detailed hydrographs of the City and University wells in the upper artesian zone are shown in Figure 17; the hydrograph of an observation well (7ddl) is shown in Figure 15; a generalized hydrograph of water levels in the upper artesian zone in the vicinity of the City Pumping Plant is shown in Figure 12. Data are on file with the Idaho Bureau of Mines and Geology.

In Figure 12, the generalized, composite hydrograph has been smoothed into straight-line segments and plotted against the logarithm of time. The plot

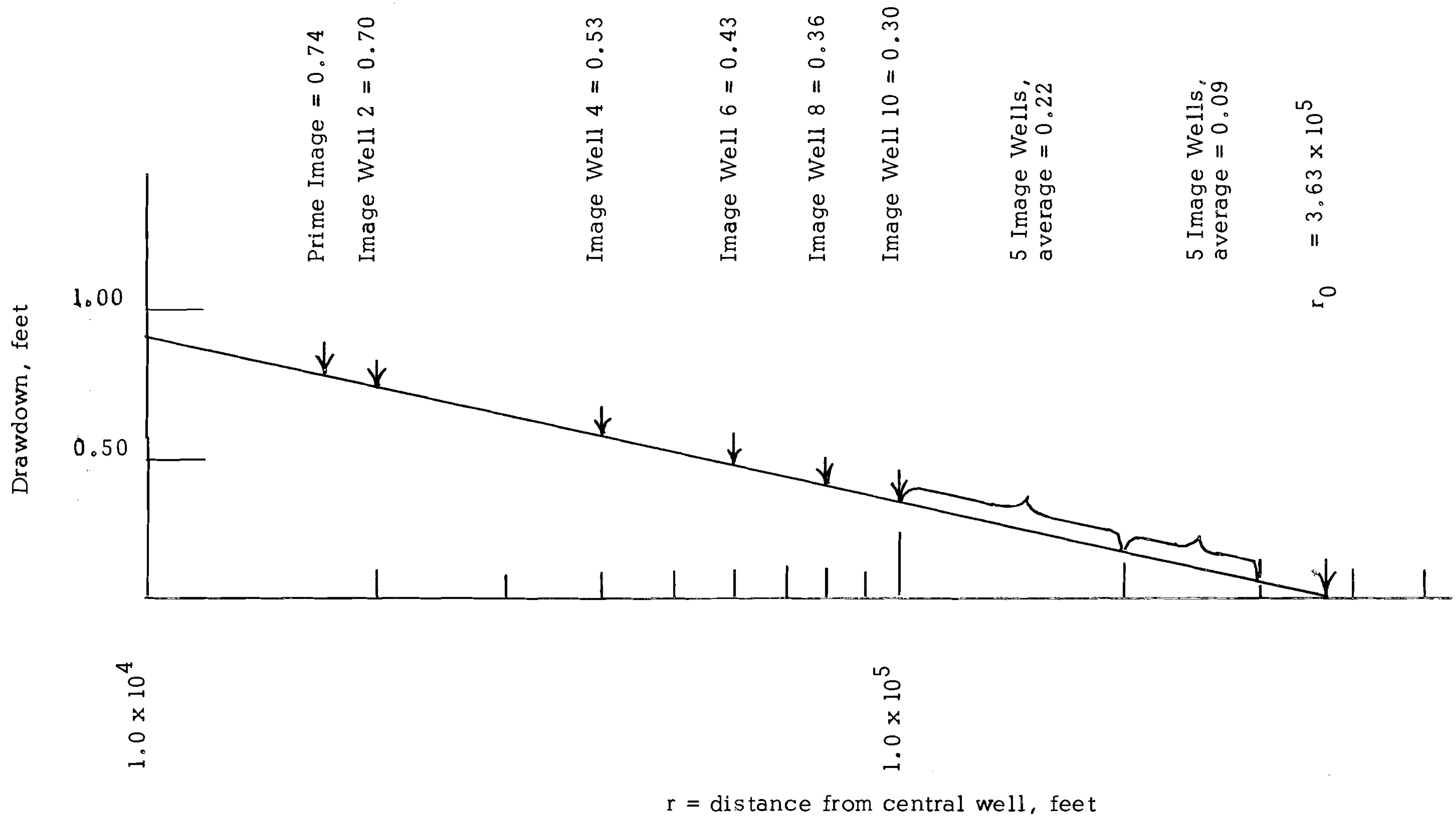


Figure 11. Sample data plot for mathematical model aquifer study, Moscow basin, Idaho. Data shown are for the upper artesian zone, 1949-1960, with coefficient of storage of 1.0×10^{-2} and coefficient of transmissibility of 1.0×10^{6} gal/day/ft.

Table 5. Sample Computation for Mathematical Model Aquifer Study, Moscow Basin, Idaho

The sample is for the upper artesian zone.

Computation:

Predicted drawdown, 1949-1960,

Using $S = 1.0 \times 10^{-2}$ and $T = 1.0 \times 10^{6}$

Well or No. of Wells in Interval	r	Average Drawdown over Interval	Draw-down
2	2.0×10^{4}		0.70
4	4.0		0.53
6	6.0		0.43
8	8.0		0.36
10	10.0		0.30
(5)	$1.0\text{-}2.0 \times 10^{5}$	0.22	1.10
(5)	2.0-3.0	0.09	0.45
(5)	3.0-4.0		
(5)	4.0-5.0		
(5)	5.0-6.0		
(5)	6.0-7.0		
(5)	7.0-8.0		
(5)	8.0-9.0		
(5)	9.0-10.0		
(50)	$1.0\text{-}2.0 \times 10^{6}$		
(50)	2.0-3.0		
(50)	3.0-4.0		
(50)	4.0-5.0		
(50)	5.0-6.0		

Data:

$T = 1.0 \times 10^{6}$ $\qquad S = 1.0 \times 10^{-2}$

$t = 4.4 \times 10^{3}$ $\qquad Q = 1060$ gpm.

Time-Drawdown for Central Well:

$$s = \frac{264Q}{T} \left(\log_{10} \frac{0.3Tt}{r^2 S} \right)$$

$$= \frac{264 \times 1060}{1.0 \times 106} \left(\log_{10} \frac{0.3 \times 1.0 \times 10^{6} \times 4.4 \times 10^{3}}{1 \times 1.0 \times 10^{-2}} \right)$$

$$= 3.12$$

Distance-Drawdown of Image Wells:

$$\Delta s = \frac{528Q}{T} = \frac{528 \times 1060}{1.0 \times 10^{6}} = 0.56$$

Distance to Edge of Cone of Influence:

$$r_0 = \sqrt{\frac{0.3Tt}{S}} = \sqrt{\frac{0.3 \times 1.0 \times 10^{6} \times 4.4 \times 10^{3}}{1.0 \times 10^{-2}}}$$

$$= 3.63 \times 10^{5}$$

Total, north half line 1		3.87
		x4
Total, four half lines		15.48
Prime Image	1.7×10^{4}	0.74
Central Well		3.12
Total for Model		19.34
Depth to Water at Start		23.27
Depth to Water at End		42.61

shows three periods of more-or-less constant rate of decline of water levels: 1895-1925, 1925-1949, and 1949-1960. The changes in rate seem to be related to major improvements in the City of Moscow water system, such as the drilling of City Well 3 in the late 1920's and the installation of large-diameter mains in the late 1940's. It should be noted that this composite drawdown is based on water levels in several wells that mutually interfere with each other.

Drawdown data used in model aquifer calculations were picked from Figure 12 and are:

Time Interval	Drawdown in Feet Increment	Total
1895-1925	45	--
1925-1949	40	85
1949-1960	40	125

The model aquifer of the upper artesian zone was studied for each of the three periods. Pumping data for 1949-1960 are largely from actual records; data for 1925-1949 and 1895-1925 were deduced from indirect evidence, such as sales of water.

Pumping and Water Level Data, 1895-1960

Records of annual pumpage for the City of Moscow and the University of Idaho for 1955-1960 are shown on Table 6a. During this period, University pumpage averaged 38 percent of City pumpage. Records of annual pumpage for the City of Moscow for 1948-1954 are shown on Table 6b. Pumpage for the University of Idaho shown on Table 6b was computed at 38 percent of City pumpage. Average annual total pumpage for the period 1948-1960 is 552,136,000 gallons. This is the equivalent of a single well pumping 1060 gallons per minute continuously for one year.

Records of sales of water by the City of Moscow for 1948 through 1957 are shown on Table 6a and b. Records of sales were furnished by the Moscow City Clerk. Computations on the relations of sales to City pumpage, shown on Table 7a, show that for 1948 through 1957, gallons pumped equals sales in dollars divided by the intermediate water rate of $0.22 per thousand gallons. Table 7b shows pumpage for 1940 through 1947 calculated by dividing annual sales in dollars by the then prevailing intermediate rate of $0.15 per thousand gallons. Data are not available on sales prior to 1940. The years 1940 through 1945 were selected as base years for the 1925-1949 period, and the average annual pumpage as calculated from 1940-1945 sales of water is used as the average annual pumpage of the entire period. The pumpage is 310,104,000 gallons, which is the equivalent of a single well pumping 590 gallons per minute continuously for a year. Water-level measurements taken in City Well 1 and in the observation well (7ddl) at irregular intervals are available over most of the period.

Prior to 1929, few data are available. Laney and others (1923, p. 11) computed the 1923 pumpage at 230,000,000 gallons based on 12 times the Agust, 1923, consumption of water by the City of Moscow (15 million gallons, including an allowance for leakage). We believe this figure is too large because August consumption is at a much higher rate than the average annual consumption. During the period 1955-1959, August pumpage for the City of Moscow averaged 11½ percent of average annual pumpage, rather than the 8 percent assumed by Laney and others. Recalculation using the 11½ percent figure indicates that the annual pumpage of the City of Moscow in 1923 was 130,000,000 gallons. Adding the 1919 pumpage for University of Idaho (Laney and others, 1923, p. 11), the total is 152,000,000 gallons. This is the equivalent of a single well pumping 290 gallons per minute for one year. Only two water levels were reported during this period for City Well 1; these two levels were "close to the surface" in 1895, and 44 feet from the surface in 1923 (Laney and others, op, cit., p. 6).

Table 6. Water Pumped from the Upper Artesian Zone, 1948-1960

a. 1955-1960; data from flow meters on both City and University wells.

Pumpage, Millions of Gallons				City Sales	
Year	City	University	Total	Dollars	Gals./Dol.
1960	334.6	153.8	488.4		
1959	432.2	178.9	602.1		
1958	372.0	190.0	562.0		
1957	353.2	191.2	548.4	95,100	4,150
1956	298.7	143.3	442.0	89,100	4,380
1955	530.9	159.8	690.7	77,700	6,410

b. 1948-1954; data from flow meters on City wells and estimated for University wells by University Pumpage = 0.38 City Pumpage.

Pumpage, Millions of Gallons				City Sales	
Year	City	University	Total	Dollars	Gals./Dol.
1954	398.6	150.0	548.6	82,000	4,580
1953	390.5	147.0	537.5	83,600	4,390
1952	383.3	145.0	528.3	83,500	4,350
1951	410.5	155.0	565.0	111,800	3,450
1950	436.9	165.0	620.9	82,800	4,350
1949	474.6	179.0	653.6	84,000	4,350
1948	350.0	132.0	482.0		

Model Aquifer Drawdown, 1896-1960

We ran two sets of drawdown computations for the upper artesian zone for the period 1896-1960. In the first set, we tested a broad range of various combinations of possible values of T and S to see which ones resulted in model aquifer drawdowns that most closely approached the real aquifer drawdowns for each of the three periods of approximately equal rate of water-level decline. In the second series of computations, we held T constant at one of the values that produced matching results in the first series of computations and varied S through small intervals until we obtained a value of model aquifer drawdown that most closely approached the real aquifer drawdown during each of the three pumping periods.

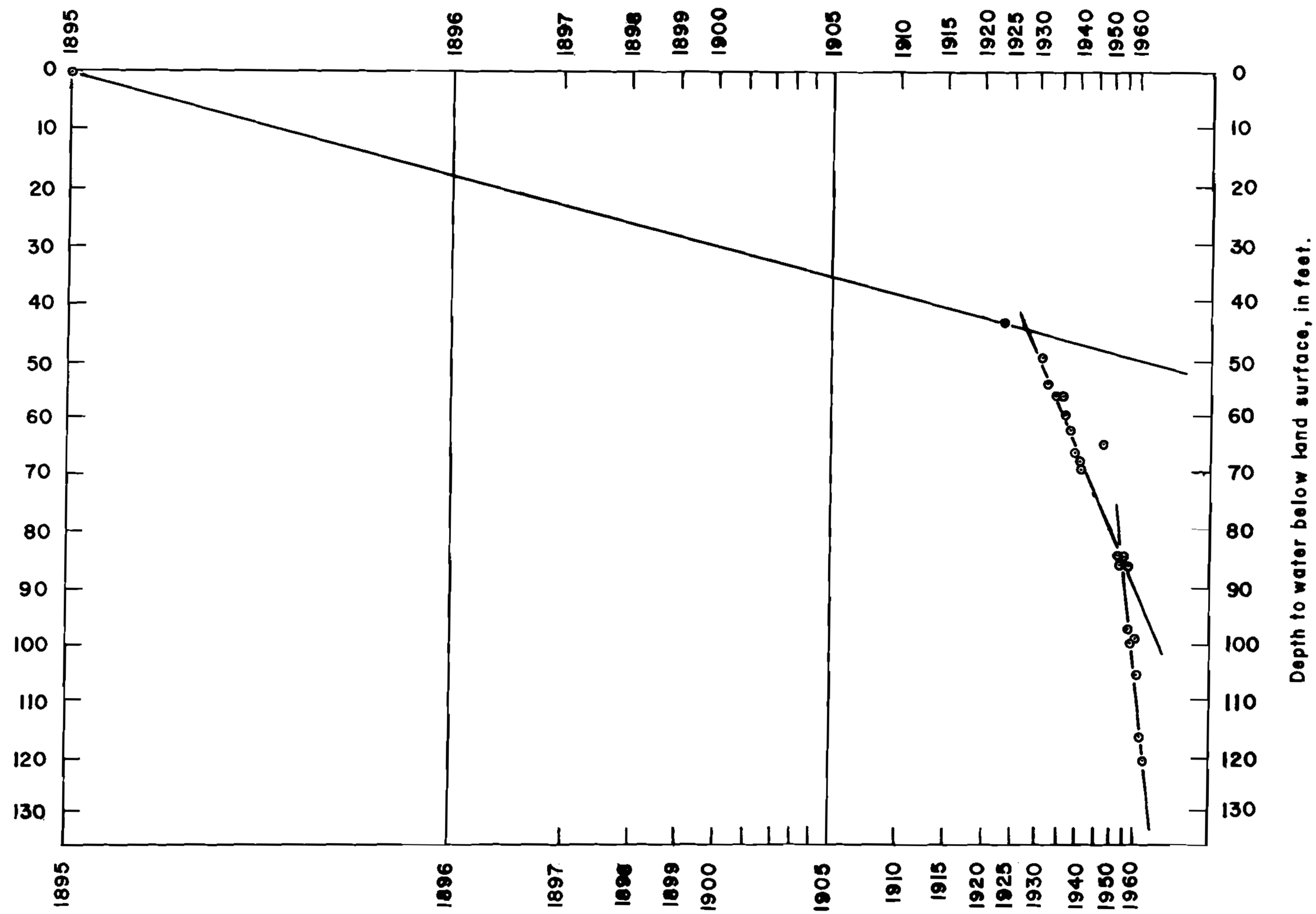


Figure 12. Generalized composite water level trends, Upper artesian zone, 1895–1960, Moscow basin, Idaho. Lowest annual water levels of City of Moscow pumping plant are plotted against logarithim of time in years.

Table 7. Water Pumped from the Upper Artesian zone, 1940- 1947

a. Computations for estimating City pumpage from City Sales

Average sales in dollars, 1948-1957 (From Tab. 6)

4,490 gallons/dollar
$0.22 dollars/thousand gallons

Water Rates

Cost per Thousand Gallons		
1948 1957	1895 1947	
$0.30	$0.20	4,000 to 10,000 gallons per month
0.22	0.15	10,000 to 20,000 gallons per month
0.16	0.12	over 20,000 gallons per month

Relation of Pumpage to Sales

$$\text{Pumpage 1948-1957} = \frac{\text{Dollars sold}}{0.22}$$

$$\text{Pumpage 1940-1947} = \frac{\text{Dollars sold}}{0.15}$$

b. Estimated Pumpage; City from Dollars Sold/0.15; University from 0.38 of City Pumpage

	City		University	Total
Year	Dollars Sold	Millions of gals.	Millions of gals.	Millions of gals.
1947	48,678	324.5	122.5	447.0
1946	41,545	277.0	104.5	381.5
1945	35,065	233.8	88.2	322.0
1944	34,712	231.4	87.3	318.7
1943	31,708	211.4	79.8	291.2
1942	35,105	234.0	88.3	322.3
1941	34,953	233.2	87.6	320.8
1940	38,242	254.9	96.2	351.1

Table 8. Results of Studies of Models of Upper Artesian Zone Using Different Values of Model Aquifer Constants, 1896-1960.

A. First series of computations: Determining values of model drawdowns for each of the three pumping periods using a range of probable values of T and S.

1949-1960; actual increment of drawdown over period, 40 feet; cumulative drawdown at end of period, 125 feet.

Drawdown Increment / Cumulative Drawdown

T gal/day/ft	S = 1.0×10^{-2}	2.3×10^{-3}	1.0×10^{-3}	1.0×10^{-4}
1.0×10^{6}	19.34 / 42.61	40.78 / 92.56	58.21 / 197.95	119.75 / 337.71
4.0×10^{5}	33.30 / 67.03	64.60 / 141,78	127.39 / 221.64	293.30 / 667.52
2.0×10^{5}	50.4 / 106.7	100.1 / 208.3		

1925-1949; actual imcrement of drawdown over period, 40 feet; cumulative drawdown at end of period, 85 feet.

Drawdown Increment / Cumulative Drawdown

T gal/day/ft	S = 1.0×10^{-2}	2.3×10^{-3}	1.0×10^{-3}	1.0×10^{-4}
1.0×10^{6}	15.00 / 23.27	30.07 / 51.78	45.79 / 70.73	140.49 / 217.96
4.0×10^{5}	25.06 / 33.73	47.98 / 77.18	88.57 / 127.39	250.41 / 374.22
2.0×10^{5}	36.1 / 56.3	67.5 / 108.2		

1896-1925' actual increment of drawdown over period, 45 feet.

Drawdown Increment

T gal/day/ft	S = 1.0×10^{-2}	2.3×10^{-3}	1.0×10^{-3}	1.0×10^{-4}
1.0×10^{6}	8.27	21.71	24.94	77,47
4.0×10^{5}	8.67	29.20	38.82	123.81
2.0×10^{5}	20.2	40.7	55.3	166.1

Table 8. (continued)

B. Second series of computations: Determining value of S that results in model drawdowns that are closest to actual drawdown for each period when T = 1.0×10^6 gal/day/ft.

1949-1960, actual increment of drawdown, 40 feet.

S	Drawdown
2.3×10^{-3}	40.78

1925-1949, actual increment of drawdown, 40 feet.

S	Drawdown
1.4×10^{-3}	36.66
1.3×10^{-3}	39.14
1.2×10^{-3}	41.00

1896-1925, actual increment of drawdown, 45 feet.

S	Drawdown
4.0×10^{-4}	30.12
3.0×10^{-4}	44.18
2.0×10^{-4}	51.52

In the first series of computations, the values used are:

<u>Coefficient of Transmissibility</u>	
1.0×10^{6}	Specific capacity of University well 2
4.0×10^{5}	Specific capacity of City well 3
2.0×10^{5}	Specific capacity of City well 2
<u>Coefficient of Storage</u>	
1.0×10^{-2}	Maximum value reported for Columbia River Basalt
2.3×10^{-3}	Average value reported for Columbia River Basalt
1.0×10^{-3}	Minimum value reported for Columbia River Basalt
1.0×10^{-4}	Middle of reported "normal" range for artesian aquifers

Results of the first series of computations are shown in Figure 13 and Table 8a. Figure 13 shows that no one set of values of T and S reproduced the actual drawdown over the entire pumping period. Table 8a shows that certain sets of values of T and S did reproduce the actual drawdown during one part of the pumping period, but these same values did not reproduce the actual drawdown in any other part of the pumping period.

For the period 1949-1960, real and model drawdowns match when $T = 1.0 \times 10^{6}$ gal/day/ft and $S = 2.3 \times 10^{-3}$; a match would also be obtained for $T = 4.0 \times 10^{5}$ gal/day/ft if S were somewhat less than 1.0×10^{-2} and for $T = 2.0 \times 10^{5}$ gal/day/ft if S were somewhat greater than 1.0×10^{-2} ; however, 1.0×10^{-2} seems much too large a value of S to be able to occur in an artesian aquifer.

For the period 1925-1949, none of the values used produces an exact match. Matches would be obtained for $T = 1.0 \times 10^{6}$ gal/day/ft when S is slightly greater than 1.0×10^{-3}, for $T = 4.0 \times 10^{5}$ gal/day/ft when S is somewhat great than 2.3×10^{-3}, and for $T = 2.0 \times 10^{5}$ gal/day/ft when S is somewhat smaller than 1.0×10^{-2}.

For the period 1896-1925, none of the values used produced an exact match. Matches would be obtained for $T = 1.0 \times 10^{6}$ gal/day/ft when S is significantly smaller than 1.0×10^{-3}, for $T. = 4.0 \times 10^{5}$ gal/day/ft when S is somewhat smaller than 1.0×10^{-3}, and for $T = 2.0 \times 10^{5}$ gal/day/ft when S is somewhat smaller than 2.3×10^{-3}.

The results of the first series of computations show that more than one set of combinations of T and S can produce model drawdown values that match real aquifer drawdowns, and that combinations of T and S that will produce matching values in any one period will not produce matching values in any other period. In general, for any one value of T, the value of S that will result in matching real and model drawdowns increases substantially in each successive period.

In our second series of calculations, we determined more exactly the probable matching values of S when $T = 1.0 \times 10^6$ gal/day/ft. The results, given in Table 8b, show an increase in value of S of about an order of magnitude since 1896.

These changes with time in model aquifer constants indicate that some of the assumptions on which the model aquifer studies are based may not be valid. Possible reasons are:

1. The aquifer constants of the real aquifers have changed with time and with pumpage.

2. The real aquifer has received significant amounts of recharge.

3. Some combination of both of the above. Evaluation of these reasons is deferred to a later section.

Predicted Drawdown, 1965-2000

Because water from the public-supply wells in the upper artesian zone contains very large amounts of iron, the City of Moscow and the University of Idaho have avoided pumping from the upper artesian zone since 1965 except in emergencies or during high-demand periods in the summer. Several irrigation and industrial wells and numerous domestic wells still utilize the zone, but demands are much smaller than they were when the zone supplied most of the water used in Moscow basin. We do not know the future demands of the upper artesian zone; therefore, we did not attempt to compute a year 2000 water level based on anticipated demand.

We did compute the water remaining in storage as of 1965. Using the coefficient of storage from specific capacity of University Well 2 ($T = 1.0 \times 10^6$ gal/day/ft) and the best-fit model value of S for 1949-1960 ($S = 2.3 \times 10^{-3}$), we computed that the upper artesian zone would yield about 262.8 million gallons a year, or 500 gpm, until the year 2000. At that time, depth to water would be about 150 feet, which is about 25 feet above the top of the producing zone in City Wells 2 and 3, and about 15 feet above the critical depth of 165 feet. Total water in usable storage is 35 years times 262.8 million gallons, or 9.2 billion gallons. This figure is considerably lower than the amount reported earlier (Jones and Ross, 1969, p. 82) owing to a mistake in the original calculations.

This figure on the amount of water in storage is based on the assumption that no recharge takes place and that coefficient of storage will not increase again. Discussion of these assumptions is postponed to a later section.

Pumping of the Middle Artesian Zone

Pumpage and Water Level Data, 1964-1968

Since University Well 3 went into service, daily records have been kept of pumpage as shown by a recording meter and of static and pumping water levels

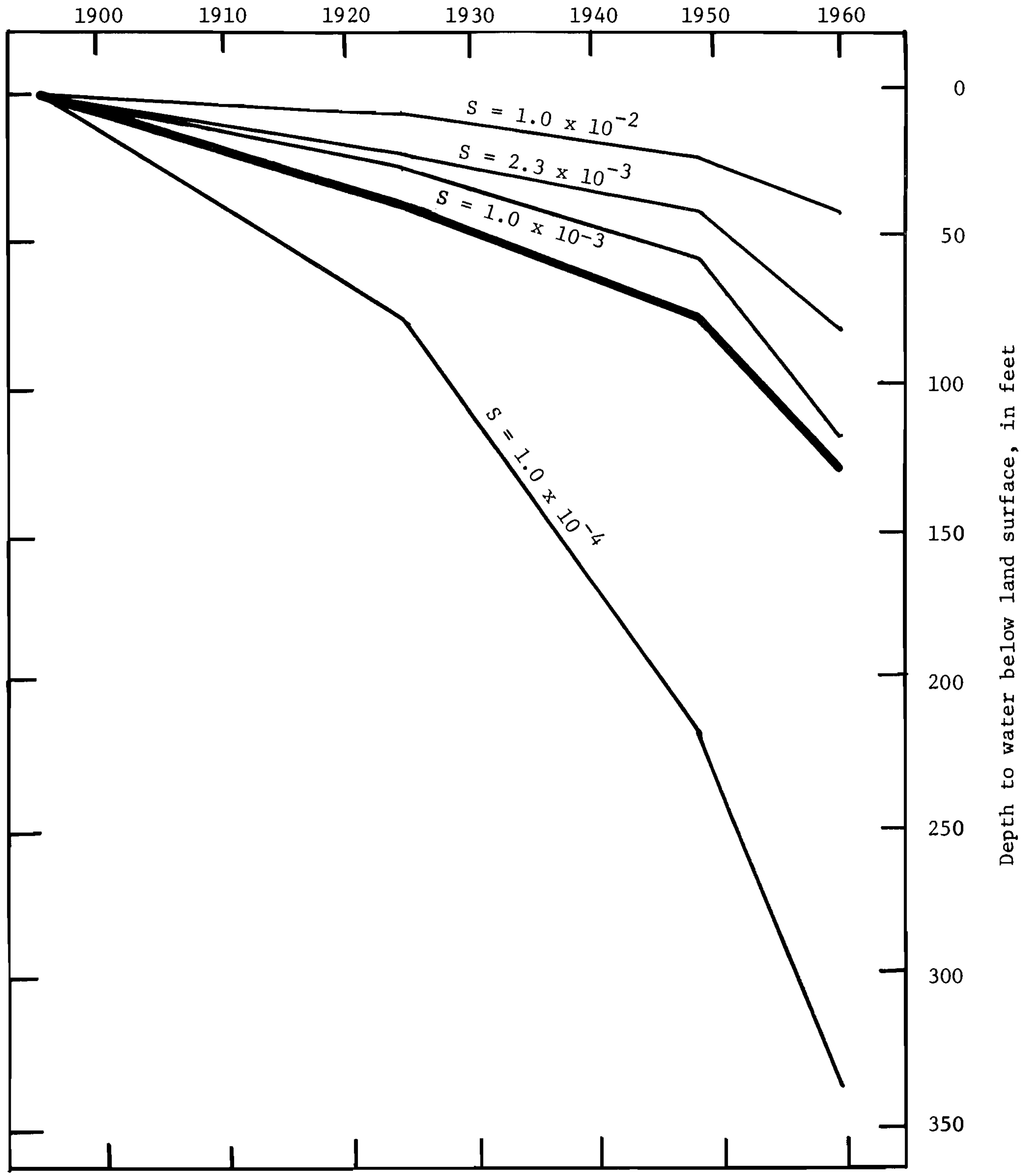


Figure 13. Comparison of actual drawdown in upper artesian zone, Moscow basin, Idaho (heavy line), with drawdowns predicted from model aquifer studies.

a. Model value of Coefficient of Transmissibility (T) = 1.0×10^6 gal/day/ft. Coefficients of Storage (S) as shown.

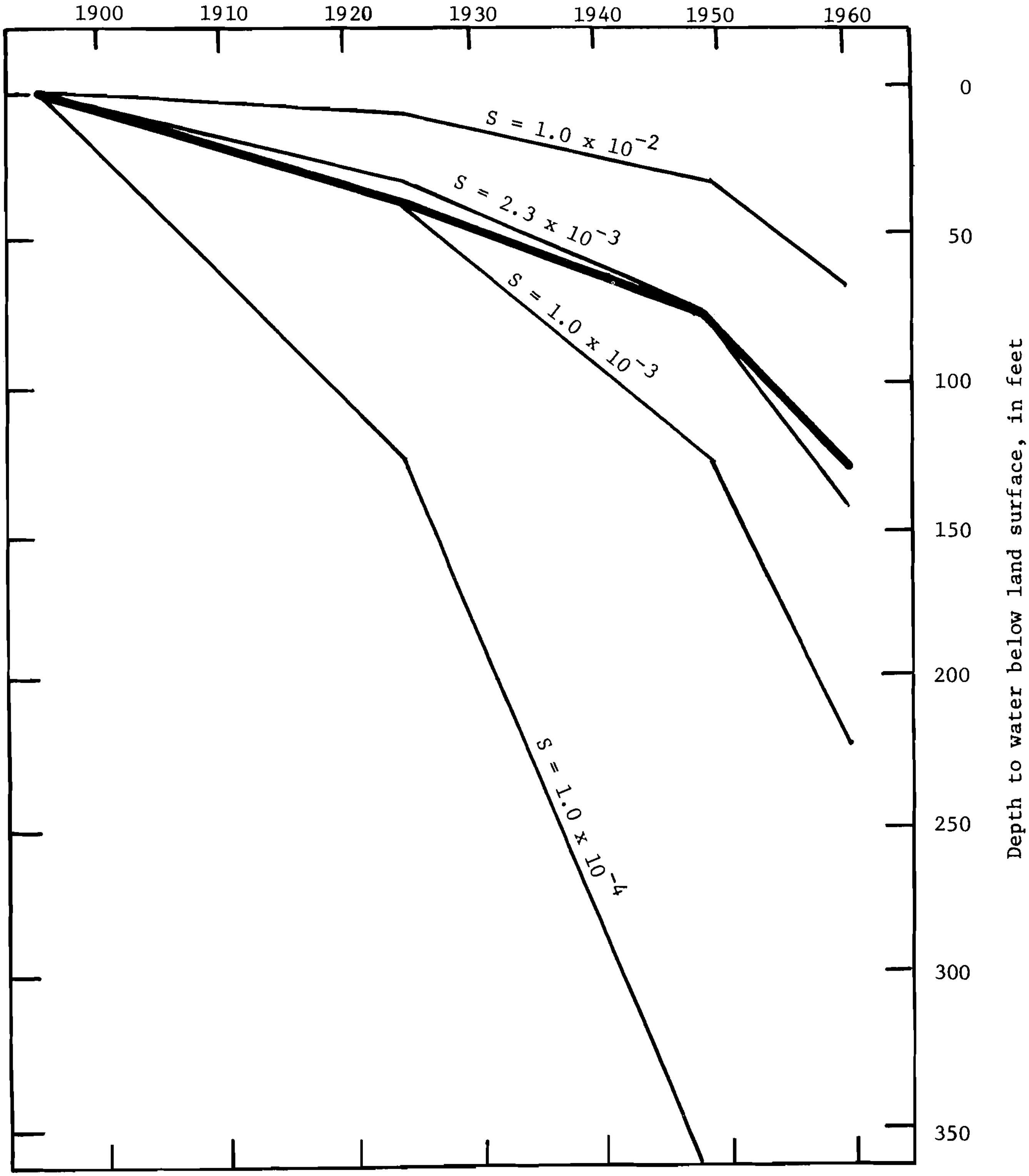


Figure 13 (cont.) Comparison of actual drawdown in upper artesian zone, Moscow basin, Idaho (heavy line), with drawdowns predicted from model aquifer studies.

b. Model value of Coefficient of Transmissibility (T) = 4.0×10^5 gal/day/ft. For S = 1.0×10^{-4}, depth to water is 374 feet in 1949 and 667 feet in 1960 and cannot be shown on the drawing at the scale used.

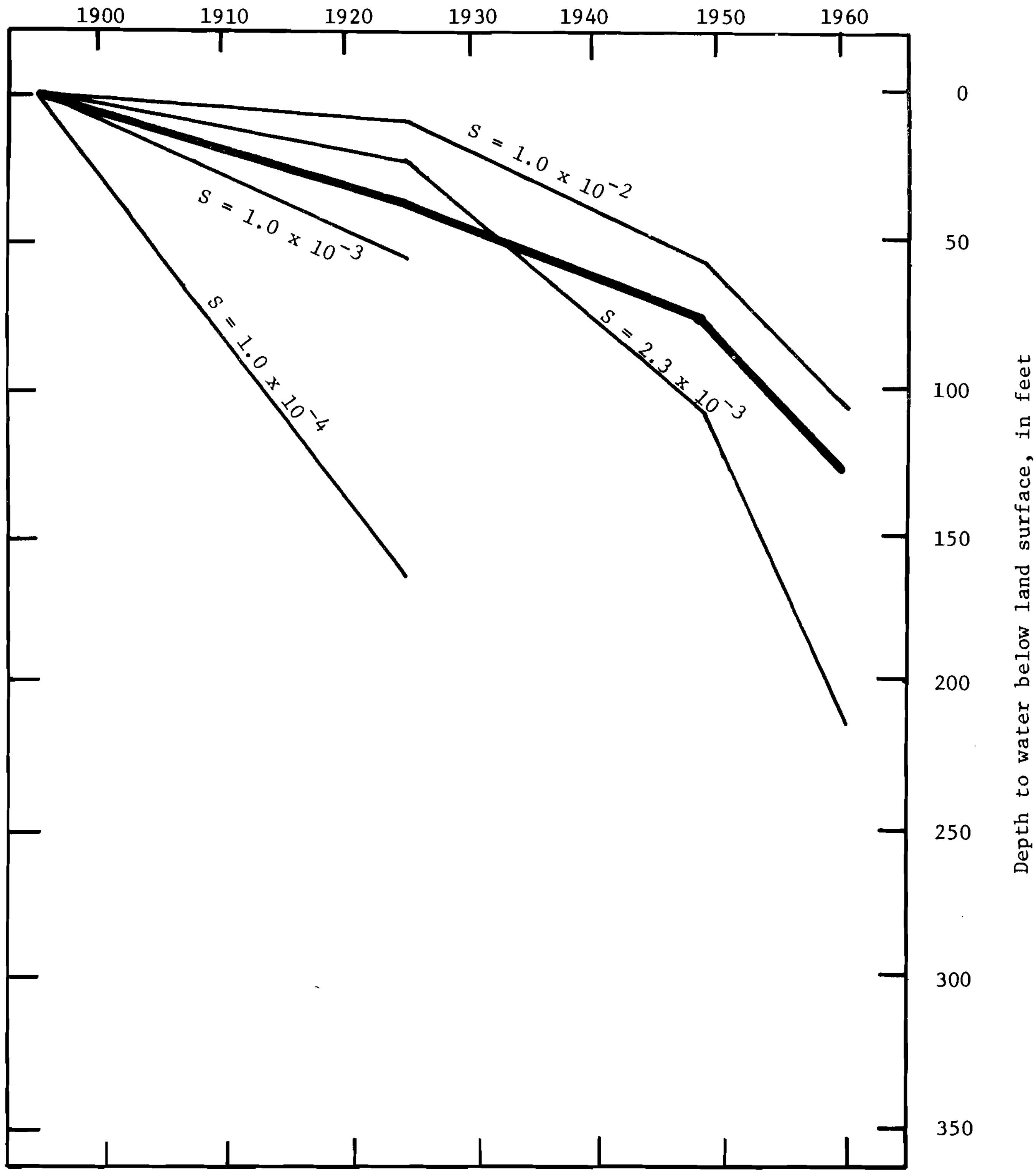


Figure 13 (cont.) Comparison of actual drawdowns in upper artesian zone, Moscow basin, Idaho (heavy line), with drawdowns predicted from model aquifer studies.

c. Model value of Coefficient of Transmissibility (T) = 2.0×10^5 gal/day/ft. Computations for S = 1.0×10^{-3} and 1.0×10^{-4} were not carried past 1925 because they obviously would be very much below actual drawdowns.

taken by airline. University Well 3 also was a project observation well; a recording gage was installed on it for a short time and periodic tape measurements of static water level were taken after the well went into service. The detailed hydrograph is shown on Figure 17; data on water levels and on pumpage are on file with the Idaho Bureau of Mines and Geology. Sokol (1966, p. 13-19) discussed the short-term water level fluctuations in University Well 3.

Model Aquifer Drawdown, 1964-1968

The model of the middle artesian zone (Fig. 9) was tested for the period of record, 1964-1968. The record of water level fluctuations indicates a decline of water levels of 8 feet for the period 1964-1966 and of 10 feet by the end of 1968. Pumpage averaged 164 million gallons a year, or 312 gallons a minute.

Using 1.0×10^6 gal/day/ft, the value of coefficient of transmissibility determined from specific capacity data and coefficient of storage assumed to be the middle value (1.0×10^{-4}) of the "normal" artesian range of Ferris and others (1962, p. 76) the 1966 and 1968 predicted water levels are 288 and 301 feet, much lower than the actual water levels of 258 and 260 feet. Predicted water levels are 259 and 261 feet, almost the same as actual water levels, when the coefficient of storage is assumed to be the reported average value (2.3×10^{-3}) of Columbia River Basalts.

Because of the closeness of agreement of predicted and actual water levels, we believe that the model values of $T = 1.0 \times 10^6$ gal/day/ft and $S = 2.3 \times 10^{-3}$ are at least in the same ratio as the actual values of T and S for the real middle artesian zone. Because the value of T is the same as the value derived from specific capacity for the real quifer, we believe that these figures are very close to the actual values of the real aquifer. The aquifer has not been pumped long enough for any change in coefficient of storage to become evident.

Predicted Drawdown, 1964-2020

The average pumpage for the University of Idaho from 1964 to 2000 probably will be about 300 million gallons per year, or about 570 gallons per minute.

The predicted year 2000 water level is 305 feet in a model that uses the best-fit values of coefficient of transmissibility and coefficient of storage ($T = 1.0 \times 10^6$ gal/day/ft; $S = 2.3 \times 10^{-3}$). In a model that uses the middle of the "normal artesian" range as the value for S (1.0×10^{-4}), the water level is about 470 feet when $T = 1.0 \times 10^6$ gal/day/ft.

To allow for unexpected increased pumpage from the middle artesian zone, we also predicted year 2000 water levels at a pumping rate of 1.3 billion gallons a year, or 2500 gpm. The year 2000 water level is 506 feet in a model that uses the best-fit values of T and S.

Total water in usable storage was computed for the best-fit values of T and S at a pumping rate of 1.576 billion gallons a year, or 3000 gpm. In the year 2020, the water level would be 616 feet and in 2025 648 feet. Critical depth in University Well 3 is 630 feet. Total water in usable storage was computed on the basis of the 2020 water level as 55 years times 1.576 billion gallons per year, or 86.7 billion gallons. The figure on water in storage is based on the assumptions that no recharge takes place and that coefficient of storage will not change. These assumptions will be discussed in a later section.

Pumping of the Lower Artesian Zone

Pumpage and Water Level Data, 1960-1968

Since City Wells 6 and 8 went into service, daily records have been kept of pumpage as shown by non-recording meters and of static and pumping water levels as taken by airline. In addition, a few water-level measurements have been taken by steel tape or electric sounder and a recording water-level gage was operated for a short time on City Well 8. The detailed hydrographs of City Wells 6 and 8 are shown on Figure 17; data on water-level measurements and pumpage are on file with the Idaho Bureau of Mines and Geology. Sokol (1966, p. 23-24) discussed the short-term water level fluctuations of City Well 8.

Model Aquifer Drawdown, 1965-1968

The model of the lower artesian zone (Fig. 9) was tested for the period of record, 1965-1968. The water-level fluctuations in City Well 8 (Fig. 17) indicate a decline of about 12 feet over the period. Data for City Well 8 were used in the model because the real position of City Well 8 is very close to the position of the central well of the model. Pumpage from the lower artesian zone during 1965-1968 averaged 490 million gallons a year, or 930 gpm.

When the value of coefficient of transmissibility obtained from specific capacity (8.0×10^4 gal/day/ft) is used in the model, predicted water levels are very much below actual water levels even when coefficient of storage is very large. When 1.0×10^{-2} (maximum reported value of S for Columbia River basalts) is used, the predicted 1968 water level is 398 feet compared to the actual 1968 water level of 320 feet. When 2.3×10^{-3} (average reported value for S for Columbia River basalts) is used, the predicted 1968 water level is 720 feet. We did not predict the 1968 water level with the value of S that was obtained by using the specific capacity value of T in the ratio $S:T = 2.5 \times 10^{-11}$ that was obtained from the pumping test of March 29-31, 1965 (see p. 31). The value of S was very low, 2.5×10^{-6}, and obviously would result in water levels very far below the actual level.

If the specific capacity coefficient of transmissibility is ignored, and various reasonable values of coefficient of storage for Columbia River basalts are substituted in the ratio $S:T = 2.5 \quad 10^{-11}$, very high values of T are obtained. However, some of them are in the range of values reported for Snake Plains basalt. When the reported minimum value of S for Columbia River basalts (1.0×10^{-3}) is used, T is very high--4.0×10^7 gal/day/ft--but still is compatible with values that have been reported from pumping tests in Snake Plains basalts. Use of these values in the model aquifer yields a predicted 1968 water level of 319 feet, almost exactly the same as the actual water level in the real aquifer. When the reported average value of S for Columbia River basalts (2.3×10^{-3}) is used inthe ratio, T becomes 9.2×10^7 gal/day/ft. Use of these values in the model yields a predicted 1968 water level of 313 feet, only slightly higher than the actual water level. When the reported maximum value of S Columbia River basalt (1.0×10^{-2}) is used in the ratio, T becomes 4.0×10^8 gal/day/ft--a value that is unrealistically high; the predicted 1968 water level is only 1 foot below the initial 1964 water level.

We interpret these results to indicate the model values of $T = 4.0 \times 10^7$ gal/day/ft and $S = 1.0 \times 10^{-2}$ are at least in the same ratio as the actual values of T and S for the real lower artesian zone and are realistic, although very large, values. The specific capacity value of T is too low; any computations based on it would produce results less than the minimum probable performance of the aquifer. The other model values of T are too large to be realisitc, although their ratio to S may be similar to the actual ratio for the real aquifer.

Predicted Drawdown, 1965-2000

The predicted average pumpage for the City of Moscow from 1965 to 2000 is 1130 million gallons a year, or 2150 gallons per minute. We predicted the year 2000 water levels in City Well 8 for the three most probable model values of coefficient of storage and coefficient of transmissibility:

T gal/day/ft	S	Depth to Water, ft
9.2×10^7	2.3×10^{-3}	343
4.0×10^7	1.0×10^{-3}	391
8.0×10^4	1.0×10^{-2}	894

We believe that these are the probable range of values for drawdown and believe that the 391 foot level is the best value because of the close agreement of 1968 predicted and actual water levels when these aquifer constants were used. It should be noted that the best-fit model values are based on pumping test data.

Total water in usable storage was computed with two sets of aquifer constants. With the best-fit model values ($T = 4.0 \times 10^7$ gal/day/ft and $S = 1.0 \times 10^{-3}$), pumping the central well of the model aquifer at the rate of 10.6 billion gallons per year, or 20,000 gallons per minute (almost 10 times the anticipated rate), results in a year 2000 predicted water level of 1085 feet, somewhat below the critical depth of 1020 feet for City Well 8. With the same set of factors, the year 1990 predicted water level is 928 feet, nearly 100 feet above the critical depth. Rather than recalculating until we found the exact year at which the water level would be at the critical depth, we decided to compute water in usable storage on the basis of the 1990 water level. The water in usable storage is 25 years times 10.6 billion gallons, or about 262.8 billion gallons. We consider this to be the optimum estimate of water in usable storage in the lower artesian aquifer; the figure utilizies pumping-test data for coefficient of transmissibility.

We also estimated the amount of water in storage using the specific capacity value of coefficient of transmissibility (8.0×10^4 gal/day/ft and the maximum value of coefficient of storage for Columbia River basalt (1.0×10^{-2}). We believe that these results are considerably below those to be expected from the lower artesian zone, but we also believe that the computations should be run because the coefficient of transmissibility is based on actual data. If the lower artesian zone is pumped at the average expected 1965-2000 demand (1.130 billion gallons a year or 2150 gallons per minute), depth to water would be 906 feet in the year 2000 when $T = 8.0 \times 10^4$ gal/day/ft and $S = 1.0 \times 10^{-2}$. This water level is over 100 feet above the critical depth of 1020 feet for City Well 8. The amount of water in usable storage in this model is 35 years times 1.130 billion gallons, or 39.6 billion gallons--the expected demand by the year 2000. We believe that these figures, which use the least likely data, should be kept in mind although it is unlikely that they are realistic. Both of these figures are based on the assumption that no recharge takes place and that coefficient of storage will not change. These assumptions will be discussed in a later section.

Relations of the Model Aquifers to the Real Aquifers

The model aquifers have the same boundaries on the north, east and south as the real aquifers and were pumped for the same lengths of time and at the same rates as the real aquifers. They differ from the real aquifers in that the model aquifer calculations are based on continuous pumpage for one nearly centrally located well and that the western boundary is undefined. In the real aquifers, pumpage is intermittant and is from several wells located relatively near to each other; furthermore, some sort of a western boundary exists somewhere. Other properties of the model aquifers which may not be the same as those of the real aquifers are the absence of recharge and the values of the aquifer constants.

The model studies of the upper artesian zone show that more than one set of model values of coefficient of transmissibility and coefficient of storage can produce drawdowns that are similar to the drawdowns of the real aquifers. We have assumed that predictions of future drawdowns in an aquifer can be made with any set of values that reproduced real aquifer drawdown. We chose to use values that are compatible with available field evidence. In the upper and middle artesian zines, real aquifer drawdown was duplicated in models that used values of T estimated from specific capacities of wells. In the lower artesian zone, the specific capacity value of T did not reproduce real aquifer drawdown for any reasonable value of S. A value of T derived from the rate of spread of the cone of depression in the lower artesian zone did reproduce real aquifer drawdown.

The model aquifer studies of the upper artesian zone showed that any one set of values of model aquifer constants that reproduced real aquifer drawdown during any one of the three periods of equal rate of decline of piezometric surface would not reproduce real aquifer drawdown in any other of the periods. These differences can be caused by a change of aquifer constants with time, or by recharge taking place, or by both. Absence of a western boundary in the model aquifer does not affect our interpretations.

Changes of Coefficient of Transmissibility with Time

Commonly, a screened or perforated well in unconsolidated sediments may show a small to moderate increase in coefficient of transmissibility during the early life of the well. Pumping of the well increases the permeability of the aquifer in the vicinity of the screen, resulting in an apparent increase in T. Apparent T may also decline in the later life of such a well if the screen openings or perforations become clogged. Some of the wells in the upper artesian zone (City Well 3 and University Well 2) are in unconsolidated sands.

The data on Table 8 suggest an increase in T of about an order of magnitude in 60 years, assuming that S remains constant. We know of no reference to such a large increase in T in the literature and doubt that it could happen. If on the other hand, we assume that T has remained constant at the specific capacity value of 1.0×10^6 gal/day/ft, the best-fit values of coefficient of storage on Table 8 indicate an increase of about an order of magnitude.

Changes of Coefficient of Storage with Time

Theory of ground water storage has been discussed in a series of papers by Jacob (for a summary, see Ferris and others, 1962, p. 74-91). In an artesian aquifer, water is stored by the compression of water caused by the hydrostatic head in the aquifer and by the expansion of the skeletal framework of the aquifer as water is forced into the interstices of the framework--again by the hydrostatic head. The expansion of the aquifer during and after saturation deflects the overlying materials upwards, if both the aquifer and the overlying materials are sufficiently elastic. When the hydrostatic head is reduced--as by pumping a well, for instance--water will expand and water be will forced out of the interstices as the skeletal framework of the aquifer contracts under the load of the overlying materials. In several places in the United States, subsidence of land surface has been related to contraction of underlying artesian aquifers in which the hydrostatic head has been reduced by heavy pumpage (Poland, J.F. and Danes, G.D., 1956; Winslow, A.G. and Doyel, W.W., 1954).

Jacob (1940) noted that a time lag would be present between the time that hydrostatic pressure is reduced and the time that the water actually would be forced out of the interstices, particularly if finer-grained materials such as clay and shale are included in the aquifer system. Jacob did not attempt to evaluate the time lag or the amount of change of S because of the difficulty of the mathematical analysis necessary. Taylor (1948) presents the solution to the one-dimensional flow equation which predicts the rate of consolidation of an aquifer as a function of the rate of water removal. Domenico and Mifflin (1965) have elaborated on this process.

We have seen only a few references in the literature to magnitudes of changes of S with time. Mundorff and others (1964) reported increases in S of as much as two orders of magnitude during the first few days of pumping tests in basalts in the Mud Lake area of the Snake River Plain. They explain the increase as the result of slow draining of moderately porous but slightly permeable materials. Although the changes are similar to those indicated by the models of Moscow basin, they take place over a few days in a system in which water occurs under complex water table conditions. We do not know if the examples from the basalts of the Snake River Plain can be applied to the Columbia River basalts of Moscow basin. However, they do demonstrate that changes in S with extended pumpage are not unique to Moscow basin.

The artesian aquifers of Moscow basin have heterogeneous lithologies, and both the basalts and the sedimentary rocks contain large volumes of low-permeability materials. The silty portions of the sedimentary rocks probably have high porosity and could yield large volumes of water over long periods of time. However, the artesian aquifers are very rigid; Sokol (1966) computed barometric efficiencies of about 90 percent for each of the three artesian zones. High barometric efficiencies indicate rigid aquifers. Also, we do not know of any evidence of subsidence of land surface in Moscow basin such as would be expected in an area where an elastic aquifer has been heavily pumped.

Predictions of future drawdowns are based on the 1965 best-fit values of model coefficient of storage, 2.3×10^{-3} for the upper and middle artesian zones and 1.0×10^{-2} for the lower artesian zone. If conditions of ground water storage are the same in all three zones--which seems a reasonable assumption in view of the similarities of the lithologies of the rocks of the three zones--it might be expected that the initial values of S should be the same for each zone. The best-fit initial value of S for the upper artesian zone is 3.0×10^{-4} when the pumping rate is 290 gpm. The much higher value for the lower artesian zone, 1.0×10^{-2}, might have some relation to the much higher initial pumping rate of 930 gpm. However, relations of S to pumping rate are unclear in that the initial pumping rate of the middle artesian zone was only 312 gpm, but the initial value of S is 2.3×10^{-3}. The long-term development of the upper artesian zone might have influenced the conditions of ground-water storage in the deeper aquifers, but this process also is obscure.

In view of the apparent change of coefficient of storage with time in the upper artesian zone, use of 1965 model data to make prediction of future drawdowns has some hazards. The model S increased with time to values considerably higher than those considered "normal" for artesian aquifers; how much higher the values could go is unknown. Further increase might be possible; on the other hand, decrease is possible if water is being derived from slow drainage of low-permeability materials which are becoming depleted.

Effect of Recharge on Real Aquifer Drawdown

In the theoretical development of the nonequilibrium equation, Theis (1935) assumed that all water is derived from storage, that no recharge takes place, and that the aquifer is infinite. Under these conditions, the cone of depression would never cease to expand outwards, although the rate of expansion would eventually become negligibly small. Water levels in the pumping well would never cease to decline, although the rate of decline eventually would also become negligibly small.

In most actual groundwater systems, the aquifer is finite, has barrier boundaries and/or recharge boundaries, and recharge is received from vertical and/or lateral sources. Under these conditions, the cone of influence expands outwards until sufficient vertical and/or lateral recharge is intercepted to balance the discharge of the well--provided that adequate recharge is available. If the pumping rate is held constant, the water level in the well eventually will become constant when equilibrium is achieved between discharge and intercepted recharge.

Special forms of the nonequilibrium equation have been developed to allow for vertical recharge (see Ferris and others, 1962, p. 118-122); lateral recharge generally is handled by the method of image wells. Hantush (1964) describes a method for handling combined vertical and lateral recharge under water-table conditions.

When the unmodified (no recharge) equilibrium equation is used to study a system that actually does receive recharge, the effect is an apparent increase in coefficient of storage each time that pumping rate is increased--provided that the recharge continues to be large enough to balance the increased demand. The amount of increase depends upon the ratio of the amount of water derived from storage to the amount of water derived from recharge. Studies of the equation of hydrologic equilibrium for Moscow basin (see Ross, 1965, p. 33) indicate that the amount of water available to become ground-water recharge has been greatly in excess of pumpage through the 1960's and will contine to be in excess of pumpage until about the year 2000. The apparent progressive increases in S shown by the model aquifer studies indicate progressively greater amounts of the water available for recharge have been utilized as demand has increased. Because the estimated amount of water available for recharge is still greater than the demand, further apparent increases in S should be possible until such time as demand becomes as great as the actual maximum possible recharge about the year 2000.

The results of the model aquifer study are a direct contradiction of the hypothesis of Crosby and Chatters (1965) that recharge to Moscow basin is very small. We would like to point out that the model aquifer study is an application of principles of well-field hydraulics that are widely used in the study of ground water problems. We believe that the isotope-age data of Crosby and Chatters are incompatible with the hydraulics of Moscow basin; the Moscow basin aquifers do receive a great deal of recharge.

Effect of Western Boundary

The most obvious difference between the three-boundary model aquifer and the real aquifer is the absence of a western boundary in the model aquifer. Certainly some western boundary exists in the real aquifer, but its position and nature are not known.

Even though the no-recharge, three-boundary model is not an exact representation of the real aquifer, the results of the study of the three-boundary model can be applied to the real aquifer. In fact, the results of the study of the no-recharge, three-boundary model make it unnecessary to design and study a no-recharge, four-boundary model that might be a more exact representation of the real aquifer.

The four-boundary, no-recharge model would be much smaller than our three-boundary aquifer and would have a much smaller amount of water in storage. Furthermore, if the western boundary is a barrier boundary--as seems likely--the boundary effect would be an increase in drawdowns in a four-boundary model compared to a three-boundary model. In a four-boundary aquifer in which all boundaries are barrier boundaries, and all water is derived from storage because there is no recharge, drawdowns can only be larger than the drawdowns in the three-boundary aquifer for the same values of T, S, and pumpage.

However, drawdowns in the no-recharge, three-boundary model are already greater than the actual drawdowns in the real aquifer when values of T and S are used that apply to artesian aquifers of the type that are present in Moscow basin. In our studies, the three values of T used are all based on specific capacities of real wells and therefore are real values based on field evidence. When these values are used in our three-boundary model with values of S that are normal for artesian aquifers ($S = 1.0 \times 10^{-3}$ to 1.0×10^{-5}, Ferris and others, 1962), 1960 depths to water are from one hundred to many hundreds of feet greater than the actual depths to water in the real aquifer (Fig. 13). The only exception is for the combination $T = 1.0 \times 10^6$ gal/day/ft and $S = 1.0 \times 10^{-3}$, both of which are very large values.

If the 1960 depths to water in the no-recharge, three-boundary model are already generally much greater than the actual 1960 depths to water in the real aquifer, the 1960 depths to water in a no-recharge, four-boundary model can only be even greater--perhaps enormously greater. The three-boundary model has already demonstrated that the real aquifer did not drawdown as much as it should have compared to an aquifer that receives no recharge and therefore demonstrates that the real aquifer did receive recharge. Thus, there is no need to study a no-recharge, four-boundary aquifer.

Suggested Methods for Further Study of Recharge of Moscow Basin

Considered separately, either the water released from ground-water storage or the water available from natural recharge seems to be sufficient by itself to supply the water pumped from the artesian aquifers of Moscow basin. Most likely, the water pumped comes from both of these sources, but not in equal amounts. Relative amounts can not be determined from the data and procedures available to us; other methods and more data are needed.

One approach to the problem could be through a study of the probable range of values of coefficient of storage in artesian aquifers. In order to duplicate real aquifer drawdown, we had to use values of S as high as 1.0×10^{-2}. If it can be shown that such high values are not possible in an artesian aquifer, even after a long period of pumpage, then much of the water pumped from Moscow basin must come from recharge. To the best of our knowledge, the necessary theoretical studies of long-term coefficient of storage in artesian aquifers have not been made.

Another method would be to design model aquifers for Moscow basin that include recharge. The proportion of recharge to water derived from storage could be increased in successive analyses until model aquifer drawdowns match real aquifer drawdowns.

A field study approach to the problem would be to design pumping tests that would produce useful results in the aquifers of Moscow basin such that reliable values of T and S might be determined. With these data and additional data on other hydrologic factors, a more exact hydrologic budget could be determined and the amount of recharge measured utilizing the real system rather a model system.

Causes of Decline of Piezometric Surface of Upper Artesian Zone

The long-continued decline of water levels in the upper artesian zone has caused concern that pumpage in Moscow basin has been in excess of recharge. Although over-pumping cannot be eliminated as one working hypothesis to explain the declining water levels, an alternate hypothesis can be erected that includes adequate recharge to the basin.

Consideration of the nonequilibrium equation in a finite aquifer has shown that the cone of influence expands until pumpage is balanced by intercepted recharge where adequate recharge is available or until the cone intersects barrier boundaries in all directions. If the pumping rate is held constant, the rate of decline of the water level in the well will decrease as intercepted recharge increases and will become zero when sufficient recharge is intercepted to balance the pumpage. If available recharge is insufficient to balance the pumpage, the rate of decline of water level in the well will increase with time even with the pumping rate held constant. The rate of increase will accelerate as barrier boundaries come into effect.

In Moscow basin, the rate of pumping has never been constant. The data on pumpage from the upper artesian zone (Tables 6 and 7) show an overall increase over the period of 1896-1960, but also show differences of as much as 37 percent between the pumpage of 2 successive years. The pumpages used in the three periods of model studies are the averages over the periods; very likely, the rates were lower in the earlier part of each period and higher in the later part.

Because pumpage has increased at an irregular rate, the cone of influence in the upper artesian zone would be expected to expand and deepen at an irregular rate. The effect of the aquifer boundaries probably increased the rate of decline of the cone. It should be remembered, however, that the positions of the boundaries are known on only three sides of the basin; to the west, some sort of boundary must be present, but its position and nature are unknown. It also should be remembered that the known boundaries were assumed to be barrier boundaries for the purposes of the model studies, whereas the results of the model studies indicate that some of the boundaries may actually be recharge boundaries. The decline of water levels could be the result of the generally increasing rate of pumpage plus complicating effects of the aquifer boundaries.

Perspective on the decline of the water levels near the pumped wells in the upper artesian zone may be obtained by comparing the changes between each of the pumping periods in the amount of water pumped in millions of gallons per year per foot of drawdown. This factor is termed "annual specific capacity".

Period	Average Pumping Rate: Millions Gallons/Year	Decline in Water Levels		"Annual Specific Capacity Millions Gallons/Year	
		Increment	Cumulative	Increment	Cumulative
1895-1925	152	45	--	3.4	--
1925-1949	310	40	85	7.8	3.6
1949-1960	552	40	125	13.8	4.4

These data show that the "annual specific capacity" of the upper artesian zone increased substantially over the period 1896-1960, whereas it should have decreased if the basin were surrounded by impermeable boundaries and the recharge were small. Therefore, these data suggest that the upper artesian zone received considerable recharge over the period studied.

We believe that the hypothesis that assumes that the basin has received considerable recharge is more compatible with the "annual specific capacity" data than the hypothesis that assumes that pumpage has been considerably in excess of recharge. The increasing depths to water is the result of increasing pumping rates and image well effects and does not indicate pumpage in excess of recharge.

Artificially Recharging the Upper Artesian Zone

A proposal for artificial recharge of the upper artesian zone by utilization of seasonal runoff from intermittent streams was presented to the Sixth Idaho Engineering Geology and Soils Engineering Symposium (Jones, Ross, and Williams, 1968). The mathematical model aquifer studies for the recharge system were done before the model studies of pumping of Moscow basin. Knowledge of possible values of aquifer constants gained through the pumping model studies suggests that the annual buildup of the piezometric surface from recharge may be less than was predicted (Jones, Ross, and Williams, 1968, p. 274).

Proposed System

The details of the proposed system and the supporting data are in Jones, Ross and Williams (1968). A brief outline will be presented here to provide background for the following discussion of the mathematical model study.

Several of the intermittent streams that drain the Palouse Range on the north border of Moscow basin have rather large runoffs during the spring months when the snowpack melts, but are essentially dry during the peak demand summer months. The water is not being used in Moscow basin at this time. The outline of the proposed system to utilize this water by artificial recharge is:

1. Divert the spring runoff of the intermittent streams (March through May).

2. Treat the diverted water to bring it up to potable standards.

3. Use the treated water to recharge the upper artesian zone.

4. Withdraw the water from temporary underground storage during the peak demand season (June through September).

The artificial recharge system has advantages over a surface storage system in that capital investment is lower because no large reservoirs must be built because minimal evaporation losses can occur, and because years of abnormally low stream-flow have little effect on the water available during the peak demand season.

An artificial recharge system that diverts water at Robinson Lake on South Fork Palouse River (Fig. 3) and which operates when flow exceeds 1000 gpm, and which has a maximum recharge capacity of 4000 gallons per minute, would operate 90 days a year in 6 years out of 7 and recharge at least 375 million gallongs annually. In 3 or 4 years out of 7, the system would operate 120 days and recharge at least 460 million gallons annually. Regulation of stream flow by small reservoirs might make an additional 80-100 million gallons available annually. During a normal year, South Fork Palouse River above Robinson Lake should yield about 400 million gallons on a run-of-river basis and 480 million gallons if stream flow is regulated. This is about one-third of the anticipated 1965-2000 average annual demand of 1,130 billion gallons.

Gaging records (Jones, Ross, and Williams, 1968, p. 268-269) indicate that in about 1 year in 7, discharge of South Fork Palouse River above Robinson Lake would be in excess of 1000 gallons per minute on only 50 days and would seldom exceed 2000 gpm; the water available for artifical recharge would be only about 70 million gallons. During such dry years, Moscow basin would rely on water naturally in ground-water storage plus any carryover of artificially-recharged water. The deficit would be made up, in part, during other years in which runoff is greater. One or more dry years would have little long-term effect on the artificial recharge system. In contrast, one or more dry years would be disastrous to a surface-storage system in the Palouse Range drainage. Available reservoir sites are small, have low storage volume, and evaporation losses would be large in proportion to storage. The surface water reservoir sites available are not large enough to retain wet-year excess runoff for use during dry years.

Additional water could be obtained from Little Bear Creek, the next drainage basin east of South Fork Palouse River. Assuming that the flow is at least as large as that of South Fork Palouse River, an additional 400 million gallons may be available annually on a run-of-river basis and an additional 480 million gallons if streamflow is regulated. During a normal year, the combined flow of South Fork Palouse River and Little Bear Creek would provide the artificial recharge system with about two-thirds of the anticipated 1965-2000 demand.

Eventually, the effluent of the Moscow Waste-Water Treatment Plant could be used as an additional source of water for artificial recharge. Facilities for tertiary treatment of the effluent would need to be constructed. The Waste-Water Treatment Plant is one source of water in Moscow basin which will increase in volume as population--and demand--increases. Annual discharge, now is about 300 million gallons, should increase to about 1 billion gallons by the year 2000.

At the time that we made the study of the artificial recharge system, we had no data on amounts of water in ground-water storage in Moscow basin. We conservatively assumed that the three artesian zones would yield no more than 500 million gallons annually on a long-term basis (Jones, Ross, and Williams, 1968, p. 264). Our studies of the pumping of the mathematical model aquifers suggest that a great deal more water is available from natural sources in Moscow basin. The artificial recharge system may not be needed for many years, but it should be kept in mind as one way that additional water can be obtained for Moscow basin at a relatively low cost.

Predicted Buildup of the Piezometric Surface under Artificial Recharge

Recharge through wells seems to be the most practical system for Moscow basin. Initially, some of the existing wells could be used for recharge during the spring runoff season, then the same wells would be used as supply wells during the peak demand season. Later on, additional wells may need to be drilled.

As a first approximation, a well should accept water under artificial recharge at about the same rate that the well will yield water when pumped. City Well 3 yields 1350 gpm with about 10 feet of drawdown; presumably, it would build up a cone of impression about 10 feet high in the well when recharged at the rate of 1,350 gpm. The apex of the cone would gradually rise as water goes into storage and as barrier boundaries take effect. Recharge would become impossible if the cone of impression should rise to the surface.

We have reported earlier (Jones, Ross, and Williams, 1968) on a mathematical model aquifer study of artificial recharge of the upper artesian zone. This study was done before our pumping model studies; therefore, we had little knowledge of the values of coefficient of transmissibility and coefficient of storage for the upper artesian zone that became apparent during the pumping model studies. The artificial recharge computations were based on T derived from specific capacities of wells and S assumed as the middle of the "normal" range for artesian aquifers.

Our pumping model studies show that the value of S may be much greater than the assumed value used in the artificial recharge studies. Therefore, we have recalculated probable buildups with the larger values of S; the results are reported on Table 9.

We are not certain as to whether the largest values of S that apply to the pumping models would also apply to the artificial recharge models. The pumping models values of S increased with time; part of the increase could have been the result of compression of the aquifer that forced water out of lenses of clay and silt within the aquifer system. This process may not be fully reversibile, especially during the short time interval of 100 days during which artificial recharge would take place. However, we do believe that the two sets of values of S represent the extremes of the possible range of buildups that are compatible with the available data on the properties of the upper artesian zone.

If we assume that the pumping model data do apply to artificial recharge, we can select any of the combinations of values of T and S that reproduced real aquifer drawdowns in any one of the periods of pumping. In order to evaluate the effect of the minimum and maximum probable values of S on the buildup, one set of values should be used from the 1949-1960 period in which the maximum model values of S apply and one set of values should be used from the 1896-1925 period in which the minimum model values of S apply. For both periods, we used $T = 1.0 \times 10^6$ gal/day/ft; the values of S used are 2.3×10^{-3} for 1949-1960 and 1.0×10^{-3} for 1895-1925. The value of T is estimated from the specific capacity of University Well 2; although the values of T should be lower for City Wells 1, 2, and 3, the larger value of T was used in the pumping model and did reproduce the actual drawdown that took place in the real aquifer in the vicinity of City Wells 1, 2, and 3. Therefore, we feel justified in using the larger value to predict the buildup in the vicinity of these three wells at the City Pumping Plant. The results of the computations are shown on Table 9 and indicate that the maximum buildup would be about 32 feet at the end of recharging 2000 gpm. for 100 days through a single well at the City Pumping Plant.

If we assume that the pumping model data do not apply to artificial recharge, we can compute probable buildup from specific capacity data values of T and an assumed value of S--as we did in our original study. The amount of buildup then depends on the value selected for T. Results of computations for several values of T are on Table 9. If specific capacity data for University Well 2 are used to determine T, the maximum buildup would be 45 feet at the end of 100 days of recharging at 2000 gpm. If the specific capacity data of City Wells 2 or 3 are used to determine T, buildups would be greater--45 to 75 feet at the end of 100 days of recharging 1000 gpm. and 93 to 152 feet at the end of 100 days recharging at 2000 gpm. Because 1965 depths to water in the vicinity of

City Wells 2 and 3 was about 100 feet, recharging at 2000 gpm. through either of these wells would not be feasible because the cone of impression would reach the surface before the end of the recharging season.

The results of these calculations indicate that it is feasible to recharge artificially at the rate of 1000 gpm. for 100 days through City Well 2 or 3, regardless of which set of conditions apply. If the conditions indicated by the pumping model aquifer studies do apply--and it should be recalled that these data gave results that duplicate the actual performance of the real aquifer whereas the data utilizing the "normal" value of S did not--then it is feasible to recharge at the rate of 2000 gpm. for 100 days.

Table 9. Probable Buildup of Piezometric Surface of Upper Artesian Zone Caused by Artificial Recharge through a Single Well at the City of Moscow Pumping Plant. Buildups are Calculated for Several Values of Aquifer Constants for 100 Days of Recharge at 1000 and 2000 gpm.

	Buildups in feet	
	1000 gpm	2000 gpm
A. Buildup using aquifer constants that are conformable with the results of the pumping model studies.		
Highest probable values of S, based on 1949-1960 data: $T = 1.0 \times 10^6$ gal/day/ft; $S = 2.3 \times 10^{-3}$.	6	13
Lowest probable values of S, based on 1896-1925 data: $T = 1.0 \times 10^6$ gal/day/ft; $S = 1.0 \times 10^{-3}$.	16	32
B. Buildup using aquifer constants that are not conformable with results of the pumping model studies.		
Value of S assumed to be middle of "normal" range for artesian aquifers = 1.0×10^{-4}; T estimated from specific capacities of wells.		
Univ. Well 2, $T = 1.0 \times 10^6$ gal/day/ft	24	45
City Well 3, $T = 4.0 \times 10^5$ gal/day/ft	45	93
City Well 2, $T = 1.5 \times 10^5$ gal/day/ft	75	152

INTERPRETATIONS OF LONG-TERM WATER-LEVEL FLUCTUATIONS

Water levels in wells in Moscow basin display short-term, seasonal, and long-term fluctuations in response to natural and artificial changes in hydrologic conditions. Natural long-term changes are caused by natural changes in recharge and/or discharge. Natural recharge to Moscow basin is entirely from precipitation within the basin; therefore, recharge fluctuates with seasonal and long-term fluctuations of precipitation. The natural discharge is largely by underflow out of the basin to the west. Under natural conditions, the amount of underflow would nearly be constant.

Artificial changes in ground-water discharge are caused by pumping of wells. The steady decline of water levels in public-supply wells in the upper artesian zone from the 1890's to the early 1960's indicated long-term changes in hydraulic conditions, but not necessarily a depletion of water available for use. In the early 1960's , pumping of the upper artesian zone was greatly decreased and water levels have been rising ever since.

Short-term water level fluctuations in artesian wells in Moscow basin were discussed by Sokol (1966). His conclusions will be summarized here. The water levels in the artesian wells respond to changes in ground-water storage, to pumping in nearby wells, and to changes in atmospheric pressure. Some of the artesian wells also exhibit short-term, non-periodic fluctuations in response to earthquake shock waves or to wind blowing across the top of open casing. Atmospheric pressure fluctuations at Moscow (Sokol, 1966, Fig. 3) are as much as 0.5 feet of water in one day and as much as 1.2 feet of water over a three-day period. Barometric efficiencies of the artesian aquifers commonly are more than 90 percent; therefore, atmospheric pressure fluctuations can produce as much as 0.4 foot of water-level fluctuation in an artesian well in a single day and as much as 1.0 feet over a period of several days. These changes are large enough to mask changes caused by seasonal recharge.

The basis of our analysis of long-term water level fluctuations are water-level records compiled from various sources. The tabulated data are on file with the Idaho Bureau of Mines and Geology. Hydrographs of the records for wells in Moscow basin are shown in Figures 14, 15, 16, and 17 (Figures 15 and 17 are in the pocket at the end of the report). Sources of data are:

1. Periodic tape measurements and recording water-level gage graphs from 9 project observation wells and 2 U.S. Geological Survey observation wells for the period 1963-1966.

2. Records of periodic tape measurements taken prior to 1963 in the 2 U.S. Geological Survey observation wells, as furnished by the U.S. Geological Survey.

3. Records of periodic and random measurements, nearly all by air-lines, for 7 public-supply wells, as furnished by the City of Moscow and the University of Idaho.

Where recording water-level gage records were used, daily highest water levels are the basis for reporting data from all wells. Some of the wells are under the influence of pumping of nearby wells; therefore, daily highest water level would be the closest to the undisturbed water level of the well.

Fluctuations in Observation Wells in the Surficial Aquifers

Water level measurements were taken periodically by steel tape in 5 wells in the surficial aquifers; water-level recording gages were operated on two of the wells for about 1 year and 2 years. All are water table wells; 4 are are in surficial sediments and 1 is in granitic rocks. The records of the water levels in the surficial aquifers suggest that the water table in Moscow basin has remained at about the same levels since at least the early 1930's. Williams and Allman (1969) have worked out the relations between water level fluctuations and recharge in a portion of the surficial aquifers of Moscow basin.

The 4 observation wells in the surficial sediments are: 39N-5W-8da1, 9bc1, 10ac1, and 40N-5W-31ca2. Water level fluctuations of 10ac1 and 31ca2 (Fig. 14 and 15) reflect only natural changes in water levels. Water levels in 9bc1 are influenced by its own pumping during the summer months and those in 8da1 are affected by lawn watering during the summer.

Well 10ac1 is the key observation well for the surficial aquifers. Its position in the surficial materials overlying crystalline rocks in the valley of the South Fork Palouse River means that the water levels are influenced by the underflow moving from the Palouse Range towards the recharge area of the artesian aquifers at the basalt-crystalline rock contact about 1/2 mile south of 10ac1. In addition to reflecting the annual underflow changes, the well might also show any long-term effects that heavy pumping of the artesian aquifers might have on the overlying water-table bodies. If pumpage of the artesian aquifers is too great to be balanced by the underflow, the water-table might be expected to show a long-term decline as water moves down from the water table to replace water removed by pumpage from the artesian aquifers. Well 10ac1 also has the longest record of any water-table well in Moscow basin. It is a U.S. Geological Survey observation well that was measured in 1934-1940, 1947-1960, and 1964-1966. The complete hydrograph is shown on Figure 15 and the 1964-1966 hydrograph is shown on Figure 14. Precipitation at Moscow is shown on both figures.

The hydrograph of well 10ac1 shows annual fluctuations through a range of about 7 feet which probably reflect annual changes in the amount of underflow passing well 10ac1 in addition to local recharge and discharge of the water table. Long-term changes in the water levels probably reflect long-term changes in underflow and in local recharge-discharge, both of which are the results of changes in precipitation. The generally lower water levels of 1953 through 1955 are in response to lower than normal precipitation during these years, whereas the generally higher water levels of 1956 through 1959 reflect higher precipitation. The lack of a seasonal rise during 1966 may have been due to the unusually low winter and spring precipitation; it may also be due to pumping of wells at a trailer court that went into operation near 10ac1 in 1966. Several other wells shown on Figure 14 had a less-than-normal seasonal rise in the spring of 1966.

The record of 10ac1 suggests that the amount of underflow and local recharge-discharge in the South Fork Palouse River valley has been essentially stable for many years. Comparison of the hydrograph of 10ac1 with the hydrograph of

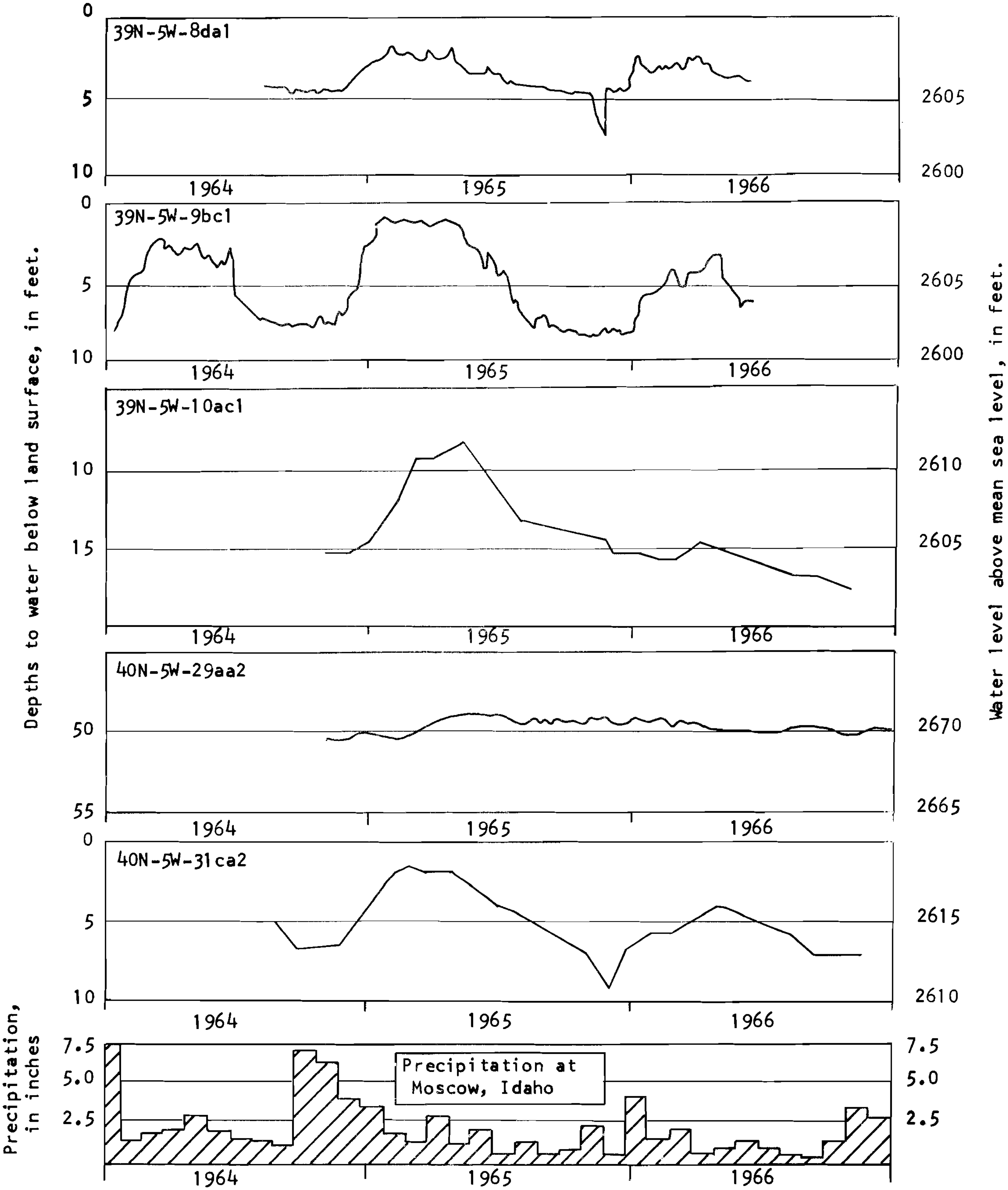


Figure 14. Water-level fluctuations in observation wells in the surficial aquifers, 1964-1966, Moscow basin, Idaho. Precipitation from U.S. Weather Bureau, 1964, and College of Mines Weather Station, 1965-1966.

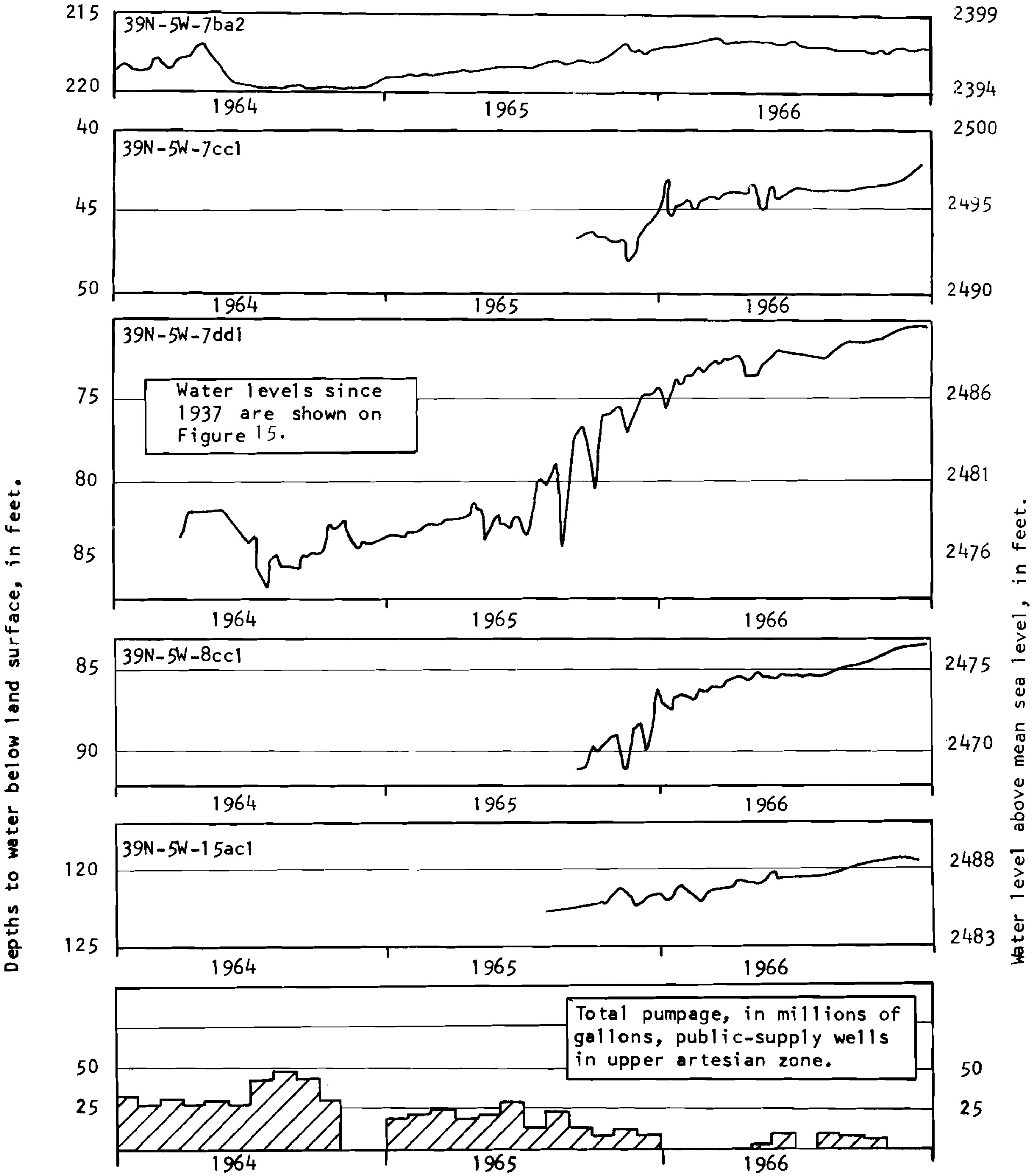


Figure 16. Water-level fluctuations in observation wells in the upper artesian zone, 1964-1966, Moscow basin, Idaho.

7ddl, both shown on Figure 15, shows that 10acl experienced no progressive decline during the period 1934-1960 when 7ddl, a well in the upper artesian zone near the middle of Moscow basin, experienced a decline of nearly 40 feet. The differences in the hydrographs suggest that heavy pumpage in the artesian aquifers has not affected the water table in Moscow basin.

Well 8dal is in a reclaimed swampy area where the position of the water table is controlled by drains and storm sewers. The amplitudes of the fluctuations are small because of the water-table control; some accidental recharge during the summer is caused by lawn watering. Except for the accidental recharge, the pattern of fluctuations is similar to those in 10acl, but of smaller amplitude.

Well 9bcl is in the valley bottom of Paradise Creek where the water table is very close to the surface. A recording gage was operated in the well from January, 1964, to June, 1966. The well is very close to the western edge of the basalt and to the recharge area of the artesian aquifers. Well 9bcl might have shown a decline of water levels if the pumpage of the artesian aquifers was greater than water available for recharge. Although only a few years of record are available for 9bcl, both Figure 14 and a more detailed hydrograph published by Ross (1965, pl. 7) show that during the late-winter to early-spring recharge season, water levels in 9bcl nearly reach the surface of the ground.

Water levels in 9bcl have risen as much as 2 feet in 1 week. During low-water periods, depths to water still are less than 10 feet. Such high water levels in the possible recharge area of the artesian aquifers are not compatible with the concept that pumpage in the artesian aquifers is in excess of the water available for recharge that passes through the recharge area. The pattern of annual fluctuations in well 9bcl is influenced by its own pumpage; fluctuations are less in 1965 than in 1964 and less in 1966 than in 1965 because pumpage was less in each of these years. In spite of summer-time pumpage, the pattern of water-level fluctuations in 9bcl is similar to that of 10acl; the hydrograph of 9bcl shows many minor fluctuations not present on the record of 10acl because the data for 9bcl are taken from a recording gage.

Well 31ca2 is in a valley bottom. It also is fairly close to the eastern edge of the basalt and to the recharge area of the artesian aquifer. Water levels come very close to the surface in the spring and depth to water is less than 10 feet in the fall. Water-level fluctuations seem to be entirely from natural causes; the hydrograph is similar to that of 10acl. The high water table for this well also is not compatible with the concept that pumpage of the artesian aquifers is in excess of the water available for recharge that passes through the recharge area.

Well 29aa2 is a water-table well in the granitic rocks of the surficial aquifers. Unlike the other wells in the surficial aquifers, it is some distance away from the contact with the basalt. Water-level fluctuations are somewhat affected by the pumping of well 29aal, 675 feet away. Daily highest water levels are shown on the hydrograph in order to minimize the effect of pumpage; sustained pumping of 29aal may draw down water levels in 29aa2 by as much as 0.4 foot. Seasonal and annual changes in water levels in 29aa2 are small compared to the other wells, probably because 29aa2 is not in an underflow channel as are some of the other wells in the surficial aquifers.

Fluctuations in Observation Wells in the Upper Artesian Zone

Wells 39N-5W-7cc1, 7dd1, and 8cc1 are artesian wells in the upper basalt and the upper part of the upper interbeds. The three wells are in the central and western parts of the City of Moscow and are in the cone of depression caused by pumping of public-supply wells in the upper artesian zone; principal water-level fluctuations are caused by fluctuations in pumping. Sokol (1966, p. 21) computed a barometric efficiency of 90.5 percent for well 7cd1 (University 2), a public-supply well in the upper basalt and upper interbed. His Figure 3 shows atmospheric pressure fluctuations that are as much as 0.5 feet of water in one day and as much as 1.1 feet of water over a three-day period. In wells with a 90 percent barometric efficiency, these atmospheric pressure changes would cause water-level changes of about 0.4 and 1.0 feet respectively. In the three wells, short-term water-level fluctuations caused by natural changes in ground-water storage are masked by the changes caused by fluctuations in pumping and/or fluctuations in atmospheric pressure.

Well 7dd1 is the key observation well in the upper basalt and upper interbed. It is a U.S. Geological Survey observation well with a record of measurements that begins in 1937. The complete hydrograph is shown on Figure 15 and the 1964-1966 hydrograph on Figure 16. Generalized water level trends are shown in Figure 12. Well 7dd1 is about 1500 feet from the center of pumpage of the City of Moscow (City 1, 2, and 3) and about 3000 feet from the center of pumpage by the University of Idaho (University 1 and 2). These public-supply wells in the upper basalt and upper interbed were essentially the only sources of water prior to 1960 and were heavily pumped until the early 1960's. Beginning in 1960, the source of the public supplies was shifted to deeper wells with City 6 going into service in late 1960; University 3, in late 1963; and City 8, in mid-1965. During this time, pumping of the upper basalt and upper interbed aquifers decreased, and virtually ceased in 1965-1966. Water levels in 7dd1 rose slightly in 1960 through 1964 in response to decreased pumpage, then rose sharply during 1965-1966. This increase in water levels in 7dd1 indicates that the cone of depression in the upper artesian zone centered on Moscow has decreased in size and that ground-water storage in the upper artesian zone has increased. As this report was written (summer of 1969), water levels in 7dd1 were still rising, but at a lesser rate than in 1965-1966. The 1969 water levels were at about 65 feet--the same depths as in the late 1940's and early 1950's but still below those of the 1930's.

The records of wells 7cc1 and 8cc1 are much shorter than that of 7dd1, but show the same rising trend of water levels during the period of record. The depth of 7cc1 is not known, but a son of the former owner (Albert Hagen, oral communication) recalled that clean white sand was penetrated at the bottom of the well. This sand may be from the upper interbed. The well is blocked by a piston pump housing at about 70 feet. However, water-level fluctuations in the well are thought to represent the fluctuations within the aquifer because comparison of the measurements from 7cc1 with recorder graphs from wells showing fluctuations caused by changes in atmospheric pressure indicate that the water levels in 7cc1 also respond to atmospheric pressure changes. Water levels in 8cc1 also seem to respond to atmospheric pressure changes.

The Parker Farm well of the University of Idaho (15bc1) is an artesian well in the upper basalt aquifer that was drilled in 1957. The 1964-1966 water level fluctuations are shown on Figure 16 and are shown again in relation to the 1957 water level in Figure 17. The principal cause of water-level fluctuations is pumping of the well, which tends to obscure changes from other causes. Two nearby wells (15ac1, Elks Golf Course; and 15ca1, Bennett Lumber Products) do not seem to be hydraulically connected with the Parker Farm well. The two wells were heavily pumped during the period of record. The Elks well is about 2100 feet away and is pumped almost continuously during the summer months at a rate of about 60 gpm. In 1962, the static water level in the Elks well was at elevation 2535, or about 60 feet below the surface, which is nearly 60 feet higher than the probable water level in the Parker Farm well (elevation 2488, or about 120 feet below the surface). Chemical analyses of waters from the two wells are very similar (Tab. 10; Fig. 25). The large difference in elevations of the piezometric surfaces suggests that the Parker Farm well and the Elks well are not connected hydraulically. The Bennet well, 1500 feet away, is pumped continuously, all year around. The static water level in the Bennett well is at about elevation 2505, or about 90 feet below the surface, which is only about 17 feet higher than the water level in the Parker Farm well. However, the chemical analyses of water from the two wells are very different (Tab. 10; Fig 25). The small difference in elevations of piezometric surfaces plus the very great difference in chemical analyses suggests that the Parker Farm well and the Bennett well are not connected hydraulically. It follows that, from the differences in elevation of piezometric surface and the differences in chemical analyses, the Elks well and the Bennett well are not connected hydraulically, even though they are only about 700 feet apart.

The hydrographs of the Parker Farm well (Fig. 16 and Fig. 17) show a steady rise of water levels during 1965-1966 to levels that are above the static water level of the pumping test in November, 1957. This pattern is very similar to that of well 7dd1 in downtown Moscow, but the rise is not as great. Moscow basin water levels were at about their lowest in 1957 and rose steadily during 1965-1966 and were considerably above 1957 levels in mid-1969. Such a similar pattern of the hydrographs suggests that the Parker Farm well is hydraulically connected with the public-supply wells in Moscow even though it is about 12,000 feet from the center of City pumpage and about 15,000 feet from the center of University pumpage. However, the chemical analysis of water from the Parker Farm well (Tab. 10; Fig. 25) is very different from the chemical analyses of water from the public-supply wells in Moscow that pump water from the upper basalt and upper interbed aquifers (Tab. 10; Fig. 25). The water-level fluctuations in the Parker Farm well may represent natural increase in ground-water storage during 1965-1966.

City Well 7 is an artesian well perforated in the lower part of the second interbed. Short-term water-level fluctuations were discussed by Sokol (1966, p. 11-12). He shows that fluctuations are caused by changes in atmospheric pressure and changes in storage. He suggests that the aquifer penetrated by City 7 is not directly connected hydraulically with any of the pumped aquifers and that changes in storage are natural.

With the exception of 7ba2 (City 7), all observation wells in the upper artesian zone showed rising water levels during the period of our field studies (1964-1966). These rising levels were in response to decreased pumpage and indicate an increase in ground water in storage. At the time this manuscript was written (summer, 1969), water levels were still rising in the one well still being measured (7dd1).

Fluctuations in Public Supply Wells in the Upper Artesian Zone

University 1, University 2, City 1, City 2, and City 3 are artesian wells in the upper basalt and upper interbed aquifers. For many years, these were the principal public-supply wells in Moscow basin and heavy pumping from them caused a deep cone of depression in the piezometric surfaces of the upper artesian zone. Individual detailed hydrographs of most of these wells since the city and University began periodic water level measurements in the early 1950's are shown on Figure 17. City 1 is omitted from Figure 17 but is shown along with City 2 and University 1 and 7ddl on Figure 12, a generalized hydrograph for the period 1896-1966.

A recording gage was in operation on University 2 during the period June 25 to July 14, 1964. Water-level fluctuations during this period were analyzed by Sokol (1966, p. 21-22). He computed a barometric efficiency of 90.5 percent and concluded that the short-term water-level fluctuations were caused by changes in atmospheric pressure, by seasonal recharge and/or recovery since shutdown of the well, and by pumping of other wells. Prior to June, 1964, the principal cause of water-level fluctuations was pumping of the well. The well has been pumped very little since University 3 went into service.

City 1, City 2, and City 3 are within a hundred feet of each other and seem to obtain water from the same zones. The water-level fluctuations in each of these wells are caused by its own pumping and probably also are caused by pumping of the other two. The water-level measurement record is most complete for City 2; the pattern of the generalized hydrograph on Figure 12 is similar to that of 7ddl: increasing demand in the 1940's and 1950's caused steady decline of water levels until the source of public supplies was shifted to the deeper wells in the early 1960's. The water levels have since risen.

University 1 is about 2200 feet from University 2, but Sokol (1966, p. 21) presents evidence that the two wells are poorly connected hydraulically. The principal cause of water-level fluctuation in University 1 is its own pumping. After University 2 went into service, pumping in University 1 declined and water levels rose after 1958. Only one measurement is available for 1963, and none thereafter.

During the period of our field studies (1964-1966) water levels in most public supply wells showed the same rising trends as were shown by the observation wells. The rise is in response to decreased pumpage and represents an increase of ground water in storage.

Fluctuations in an Observation Well in the Middle Artesian Zone

University 3 is an artesian well perforated in the middle basalt aquifer. Before the well was placed into service in September 1964, records of water-level fluctuations were obtained by a recorder between December 14, 1963 and May 7, 1964, and by tape measurements thereafter. The hydrograph is shown on Figure 17. Sokol (1966, p. 13-19) analyzed the records prior to September, 1964. He computed a barometric efficiency of 91 percent and concluded that the principal short-term water-level fluctuations are caused by fluctuations in atmospheric pressure. After removing the effects of atmospheric pressure changes from the hydrographs, he demonstrated small water-level changes from natural changes in ground-water storage. Other water level changes that he discussed are non-periodic and are caused by earthquakes and by wind blowing across the top of the casing before the well house was built.

After University 3 went into service, we took tape measurements at about weekly intervals and the University of Idaho took daily airline measurements. Both sets of data are reported and are shown together on Figure 17 for purposes of comparison. Since the well went into service, the principal cause of water-level fluctuations has been pumping of the well. Water levels declined about 8 feet by December of 1966, at the end of two years of operation. By the end of 1968, the water levels had declined an additional 2 feet. These declines were in response to the pumping of the well.

Fluctuations in Public-Supply Wells in the Lower Artesian Zone

City 8 is an artesian well in the lower basalt and City 6 is an artesian well in the lower basalt and lower interbeds. No periodic water-level measurements were obtained from City 6 prior to its being placed into service; a recording gage was operated on City 8 from March 15 to April 19, 1965. During this time, a record of the operation of the pump on City 6 was kept by a recording voltmeter installed on the pump controls. Water-level fluctuations for the period were analyzed by Sokol (1966, p. 23-24), who estimated the barometric efficiency of City 8 to be 80 percent and determined that the major cause of water level fluctuations was pumping of City 6, which is 6000 feet away. Comparison of extended-scale recorder graphs of the water levels in City 8 with the recording voltmeter graphs from City 6 (Sokol, 1966, Fig. 15) shows that the direction of trend of water-level in City 8 reverses within 1 to 5 mintes of the time that the pump on City 6 starts or stops. During the period that the recording instruments were in operation, pumping of City 6 caused daily water-level fluctuations through a range of 2 feet in City 8. The quickness with which water levels in City 8 respond to changes in pumping in City 6 and the amount of influence that pumping of City 6 has on the water levels in City 8, show that these two wells have very good hydraulic connection. Pumping of City 8 should have strong influences on water levels in City 6.

The water-level records for City 8 (Fig. 17) are from several sources. The pumping-test static level of December 11, 1964, probably was an airline measurement; it is believed to be inaccurate. The recorder graph records of March 15-April 19 were checked with steel tape measurements each time the charts were changed and are accurate. In the summer of 1968, the Moscow City Engineer discovered that the airline on City 8 was 10 feet shorter than originally reported. The data shown on Figure 17 have been corrected to remove this error.

According to R.O. Day, then Moscow City Engineer, (oral communication, 1967), recovery of City 8 is rather slow and most of the shut-down periods of the pump are too short to permit full recovery. He reported that when the pump is off for a period of more than several hours, water levels will recover to about 310 feet. The measurements furnished by City of Moscow for the period mid-1965 through 1966 range from between 318 and 330 feet and probably are random points on the recovery curve of the well. The water levels probably are also affected by drawdown caused by pumping of City 6. The reported recovery of City 8 to about 310 feet plus the several "normal" measurements in the range of 318, 319, 320, and 321 feet during late 1966 suggest that the static level of City 8 had declined about 10 feet by December, 1966, after about $1\frac{1}{2}$ years of operation.

The principal causes of water-level fluctuations in City 6 are its own pumping, and (since mid-1965) pumping of City 8. Between 1960 and 1966, City 6 was pumped at about 900 gpm. Water levels declined slightly and were about 2 to 5 feet lower at the end of 1966 than they were in 1960. On January 26, 1967, the water level was measured with a steel tape and was 289.20 feet below land surface. At the time of the measurement, the pump had been removed from the well and it had not been pumped for several weeks. In 1967, the pumping rate of City 6 was increased to about 1200 gpm.

Relations of Water-Level Fluctuations to Ground-Water Supply

Water that recharges the artesian aquifers must pass through the surficial aquifers on the way down to the artesian aquifers (Fig. 6). The water moves along the basalt-crystalline contact, or through buried stream channels (Chang-Lu, 1957, p. 76-80) and/or vertically through the top of the basalt sequence. Heavy pumpage in the upper artesian zone between 1896 and 1960 created a substantial decline in the piezometric surface that must have increased the rate of downwards movement of water. If recharge to the water table was less than the amount of water moving to the artesian aquifers, a long-term decline of water levels should have occurred in the water-table aquifers. If water available for recharge was in excess of the amount of water moving to the artesian aquifers, it would move into the water table aquifers in amounts sufficient to keep the water-level fluctuations within the same range as when pumpage of the artesian aquifers did not affect the water table bodies.

All evidence shows that the water table in Moscow basin has not declined since at least the early 1930's and suggests that the water table has remained stable since the 1890's. In this period of time, the water levels of the upper artesian zone declined nearly 120 feet in the vicinity of the pumped wells.

Water-level fluctuations in the artesian aquifers generally show seasonal rises in the spring that Sokol (1966, p. 25) relates to recharge. The long-continued rise in water levels in the upper artesian zone after phasing out of heavy pumpage in the early and middle 1960's is additional evidence of recharge.

The seasonal rises in the artesian aquifers, the long-term stability of the water table all indicate that water available for recharge in Moscow basin still was in excess of pumpage at the time we did our field studies. Water moves down from the water table to replace water pumped from the artesian aquifers. Water that formerly passed out of Moscow basin as surface flow or evapotranspiration ("rejected recharge") moves down into the water table to replace the water that went down to the artesian aquifers. Enough water has moved down from the formerly rejected recharge to keep the water table fluctuations within the same range for the period 1934-1966 and probably for the period 1896-1966.

Additional rejected recharge probably still is available. The estimates of water available for recharge in Moscow basin (see Ross, 1965, p. 30-35) suggest that recharge should exceed pumpage through the year 2000.

If pumpage from the artesian zones should begin to exceed all available recharge, the water table should begin to decline gradually. A long-term program of water-level measurements would provide the data necessary to determine if the water table is remaining stable or beginning to decline.

GEOCHEMISTRY OF GROUND WATER

Natural water never consists of pure H_2O. Dissolved inorganic solids, gasses, and perhaps dissolved organic matter are present in ground waters. In "fresh water", the dissolved solids commonly make up less than 1 percent by weight; in Moscow basin ground waters, the dissolved solids generally total 0.01 to 0.05 percent. The term "chemical composition of water" means the kinds and amounts of materials dissolved in the water.

Even rain water, though commonly regarded as pure, contains some dissolved solids and generally picks up carbon dioxide gas as it passes through the atmosphere. The carbon dioxide undergoes a chemical reaction with the water to make a very dilute solution of carbonic acid. As the slightly acid water percolated down through the soil and rocks to the water table to become ground water, and as the ground water moves through the soil and rocks, various chemical processes of solution, deposition, and reaction take place that cause changes in the chemical composition of the water. The kinds of processes that will take place are influenced by the mineralogy and other geologic features of the rocks, by the volume of water and the rate of circulation, and by physical-chemical variables such as pressure, temperature, pH, and oxidation potential. In some place, changes are caused by actions of man. These artificial changes are classed as contamination or pollution.

Geochemistry of ground water is the study of the changes in chemical composition of water as it circulates through the ground. Deduction of the reasons of the changes in terms of physical-chemical relations is the goal of geochemical study.

During the time that the public water supplies of Moscow basin were obtained from the upper artesian zone, excessive amounts of iron caused many annoying problems. One of the goals of our study was to determine the factors that control the origin and distribution of iron in the ground waters of Moscow basin. Ross (1965, p. 83-88) determined the hydrologic factors during her preliminary studies. Our follow-up studies eliminated many simple chemical and physical-chemical relations as causes of the origin and distribution of the iron but did not determine the actual causes.

Preliminary Study

During her investigations of the geohydrology of Moscow basin, Ross (1965, p. 65-88) accumulated much basic data on the chemical quality of the ground waters. She assembled data from 70 chemical analyses from various sources; most of the analyses reported only a few constituents of the water. She determined field specific conductivity of water from about 200 wells, determined iron content of water from about 150 wells, and measured temperatures of water from about 62 wells.

She presented two isogam maps of distribution of specific conductivity of ground waters in Moscow basin; her plates 9 and 10 show specific conductivity in the "shallow unconfined aquifers" and in the "upper confined and deeper unconfined aquifers" respectively. These maps have been revised; Figure 23 is distribution of specific conductivity in the surficial aquifers and Figure 27 in the upper artesian zone.

Ross also presented two isogam maps of distribution of iron in ground waters in Moscow basin. Her Plate 10 shows distribution of iron in the "shallow unconfined aquifers" and her Plate 11, in the "upper confined and deeper unconfined aquifers". These maps also have been revised; Figure 22 is distribution of iron in the surficial aquifers and Figure 26, in the upper artesian zone.

Ross discussed the hydrologic controls of the origin and distribution of iron and silica. The discussion is given below; few revisions were necessary.

Most of the wells for which chemical analyses of water were obtained were selected on the basis of data in Ross' report. Water was analyzed from wells believed to be typical of each aquifer, from wells with waters high in iron, or from wells with other anomolous characteristics. The analyses are shown on Table 10. Partial quality of water data, such as temperatures, iron, field bicarbonate, specific conductivity, and field pH were obtained on 22 additional wells after Ross completed her preliminary study. These data are on file with the Idaho Bureau of Mines and Geology.

Sources of Chemical Analyses

Chemical analyses of water from 42 wells and springs in Moscow basin are shown in Table 10. Samples for 38 of the analyses were collected by R. W. Jones and were analyzed by the Department of Agricultural Biochemistry and Soils of the University of Idaho. Two of the analyses are by the Idaho Department of Health and two are the averages of a number of analyses from various sources that were reported by Ross (1965, Tab. XVI). Data that supplement the analyses were obtained by S. H. Ross, R.W. Jones, and D.A. Myers.

Thirteen of the samples are for freshly pumped water collected from valves that bypass the storage system fed by the pump. These samples of water probably are very similar to the water circulating in the aquifer. The other samples were collected from taps in the storage system or were dipped from open wells. The water in these samples probably was modified to an unknown extent by the metal and air in the storage system or by contact with the atmosphere. Iron, bicarbonate, and pH are particularly susceptible to such alteration. To minimize changes in samples after collection, the samples were held in a refrigerator from a few hours after collection until analysis.

Procedures for Obtaining Field Analytical Data

In general, the analytical laboratories reported: SiO_2, Al, Ca, Mg, Na, K, HCO_3, SO_4, Cl, NO_3, and pH. Some of the analyses also showed NH_3, PO_4, or Mn. Although the analytical laboratories did report Fe and the Department of Agricultural Biochemistry and Soils reported specific conductivity, their data are not shown on Table 10 but are shown on Table 11. On Table 10, the data on temperature, Fe, HCO_3 (Field), F, specific conductivity, and pH (Field) were obtained by R.W. Jones, S.H. Ross and D.A. Myers. Total dissolved solids and hardness were calculated using methods in Rainwater and Thatcher (1960, p. 176-177; 271-273).

Temperature

Temperature was taken with ordinary thermometers that probably were accurate to 1 degree. Temperature was measured only when water came directly

Table 10. Chemical analyses of ground waters, Moscow basin, Idaho.

Analysis number	Well number	Date sample collected	Temperature, °F 2/	Silica (SiO_2)	Aluminum (Al)	Iron (Fe) 2/	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate (HCO_3) (Field) 2/	Bicarbonate (HCO_3) (Lab)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F) 2/	Nitrate (NO_3)	Dissolved solids (calculated)	Hardness as $CaCO_3$ (calculated)	Specific conductance (micromhos/cm at 25°C) 2/	pH (Field) 2/	pH (Lab)	Collection point	Remarks
	SURFICIAL AQUIFERS																						
	39N-5W																						
1	2bd1	11/23/65	-	25.0	0	0.0	28.8	7.1	16.8	2.0	160	155	0.43	13.9	0.2	1.99	148	101	220	7.3	7.2	tap	NH_3, 0.2 ppm.
2	8ab3	11/24/65	50	24.2	0	0.20	27.8	8.3	14.7	0.8	151	134	0.9	13.9	0.2	12.4	145	104	235	7.0	7.35	dipped	NH_3, 0.2 ppm.
3	9bc1	8/12/65	54	23.8	tr	tr	59.6	15.8	46.2	1.6	-	240	25.0	76.7	-	43.9	386	214	550	6.35	6.8	bypass	
4	9bc2	8/12/65	56	21.7	tr	tr	38.2	14.0	42.5	0.8	-	219	20.5	41.7	-	36.3	302	153	520	6.4	6.95	dipped	
5	9bd1	8/12/65	-	20.4	tr	0.06	18.0	3.6	23.8	0.8	-	151	4.5	6.7	-	0.2	132	61	205	6.6	7.2	tap	
6	9bd2	8/12/65	-	19.2	tr	4.0	19.2	4.2	32.5	1.2	-	158	7.5	6.7	-	2.0	85	65	240	6.8	7.15	tap	
7	9cc1	11/23/65	-	24.2	0	3.6	53.6	16.6	16.8	4.0	2.8	24.4	206	7.1	1.0	0.22	321	202	440	5.35	7.50	tap	
8	10cb1	8/13/65	-	19.2	tr	1-1/2	23.4	4.8	18.7	1.6	-	144	3.5	4.0	-	tr	127	78	215	7.15	7.25	tap	
9	11da1	8/11/65	52	25.8	tr	0.13	25.8	3.6	21.2	0.8	-	111	3.5	9.6	-	10.6	130	79	220	7.2	7.00	dipped	
10	12ab1	8/11/65	52	23.3	tr	tr	24.8	6.1	17.5	1.2	-	140	tr	4.0	-	1.3	124	88	205	7.55	7.00	bypass	
11	16ac1	8/13/65	-	23.3	tr	0.16	58.2	15.5	87.5	1.6	-	165	38.5	84.6	0.0	133	500	209	770	6.7	6.75	tap	NH_3, 0.3 ppm.
12	16ad2	8/13/65	-	21.7	tr	tr	23.8	4.2	18.7	0.8	-	129	10.5	6.7	-	6.9	136	77	200	7.1	7.3	bypass	
13	16bc1	11/24/65	48	16.7	0	0.05	31.6	9	33.6	1.6	193	194	22.1	13.9	0.5	0.38	208	116	330	7.45	7.50	dipped	NH_3, 0.3 ppm.
14	16bc2	11/23/65	52	32.1	0	4.25	23.0	11.3	12.6	2.8	178	159	2.4	13.9	0.5	1.55	146	104	230	7.25	7.30	bypass	NH_3, 0.3 ppm.
15	17cd1	8/13/65	-	16.3	tr	11.1	24.8	6.1	12.5	0.8	-	137	5.0	3.7	-	0.4	131	87	200	7.85	7.35	tap	
16	26bb1	8/13/65	54	20.4	tr	0.26	56.8	11.9	31.3	2.8	-	197	15.5	28.7	-	43.9	288	191	490	7.3	7.15	bypass	
	40N-4W																						
17	30ca1s	8/11/65	-	23.3	-	0.0	9.4	2.4	12.5	1.6	-	43	tr	5.3	-	0.2	52	33	80	6.8	6.60	tap	
	40N-5W																						
18	18db2	8/11/65	-	22.5	tr	1.85	16.2	1.8	15.0	0.8	-	101	1.0	6.7	-	0.00	93	48	140	7.55	7.10	tap	
19	18db3	11/18/65	54	17.5	0	0.20	9.0	2.56	7.35	0.8	46	52.6	0.43	7.1	0.0	0.89	55	33	80	6.75	6.90	dipped	NH_3, 0.30 ppm.
20	20cb1	8/11/65	-	14.2	-	0.50	32.4	8.5	18.7	2.0	-	197	8.5	3.7	-	2.2	174	126	250	7.9	7.70	tap	
21	33bd1	11/18/65	49	17.1	0	10.4	113.2	28.4	33.6	1.6	97	85.8	33.7	139	0.0	194.92	587	400	850	7.1	6.7	tap	
22	Average		-	-	tr	1.63	24.7	6.0	19.3	1.4	-	139	5.4	9.3	-	-	127	87	-	-	-	-	Ave. of 16 anal. of uncontaminated waters

1/ Chemical analyses by Department of Agricultural Biochemistry and Soils, University of Idaho, except 25 and 33 are from Idaho State Department of Health and 40 and 41 are averages of analyses from various sources reported by Ross (1965, Table XVI).

2/ Determined by S.H. Ross, R.W. Jones, or D.A. Myers by analytical techniques or by portable apparatus.

Table 10. Continued.

Analysis number	Well number	Date sample collected	Temperature, °F 2/	Silica (SiO_2)	Aluminum (Al)	Iron (Fe) 2/	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate (HCO_3) (Field) 2/	Bicarbonate (HCO_3) (Lab)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F) 2/	Nitrate (NO_3)	Dissolved solids (calculated)	Hardness as $CaCO_3$ (calculated)	Specific conductance 2/ (micromhos/cm at 25°C)	pH (Field) 2/	pH (Lab)	Collection point	Remarks
	UPPER ARTESIAN ZONE																						
	Upper Basalt Aquifers																						
	39N-5W																						
23	5bb1	8/11/65	56	25.0	tr	9.2	27.2	12.2	17.5	3.2	-	194	3.5	5.3	-	0.9	174	118	300	7.5	7.00	bypass	
24	5da1	11/17/65	-	44.2	0	250	141.0	50.1	23.1	7.6	53	tr	1111.0	tr	-	tr	1609	558	1500	5.65	2.95	tap	NH_3, 4.6 ppm.
25	8ab2	-	-	55	-	0.26	27	10	13	2.6	-	160	11	2	0.0	-	145	110	240	-	7.2	-	Calc. spec. cond.
26	9ba1	11/23/65	-	29.6	-	3.0	24.6	10.2	14.7	3.2	175	151	0.43	7.1	0.5	0.27	138	103	240	7.35	7.40	tap	NH_3, 0.3 ppm.
27	9bc4	8/12/65	-	29.2	tr	tr	29.2	8.5	17.5	3.6	-	162	2.5	3.3	-	0.00	145	108	250	7.05	7.20	tap	
28	15ac1	8/11/65	-	27.5	tr	0.23	36	10.9	22.4	3.2	-	180	15	5.3	-	0.00	182	135	300	7.85	7.50	tap	
29	15bc1	11/16/65	53	23.3	0	3-1/2	28.0	10.0	14.7	3.2	165	145	9.1	7.1	0.3	tr	147	113	240	7.25	7.50	bypass	NH_3, 0.2 ppm.
30	15ca1	8/11/65	57	26.3	tr	6.0	38.2	10.9	18.7	3.6	-	172	57.5	4.0	-	0.00	224	140	320	7.65	7.30	bypass	
31	16ad5	8/13/65	-	25.0	-	13.5	18.4	7.3	17.5	1.6	-	151	5.0	6.0	-	tr	143	118	250	7.25	7.25	tap	
32	16ba1	8/13/65	55	28.8	tr	2.9	31.2	9.1	17.5	2.8	-	184	10.5	4	-	0.2	169	115	290	7.5	7.45	bypass	
33	18ba1	1960	-	-	-	9.24	36	15.8	15.5	-	-	146	38.4	11	0.64	0.25	200	156	330	-	8.3		Calc. spec. cond. Mn, 0.22 ppm. PO_4, 0.414 ppm.
34	19ba3	11/24/65	54	22.1	tr	0.4	65.2	23.1	18.9	3.6	279	268	37.3	41.6	0.2	3.54	327	258	480	7.35	7.40	tap	NH_3, 0.2 ppm.
35	19bb4	8/13/65	-	23.3	-	tr	29.4	8.5	17.5	1.6	-	172	1.5	9.0	-	12.4	165	108	280	7.3	7.5	tap	
36	19da1	11/23/65	-	27.5	0	<1/2	25.4	8.54	16.8	2.4	210	159	57.6	7.1	0.2	2.44	199	98	280	7.25	7.10	tap	NH_3, 0.3 ppm.
37	Average		-	28-5	-	3.73	32	11	17	3.6	-	190	19.2	8.6	0.3	1.7	191	125	-	-	-		Ave. of 13 anal.
	Upper Interbed Aquifers																						
	39N-5W																						
38	5ad1	11/17/65	-	22.5	0	3-1/2	42	13.2	18.9	3.6	300	239	6.7	7.1	0.7	0.44	220	159	380	7.55	6.90	tap	NH_3, 090 ppm.
39	6dc1	11/23/65	53	27.5	0	3-1/2	32.2	11.5	14.7	2.8	189	176	7.2	7.1	0.3	0.22	168	129	270	7.2	6.90	bypass	
40	7cd1	-	-	62	-	2.5	32	14	16	-	-	161	34	7	0.35	0.3	185	137	310	-	-		Calc. spec. cond.; Ave. of 5 Anal. (Ross 1965)
41	7da2	-	-	56	-	3.8	30	13	-	-	-	140	20	2	-	-	154	128	260	-	-		Calc. spec. cond.; Ave. of 8 Anal. (Ross 1965)
42	Average		-	42	-	3.3	35.1	13.2	16.5	3.2	-	178	17	6.0	0.5	0.3	193	142	-	-	-		Ave. of 4 Anal.
	Middle Artesian Zone																						
	39N-5W																						
43	7bc1	11/24/65	68	34	0	1.2	22	11	25	6	204	190	4	7	0.6	0.22	170	101	260	7.5	7.2	bypass	NH_3, 0.30 ppm.
	Lower Artesian Zone																						
	39N-5W																						
44	7ba1	11/16/65	75	27.5	0	0.6	25.4	7.95	84	6.9	370	328	12.9	13.9	1.2	0.66	317	96	450	7.85	7.00	bypass	NH_3, 0.5 ppm.
45	8ba1	11/16/65	77	22.5	0	0.6	20.6	4.9	71.5	6.4	271	274	15.8	13.9	1.4	0.22	284	72	420	8.55	7.60	bypass	NH_3, 0.20 ppm; $CO_3^{=}$, 1.29 ppm.
46	Average		-	25.5	-	0.6	23	6.4	72	6.6	-	301	14	14	1.3	0.4	286	84	-	-	-	-	Ave. of 2 Anal.

from the pump bypass valve, without going through the storage system, and when the pump had been operating long enough for the water temperature to stabilize. Data reported on Table 10 probably are very close to actual temperature of water in the aquifer.

Iron

Accurate determination of dissolved iron actually present in the ground water circulating in an aquifer is difficult. According to Hem (1959, p. 60), iron in ground water containing bicarbonate mostly will be in the ferrous state when pH is between 6 and 8. Most of the waters reported on Table 10 contain bicarbonate and have pH's between 7 and 8. At least 100 ppm (parts per million) bicarbonate is present in 37 of the 42 analyses. When waters containing bicarbonate and iron are exposed to oxygen, as when a sample is collected at a pump, the following reaction can take place:

$$2Fe^{++} + 4HCO_3^- + H_2O + \tfrac{1}{2}O_2 \longrightarrow 2Fe(OH)_3 + 4\,CO_2$$

The removal of bicarbonate ions from solution may have marked effect on the pH of the water. If the sample is allowed to stand, the ferric hydroxide will settle to the bottom of the container. Even if the analyst shakes the container carefully prior to withdrawing the aliquot for the iron determination, some of the iron hydroxide may stick to the sides of the container and the distribution of the suspended iron hydroxide within the shaken sample may not be uniform. If the container is not shaken, the amount of iron reported may depend upon the depth into the container to which the pipette was inserted when the aliquot was withdrawn. Another problem is caused by contamination by iron rust from well casings, pumps, storage tanks, and pipes in the distribution system. The problems of iron in water are discussed by Hem (1959, p. 58-66).

Ross (1965, p. 75-76) minimized errors caused by oxidation of dissolved iron by making determinations by standard laboratory colorometric methods within 12 hours of collection. Most of the data on iron in Table 10 are from Ross or were also determined within 12 hours of sample collection. Where the sample was collected from a valve that bypassed the storage system, the data on iron on Table 10 probably are very close to the dissolved iron actually present in the water circulating in the aquifer (plus any contamination from the well or pump). The iron reported for analyses 8, 29, 36, 38, and 39 was determined in the field with a La Motte Chemical Products Company kit that is accurate to about $\frac{1}{2}$ ppm iron in the range of 0 to 4 ppm and to 1 ppm in the range of 4 to 10 ppm. Determinations were made immediately upon collection of the sample. Determinations by the field kit are reported in fractions rather than decimals.

Iron reported in analyses 2, 25, 33, 34, 40, 41, 43 and 45 on Table 10 and the "laboratory" iron reported on Table 11 were reported by the analytical laboratories. The data probably represent the iron in solution and part of the iron hydroxide in suspension at the time of analysis.

Acidization of water samples to keep iron in solution until time of analyses was attempted (Rainwater and Thatcher, 1960, p. 28). Results were erratic and time did not permit working out the problems with the acidization procedures.

Bicarbonate (Field)

Bicarbonate and carbonate content (alkalinity) of water samples can change with time after collection of the sample. In the aquifer, the carbon dioxide-bicarbonate-carbonate system is in equilibrium with respect to the pressure and temperature of the ground water. Bringing the water to the surface reduces the pressure and commonly increases the temperature; solubility of carbon dioxide in water decreases with decrease in pressure and increase in temperature. The carbon dioxide-bicarbonate-carbonate system in the water sample tends towards a new state of equilibrium with respect to atmospheric pressure and prevailing temperature, generally losing carbon dioxide in the process. (Hem, p. 46). One of the possible reactions is:

$$\underset{pH > 8.2}{H^+ + CO_3^{--}} \rightleftharpoons \underset{pH\ 4.5 - 8.2}{HCO_3^- + H^+} \rightleftharpoons H_2CO_3 \rightleftharpoons H_2O + CO_2 \uparrow\downarrow$$

Because hydrogen ions are used up by this reaction, pH of the water will change unless some buffering reaction also takes place. A bicarbonate-carbonate determination made as soon as a sample is collected directly from a pump should be very close to the actual bicarbonate-carbonate content of the water within the aquifer. If the sample is collected from a storage system, the bicarbonate-carbonate content may reflect the equilibrium under the conditions in the storage system, rather than in the aquifer.

We determined the bicarbonate-carbonate content of waters from 17 wells in the field using a LaMotte Chemical Products Company kit. End points of the titrations were determined with a Beckman N2 pH meter. The field procedure is substantilly the same as standard laboratory procedure and has equivalent accuracy. Only 6 of the samples were collected directly from the pump.

Fluoride

Most of the fluoride determinations reported on Table 10 were done with a LaMotte Chemical Products Company kit. The kit probably is accurate to 0.1 ppm fluoride. The fluorides reported for analyses 25, 33, 40, 43, and 45 were determined by the analytical laboratory. Fluoride content probably does not change with time after collection of a sample.

Specific Conductivity

Specific conductivities reported on Table 10 were measured in the field with a battery-powered Industrial Instruments "SoluBridge" meter equipped with a polystyrene dip cell. Specific conductivities were also measured by the Department of Agricultural Biochemistry and Soils on another "SoluBridge". Side-by-side comparison readings on the same water samples showed that the laboratory instrument consistently read 10 percent higher than the field instrument (Ross, 1965, p. 68). Because we have many more readings with the field instrument than the laboratory instrument, all laboratory readings were converted to equivalent field readings by deducting 10 percent from the laboratory readings. On Table 10 and 11, data from the Department of Agricultural Biochemistry and Soils are reported as equivalent field readings.

Specific conductivity can change between the time that a water sample is collected and the time of the analysis if some of the dissolved constitutents precipitate. Precipitation of calcium carbonate is an example.

pH (Field)

The pH measured at the time a sample is collected commonly differs from the pH measured when the sample is analyzed. In addition to changes caused by upsetting of the carbon dioxide-bicarbonate-carbonate equilibrium, changes may be the result of hydrolysis, oxidation, precipitation (such as calcium carbonate), or absorption of fumes from the laboratory (Rainwater and Thatcher, 1960, p. 237-238).

A pH measured at the time the sample is collected directly from the pump should be close to the actual pH of water circulating in the aquifer. If the sample is collected from a storage system, the pH may reflect conditions in the storage system rather than in the aquifer.

We measured pH in the field for 38 of the water analyses reported on Table 10 using a Beckman N2 portable pH meter. The apparatus probably was accurate to 0.1 pH unit. The water was collected directly from the pump in 13 of the samples.

The pH is one of the most important factors influencing reaction rates in aqueous systems. Data on field pH of freshly pumped waters should be useful in studies of geochemistry of ground water. Unfortunately, we were able to collect too few data to serve as a basis for conclusions.

Comparison of Field and Laboratory Data

The differences between field data and laboratory data on iron, bicarbonate, pH and specific conductivity are summarized on Table 11. Samples collected directly from a pump bypass valve are considered separately from samples collected from a storage system or dipped from an open well. The samples collected from the bypasses probably are very close to the composition of water circulating in the aquifer whereas samples collected from storage systems or dipped from open wells probably have been modified.

Because of the close relations between changes in iron, bicarbonate, and pH, the causes will be discussed together following a description of the changes. Causes of change in specific conductivity will be discussed with the description of the changes.

Iron

Field iron was at least 0.1 ppm in 20 of the samples collected. Table 11 shows that field iron was higher than laboratory iron in 15 of the 20 samples. Of the 7 samples collected directly from pump bypass valves, field iron is higher in 4 and the same as laboratory iron in 1. Of the 13 samples collected from storage systems or dipped from open wells, field iron is higher than laboratory iron in 11. These data show that at least part of the dissolved iron originally present in the samples generally had precipitated prior to the laboratory analysis. Field determination, even for water from storage systems or open wells, is more representative of the iron present in the ground waters in the aquifer than laboratory determination from samples that have been standing for various lengths of time prior to analysis.

No geochemical reason exists for laboratory iron to be higher than field iron. The 5 analyses with higher laboratory iron probably are erroneous; causes could be contaminated sample containers or mistakes by the analyst in either the field or laboratory determination.

Bicarbonate

According to Hem (1959, p. 97), reproducibility of bicarbonate-carbonate determinations (alkalinity) cannot be expected to be closer than 2 to 5 percent from duplicate samples. The data on Table 11 show differences of at least 5 percent in 14 of the 17 duplicate analyses. Field bicarbonate is higher in 12 of the 14. The data show a change in the carbon dioxide-bicarbonate-carbonate equilibrium in the samples between the time of collection and the time of analysis.

The two analyses (7 and 19) that show increases of bicarbonate in the laboratory analyses had field bicarbonates that were much lower than most of the Moscow basin waters. The increase may be due to absorption of carbon dioxide at the time the sample was collected followed by displacement of the equilibrium while the sample was in storage prior to analysis.

pH

According to Hem (1959, p. 49), accuracy of pH measurements with pH meters is 0.1 pH unit, either in the laboratory or in the field. However, differences of several tenths between field and laboratory measurements can be the results of changes in the sample after collection.

Field pH differs from laboratory pH by at least 0.1 pH unit in 31 of the 38 samples reported in Table 11. Field pH is higher than laboratory pH in 19 of the 31 samples.

For 11 of the 13 samples collected directly from the pump bypass, field pH differs from laboratory pH by at least 0.1 pH unit, and is higher in 8. For 18 of the 25 samples collected from storage systems or dipped from open wells, field pH differs from laboratory pH by at least 0.1 pH unit, and is higher in 10.

Interrelations Between Iron, Bicarbonate, and pH

Bringing a sample of Moscow basin ground water to the surface changes three physical-chemical conditions that influence the carbon dioxide-bicarbonate-carbonate equilibrium and buffering system and the solubility of dissolved iron:

1. Pressure on the water is reduced to atmospheric pressure.

2. Temperature becomes greater or smaller, depending upon relations between ground water temperature and air temperature at the time of collection and during storage prior to analysis.

Table 11. Comparison of Laboratory Data with Field Data from Chemical Analyses of Ground Waters from Moscow, Basin, Idaho

Differences: Field value minus laboratory value

Source	Number Table 10	Iron 1/ ppm	Bicarbonate ppm	Bicarbonate %	pH	Specific Conductivity Micromhos	Specific Conductivity %
Surficial Aquifers							
Samples collected from bypass valves on pumps.	3				+0.45	-10	-2
	10				-0.55	-25	-13
	12				+0.20	-20	-10
	14	-4.15	-19	-11	+0.05	+40	+17
	16	+0.24			-0.15	-40	-8
Samples collected from taps in pressure systems or dipped from open wells.	1		-5	-3	-0.1	+95	+43
	2		-17	-11	+0.35	-10	-4
	4				+0.55	-25	-5
	5				+0.6	-25	-12
	6	-1.5			+0.35	-15	-6
	7	-2.3	+22	+90	+2.15	+10	+2
	8	-1			+0.10	+10	+5
	9	-0.08			-0.2	-40	-18
	11	+0.39			+0.05	+40	+5
	13		+1	+½	+0.05	-15	-5
	15	-10.6			-0.50	-20	-10
	17				-0.2	+10	+12
	18	-1.60			-0.45	+5	+4
	19	+0.20	+7	+12	+0.15	+100	+125
	20	-0.45			-0.2	+25	-10
	21	-10.0	-11	-11	-0.4	+50	+6
Upper Artesian Zone							
Samples collected from bypass valves on pumps.	23	-7.55			-0.5	-30	-10
	29	0.0	-20	-12	+0.25	+30	+12
	30	-5.05			-0.35	-50	-16
	32	-21			-0.05	-20	-7
	39		-13	-7	-0.30	0	0
Samples collected from taps in pressure systems or dipped from open wells.	24	-241.07	-53	-100	-2.70	+210	+12
	26	-0.1	-24	-14	+0.05	-15	-6
	27				+0.15	-25	-10
	28	+0.27			-0.35	-30	-10
	31				0.0	-70	-28
	34		-11	-4	+0.05	-30	-6
	35				+0.2	-100	-36
	36		-51	-24	-0.15	+35	+12
	38		-61	-20	-0.65	-20	-5
Middle and Lower Artesian Zones							
Samples collected from bypass valve on pump	43		-14	-7	-0.30		
	44	+2.9	-42	-11	-0.85		
	45		-14	-11	-0.95		

1/ Differences in iron are shown only for those analyses in which field iron is greater than 0.1 ppm.

3. Oxygen is added to the system.

Solubility of carbon dioxide in water varies directly with pressure and inversely with temperature. Reduction of pressure causes loss of carbon dioxide until equilibrium is reached with the partial pressure of atmospheric carbon dioxide. An increase in temperature could also cause loss of carbon dioxide. On the other hand, most of the samples collected in Moscow basin were held under refrigeration from a few hours after collection until analysis. If air was present in the container, refrigeration at a temperature below ground water temperature could result in carbon dioxide going into solution.

Exposure of the sample to atmospheric oxygen is almost impossible to avoid during collection. The oxygen will oxidize ferrous iron and cause it to precipitate as ferric hydroxide, removing bicarbonate from the system in the process.

The net result of these three changes should be lower laboratory values for determinations of iron, bicarbonate, and pH. In Moscow basin samples (Tab. 11), laboratory values are significantly lower in many of the duplicate determinations:

Factor	Percent of Samples Analyzed in Which Laboratory Values are Lower than Field Values
Iron	75
Bicarbonate	85
pH	61

These data show that precipitation of iron and changing of the carbon dioxide-bicarbonate-carbonate equilibrium and buffering system were common processes during the storage of the Moscow basin samples. Other processes might also be at work. In 16 of the analyses on Table 11, at least one of the three factors is significantly higher in the laboratory analysis. Higher laboratory iron has been discounted as probably the result of errors, and too few data are available on field bicarbonate to warrant conclusions. Only two analyses (7 and 19) show significant increases in both bicarbonate and pH; 14 analyses show significant increases in pH.

Specific Conductivity

According to Hem (1959, p. 43), differences of as much as 10 percent between field and laboratory determinations of specific conductivity should not be regarded as significant. Laboratory specific conductivity shown on Table 11 has been corrected to remove the effect of the consistently 10 percent higher readings with the laboratory instrument. After the correction is applied, 13 of the 35 duplicate determinations differ by more than 10 percent; 4 are from waters collected from pump bypass valves.

No pattern is apparent; in about half of the readings, field specific conductivity is higher and in about half of the readings, laboratory specific conductivity is higher. Field specific conductivity is higher in about half of the samples drawn from the pump bypasses and is higher in about half of the samples withdrawn from storage systems or dipped from open wells. The lack of consistent pattern suggests that the differences are largely caused by mistakes in taking the readings.

One cause of mistakes would be in the setting of the temperature correction on the instrument; relatively small errors in temperature correction can lead to relatively large errors in specific conductivity.

Chemical Quality of Ground Water in Moscow Basin

Quality of Water Diagrams

To aid discussion of chemical quality of water, two standard diagrams are used: the Collins diagram and the Piper diagram. The Collins diagram is a bar diagram used to show the relations of cations and anions for typical and unusual waters from each aquifer. (For an example, see Fig. 20). Most of the Collins diagrams used are in weight units; one Collins diagram is in percent by weight.

The Piper diagram is three-field, tri-linear diagram on which an analysis in percent by weight plots as a point in each field (for an example, see Fig. 21). The left tri-angular field is the cation field, the right triangular field is the anion field, and the combined relations of cations and anions are shown in the central diamond field. Construction and interpretation of the diagrams are discussed by Hem (1959, p. 168-172; 178, 182-184).

Both of the diagrams utilize equivalent weight units (equivalents per million) rather than the parts per million used in Table 10. The Department of Agricultural Biochemistry and Soils reported analyses in equivalent weight units that were converted to parts per million by use of conversion factors in Hem (1959, p. 32). In the Collins diagrams, iron and silica in parts per million are shown by auxiliary bars.

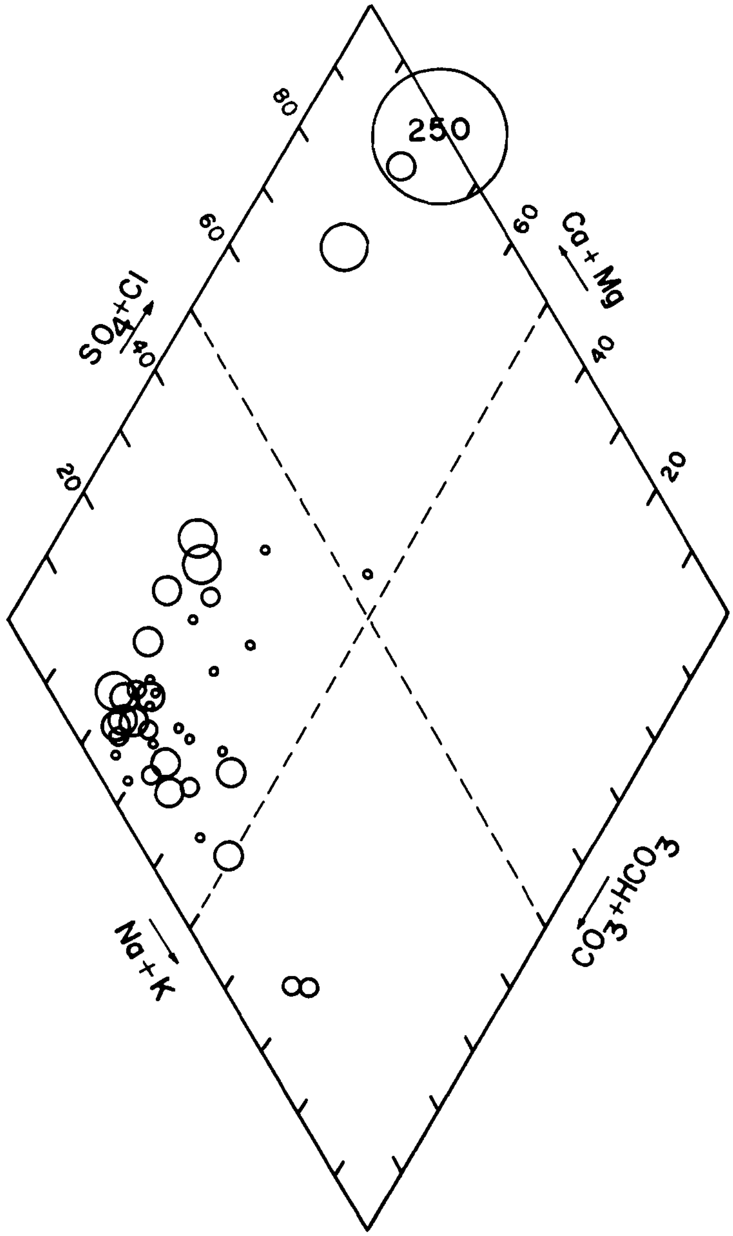
Ratios and percentages of chemical data from Table 10 (expressed in equivalent units) are given in Table 12.

General Quality of Moscow Basin Ground Waters

Most of the ground waters in Moscow basin have rather similar chemical analyses. Nearly all of them plot in the left-central portion of the diamond field of the Piper diagram (Fig. 18). Analyses plotting in this part of the field are characterized by calcium plus magnesium being more abundant than sodium plus potassium and by bicarbonate plus carbonate being more abundant than sulfate plus chloride. The percentage data on Table 12 show that in most of the analyses, calcium makes up almost half of the cations and bicarbonate makes up at least three-fourths of the anions.

In a simple classification of waters based on dominant cation and dominant anion, most of the Moscow basin waters are calcium bicarbonate waters. The three analyses that plot in the upper part of the central field are calcium sulfate waters and the two that plot in the lower part of the field are sodium bicarbonate waters.

Distribution of iron content, as shown on Figure 18, shows little relation to the distribution of major cations and anions. The three calcium sulfate waters do have high contents of iron, but so do some of the calcium bicarbonate waters. Silica contents of Moscow basin waters are relatively high compared to most ground waters. High-iron waters are common in both the surficial aquifers and the upper artesian zone in the vicinity of the clay deposits east



EXPLANATION

Area of circle in diamond field indicates iron in parts per million.

Less than 0.3 ppm iron.

03-1.0 ppm iron.

1 - 5 ppm iron.

5 - 10ppm iron.

10- 15 ppm iron.

250 more than 15 ppm iron; figures in middle are iron in ppm.

FIGURE 18.

SUMMARY OF CHEMICAL CHARACTER OF WATER FROM ALL AQUIFERS

Moscow Basin, Idaho

Table 12. Ratios and Percentages of Equivalent Units for Chemical Analyses of Ground Waters from Moscow Basin, Idaho. Numbers Refer to Analyses on Table 10

Analysis	Ca/Mg	Na/Cl	% Cations Mg	% Cations Ca	% Cations Na	% Anions HCO_3	% Anions SO_4	% Anions Cl
Surficial Aquifers								
Normal Waters								
1	2.5	2.0	20	50	28	84	2	12
2	2.0	1.8	24	48	24	84	1	16
5	3.0	4.8	14	42	43	89	3	7
6	2.7	6.5	13	35	49	88	5	7
8	2.9	6.2	17	50	30	92	3	5
9	4.3	2.5	12	53	36	83	3	13
10	2.5	5.8	20	50	29	95	0	5
12	3.4	3.8	15	52	32	84	8	8
13	2.1	4.1	18	40	40	79	11	9
14	1.3	1.5	33	41	22	86	1	12
15	2.5	3.5	22	55	22	90	4	6
16	2.9	1.4	19	56	24	73	7	20
17	2.3	3.1	17	39	41	81	0	19
18	5.4	3.0	10	51	38	88	1	11
19	2.1	1.8	20	45	34	74	7	17
20	2.3	5.3	22	52	24	91	5	4
22(ave)	2.6	3.2	18	48	32	86	4	10
Waters that may be contaminated								
3	2.5	0.8	21	48	30	58	8	34
4	1.7	1.4	24	40	35	69	8	24
11	2.8	1.4	14	39	46	45	13	42
21	2.4	0.4	24	58	17	23	11	65
Abnormal Water								
7	2.0	4.0	27	55	16	8	88	4

Analysis	Ca/Mg	Na/Cl	% Cations Mg	% Cations Ca	% Cations Na	% Anions HCO_3	% Anions SO_4	% Anions Cl
Upper Basalts								
Normal Waters								
23	1.4	4.4	32	44	22	93	2	5
25	1.7	11.0	29	48	19	91	8	1
26	1.5	3.5	29	43	25	89	3	7
27	2.1	7.0	23	50	24	95	2	4
28	2.0	5.6	25	49	25	87	9	6
29	2.0	3.5	27	46	23	86	7	7
30	2.0	6.3	25	53	21	69	29	3
31	2.0	5.9	25	41	31	89	7	4
32	2.0	5.8	24	51	23	90	6	4
33	1.5	1.8	33	49	18	67	22	11
34	1.2	0.8	31	53	15	69	12	18
35	2.0	2.6	24	51	24	90	1	8
36	1.8	4.0	25	45	28	89	4	6
37(ave)	1.8	3.0	27	48	22	83	11	6
Abnormal Water								
24	1.7	---	33	56	9	4	96	0
Upper Interbeds								
38	1.9	4.5	26	50	23	92	3	5
39	1.6	3.5	29	48	22	89	4	6
40	1.4	5.9	33	46	20	75	20	5
41	1.4	---	33	46	21	83	15	2
42(ave)	1.7	4.2	30	48	19	85	10	5
Middle Basalts								
43	1.3	5.3	28	35	32	91	3	6
Lower Basalts								
44	2.0	10.0	16	20	65	89	4	6
45	2.6	9.0	8	21	71	88	6	6
46(ave)	2.2	8.0	11	23	63	88	5	7

of Moscow (Figs. 22 and 26). Origin of iron and silica in Moscow basin waters will be discussed below.

Some waters in Moscow basin, especially in the surficial aquifers, show evidence of contamination from fertilizers or septic tank effluents. High-nitrogen fertilizers commonly are used in farming in the vicinity of Moscow and septic tanks are numerous outside of the city limits. According to Hem (1959, p. 118), where nitrate in water is derived from organic pollution, the high nitrate commonly is accompanied by high chloride. In Moscow basin, waters believed to be free from contamination generally contain less than 1 ppm nitrate and less than 20 ppm chlorides. Higher nitrates alone suggest contamination from fertilizers whereas higher nitrates plus high chlorides suggest contamination from septic tank effluent or barn-yard wastes. For the purposes of our Moscow basin study, we suggest that waters containing more than 10 ppm nitrates but less than 30 ppm chlorides should be suspected of possible contamination from fertilizers whereas waters containing more than 10 ppm nitrates plus more than 30 ppm chlorides should be suspected of possible contamination from septic tank effluent.

Relations of Specific Conductivity to Total Dissolved Solids

The relation between specific conductivity and total dissolved solids may be expresses as (Hem, 1959, p. 49):

$$(\text{s.c.})\ (A) = (\text{T.D.S.})$$

If specific conductivity is expressed in micromhos/cm at 25° C, total dissolved solids will be in parts per million. The factor "A" will have a value ranging from 0.5 to 1.0 and, unless the water is of unusual composition, will be between 0.55 and 0.75. In a group of waters of generally similar composition that have common origin, A will approach a constant and can be used to convert specific conductivities to reasonable approximations of total dissolved solids.

In Figure 19, field specific conductivity of the analyzed samples of Moscow basin waters are plotted against calculated total dissolved solids. Waters that are thought to be contaminated are omitted. Waters from the surficial aquifers are plotted separately from the waters from the artesian aquifers. In both plots, the analyses fall very close to a straight line and the slope of the line indicates that the value of A is 0.60 for waters both from the surficial and the artesian aquifers.

Distribution of pH

Ross (1965, p. 64-65) briefly discussed pH of waters in Moscow basin. She published an isogam map of pH in the "upper confined aquifer" that indicated regular changes from one part of the aquifer to another. The map was based on unpublished data in the files of the Idaho Bureau of Mines and Geology. Source of the data and method of determination are not known.

The pH data that we collected, especially the field pH data, do not show any simple, regular change in pH from one part of of an aquifer to another. The differences between laboratory pH and field pH indicate the unreliability of pH data from unknown sources obtained by unknown means.

Distribution of pH in the Moscow basin ground waters is more complex than it seemed to be at the time of Ross' preliminary report. The data at hand at the end of our study do not justify any attempts at isogam maps of pH in Moscow basin ground waters.

Quality of Water from the Surficial Aquifers

Ground waters from two different geologic environments, surficial sediments and crystalline rocks, have been included in the surficial aquifers. Little significant differences are present between the waters from the two environments. Relatively high nitrates and chlorides in several analyses probably are the result of contamination.

Four of the analyses are compared in a Collins diagram in Figure 20. Analyses 10 and 13 are typical waters; analysis 11 shows possible contamination; analysis 7 is an abnormal natural water. Analysis 10 is from well 39N-5W-12ab1, a dug well about 50 feet deep in weathered granitic rock near the upper edge of Moscow basin. Total dissolved solids is 127 ppm, about average for the uncontaminated waters from the surficial aquifers. Analysis 13 is from well 39N-5W-16bc1, a dug well 18 feet deep in surficial sediments near the center of the basin. Total dissolved solids is 208 ppm, which is high for uncontaminated waters from the surficial aquifers, but chlorides and nitrates are low. Analysis 11 is from well 39N-5W-9bc1, a dug well 18 feet deep in surficial sediments near the center of Moscow basin. Total dissolved solids is 500 ppm and chlorides and nitrates both are high; the water in this well may have been contaminated by septic tank effluent. Analysis 7 is from well 39N-5W-9cc1, a drilled well 160 feet deep in surficial materials and perhaps into weathered granitic rocks in the area of high-iron waters near the clay deposits east of Moscow (Figs. 22 and 26). Total dissolved solids is 321 ppm, rather high for uncontaminated waters from the surficial aquifers. The water is abnormal in that sulfate is the dominant anion; iron is high, 3.6 ppm. Nitrates and chlorides are normal; the water probably is not contaminated.

Relations of 21 analyses of water from the surficial aquifers are shown on a Piper diagram in Figure 21. High-nitrate analyses 3, 4, 11, and 21 also are high chloride and indicate possible contamination from septic tank effluent; high nitrate analysis 16 is barely below the minimum value that indicates septic tank effluent contamination; high-nitrate analyses 2 and 9 indicate possible contamination from fertilizers. Analysis 7 plots near the upper vertices of the central field and anion field. Except for analyses 7 and 21, the analyses with more than 0.3 ppm iron plot to the left of the central field and cluster near the bicarbonate vertex of the anion field; however, these analyses are scattered in the cation field. When analyses 3, 4, 7, 11, and 21 are excluded, the average of the analyses plots well to the left of the central field.

Only 8 of the 21 analyses, or 38 percent, show more than 0.3 ppm iron. Iron contents of water from 80 wells and springs in the surficial aquifers (Ross, 1965) are shown in Figures 22 and 26. Distribution of the iron contents is:

Iron ppm	Number of Analyses	Iron ppm	Number of Analyses
0.0-0.1	48	3-10	8
0.1-0.3	12	10-15	2
0.3-1.0	10		

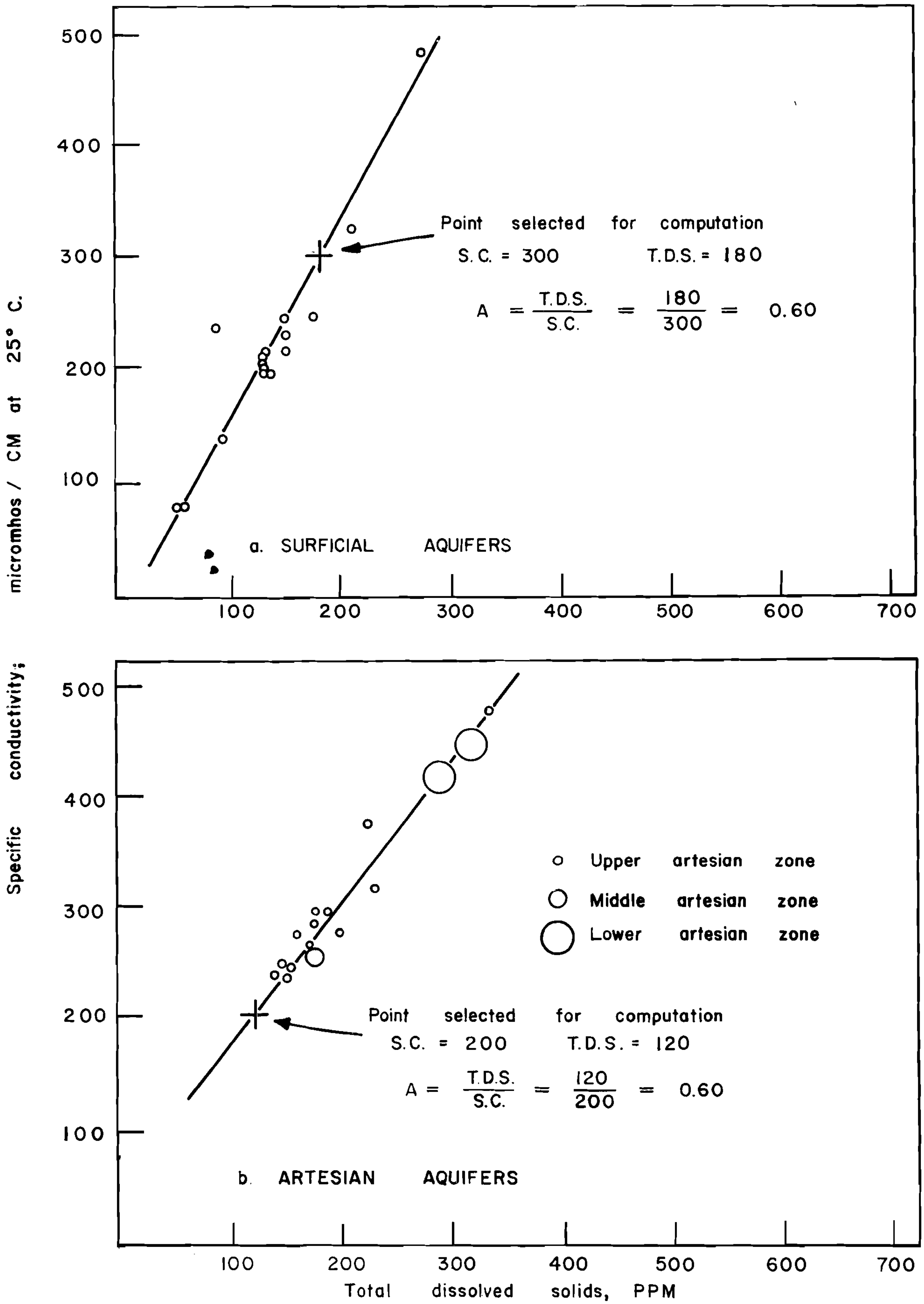


Figure 19. Relations between specific conductivity and total dissolved solids, Moscow basin, Idaho. Analyses with more than 10 PPM chlorides or nitrates omitted because of possible contamination

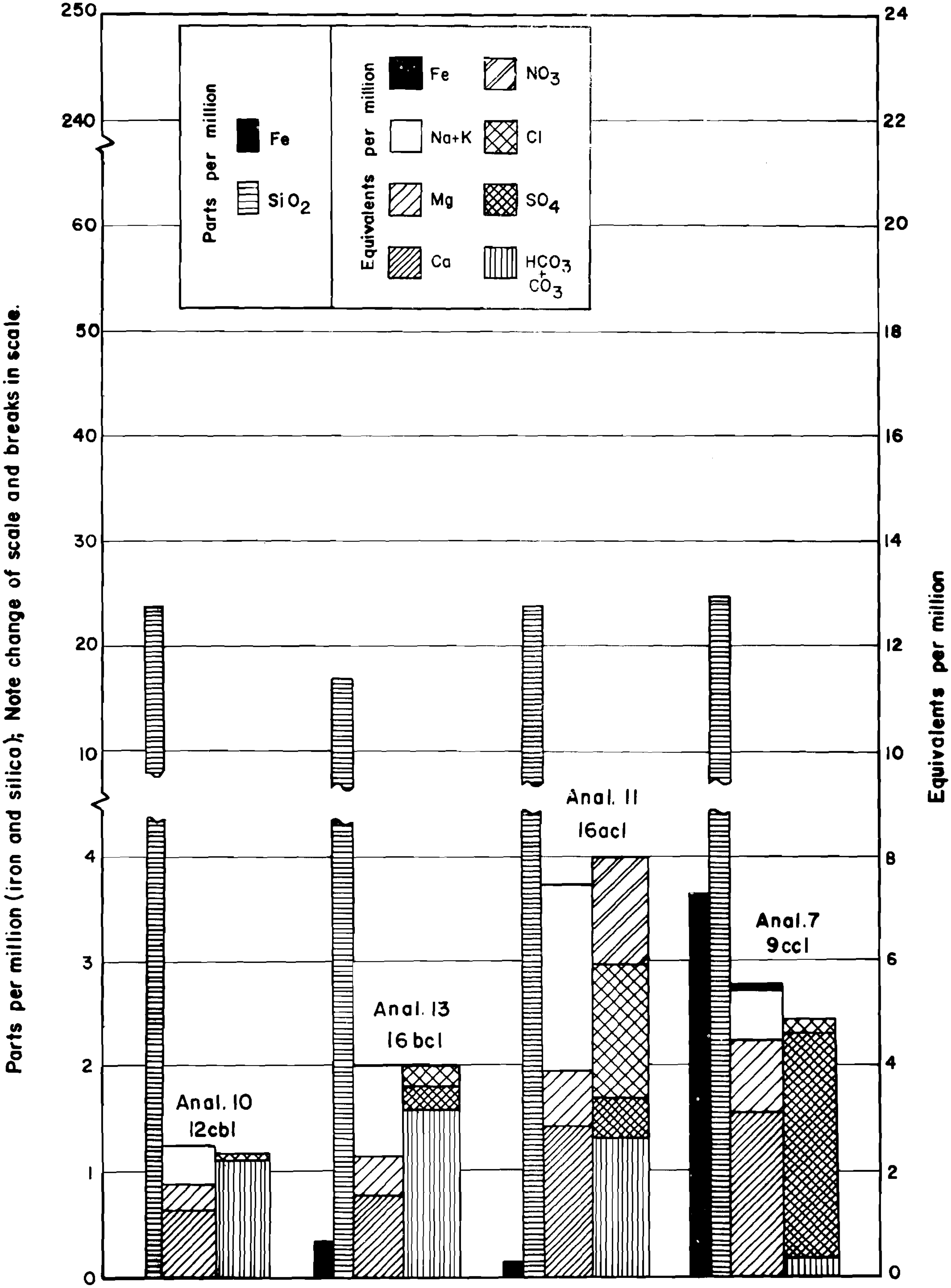
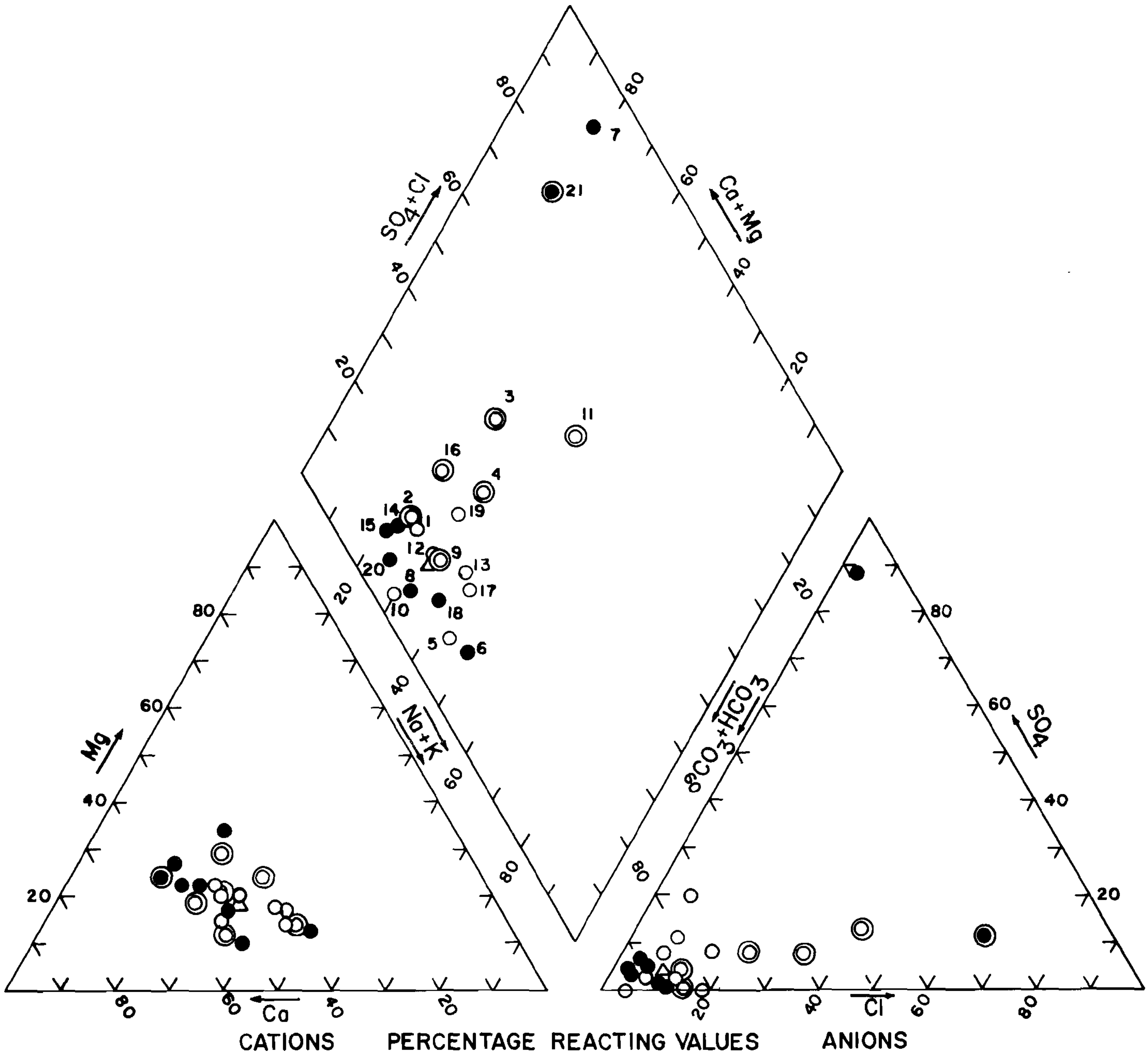


Figure 20. Chemical character of waters from several wells in the surficial aquifers, Moscow Basin, Idaho



○ Analysis; number refers to number of analysis on Table 10.

● Analysis, more than 0.3ppm iron.

◎ Analysis, more than 10ppm NO_3.

◉ Analysis, more than 0.2ppm iron and more than 10ppm NO_3.

△ Average of 16 analyses of waters probably not influenced by contamination (well 9cc$_1$ excluded).

FIGURE 21.

CHEMICAL CHARACTER OF WATER FROM WELLS IN THE SURFICIAL AQUIFERS

Moscow Basin, Idaho

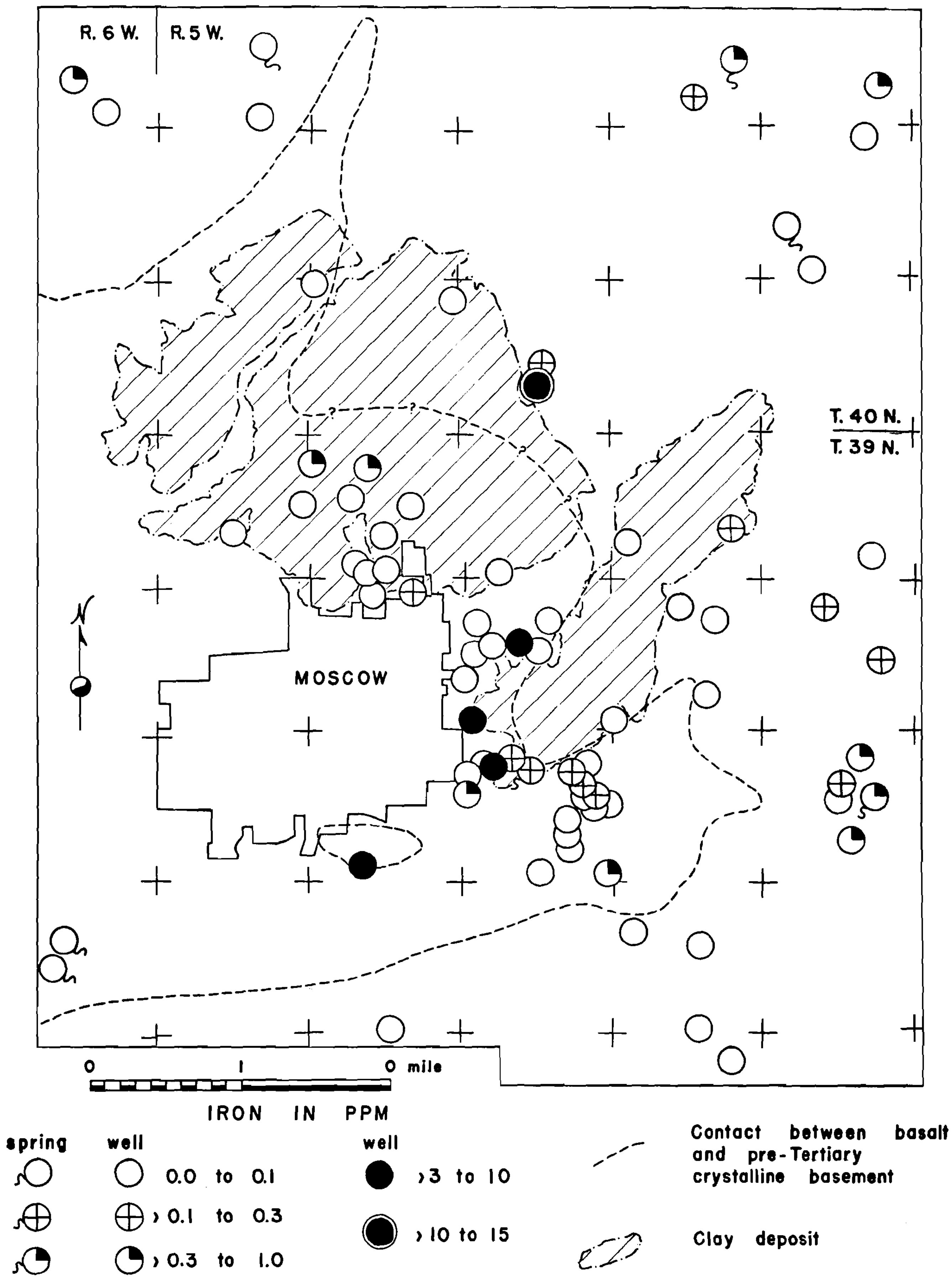


Figure 22. Iron in water in surficial aquifers in part of Moscow basin, Idaho

These data indicate that excessive iron is only moderately common in waters from the surficial aquifers; only 20 of 80 analyses, or 25 percent, have iron in excess of the U.S. Public Health Service recommended maximum of 0.3 ppm. The somewhat higher percentage shown in Figure 21 probably is the result of selective sampling of high-iron waters. The data points on Figure 22 seem too scattered to justify drawing of contours. However, higher values seem to be more common in the vicinity of the high-iron clay deposits to the east of Moscow.

Distribution of specific conductivity of waters from about 135 wells in the surficial aquifers (Ross, 1965) is shown in Figure 23. Two controls of specific conductivity are present; one is natural and the other is artificial. The natural control is length of travel as groundwater and length of time that the water has been in contact with the rocks. Waters at the outer--and higher--margins of the basin generally have lower specific conductivity than waters nearer the center of the basin; part of the waters in the center of the basin have moved down the water-table gradient from higher areas in the basin. This control is shown by the 100 and 200 micromho isopleths.

The artificial control is contamination that causes local, anomalously high readings. In drawing the isopleths for Figure 23, readings that seemed anomalously high for the position in the basin were ignored because of possible contamination. However, some of the anomalous readings could be the result of natural processes. The 300 micromho isopleth may be influenced by wells which have water that is affected by contamination. The area inside of the 300 micromho isopleth contains many septic tanks and some farm land; several chemical analyses from wells inside the 300 micromho isopleth contain both high nitrate and high chloride.

Quality of Water from the Upper Artesian Zone

We studied the waters of the upper basalt separately from the waters of the upper interbed to see if any significant differences were present. None were detected. Waters from only two wells in the upper artesian zone showed evidence of possible contamination.

Four of the analyses are compared on a Collins diagram in Figure 24. Analysis 32 is typical water from the upper basalt; analysis 39 is typical water from the upper interbed; analysis 40 is more-or-less normal water from the upper interbed that shows the highest silica content reported from Moscow basin; and analysis 24 is an abnormal water from the upper basalt that has the highest iron content reported in Moscow basin.

Analysis 32 is from well 39N-5W-16ba1, a drilled well in basalt 160 feet deep near the contact with the crystalline basement and in the area of high-iron waters near the clay deposits east of Moscow. Total dissolved solids is 169 ppm, which is rather low compared to the average for waters from basalts. Analysis 39 is from well 39N-5W-6bc1, a drilled well 376 feet deep in the upper interbed near the central part of Moscow basin and about a mile from the contact with the crystalline basement. Total dissolved solids is 168 ppm, which is low for waters from the upper interbed. Analysis 40 is the average of 5 analyses (Ross, 1965, Tab. XVI) of waters from well 39N-5W-7cd1 (University 2), a drilled well 355 feet deep in the upper interbed near the center of Moscow basin. Total dissolved solids is 185 ppm, about average for waters from the upper interbed. Silica was reported in only 1 of the 5 analyses; therefore, the 62 ppm value lacks verification. Differences between analysis 40 and and analyses 32 and 39 are small; analysis 40 does show slightly higher sulfate.

Analysis 24 is from well 39N-5W-5da1, a drilled well 186 feet deep in the upper basalt near the contact with the crystalline basement and in the area of high-iron waters near the clay deposits east of Moscow. Total dissolved solids is 1609 ppm, almost a thousand ppm higher than that of any uncontaminated Moscow basin water. Iron content is very high; 2 analyses by R. W. Jones were 240 and 260 ppm. An analysis in the files of the Moscow office of Culligan Soft Water Service showed 268 ppm (Culligan analysis N216786, File No. 11-120, Culligan Laboratories, Northbrook, Illinois; analyzed October 7, 1965). The analysis by the Department of Agricultural Biochemistry and Soils shows only 12 ppm, but the analyst stated that the results were not accurate owing to precipitation of very large amounts of iron hydroxide. In addition to high iron and high total dissolved solids, analysis 40 also is abnormal in that sulfate is by far the most abundant anion.

The high-iron water shown in analysis 24 has limited areal extent. Analysis 38 is from well 5ad1, which is only a few hundred yards from well 5da1. Well 5ad1 is deeper--405 feet--and taps the upper interbed. Iron in analysis 38 is $3\frac{1}{2}$ ppm, a value that is rather high but within the range of values shown by a number of other wells in Moscow basin. The waters in analysis 38 are normal calcium bicarbonate waters. Nearby wells in the surficial aquifers, 5ab1 and 5 ac1, apparently have normal waters although complete analyses were not obtained. Well 5ab1 is a drilled well 80 feet deep tapping a sand bed in the surficial aquifers; field specific conductivity is 195 micromhos and field pH is 6.95. Iron was not determined but no evidence of high iron content was present in the water. Well 5ac1 is a drilled well 110 feet deep; field specific conductivity is 215 micromhos, field pH is 7.15, and field iron is 0.0 ppm.

Relations of 18 analyses of waters from the upper artesian zone are shown on a Piper diagram in Figure 25. Compared to the analyses from the surficial aquifers (Fig. 21), most of the analyses from the upper artesian zone plot within a smaller area, indicating a greater uniformity of quality of water. The area in which the upper artesian zone analyses plot is overlapped by the area in which the surficial aquifer analyses plot.

Very little evidence of contamination is present in the upper artesian zone analyses. Analysis 35, the only analysis with more than 10 ppm nitrate, has only 9 ppm chlorides and plots in the middle of a group of low-nitrate analyses. Analysis 34, the only analysis with more than 30 ppm chloride, has only 3.54 ppm nitrate. Analysis 35 may show evidence of contamination from fertilizers, but analysis 34 is below the limits that are considered evidence of contamination for both nitrate and chloride.

Unlike the surficial aquifers, the analyses with more than 0.3 ppm iron do not group, but are scattered in all three fields of the Piper diagram. The anion triangle shows that several analyses (30, 33, 34, 40) have sulfate

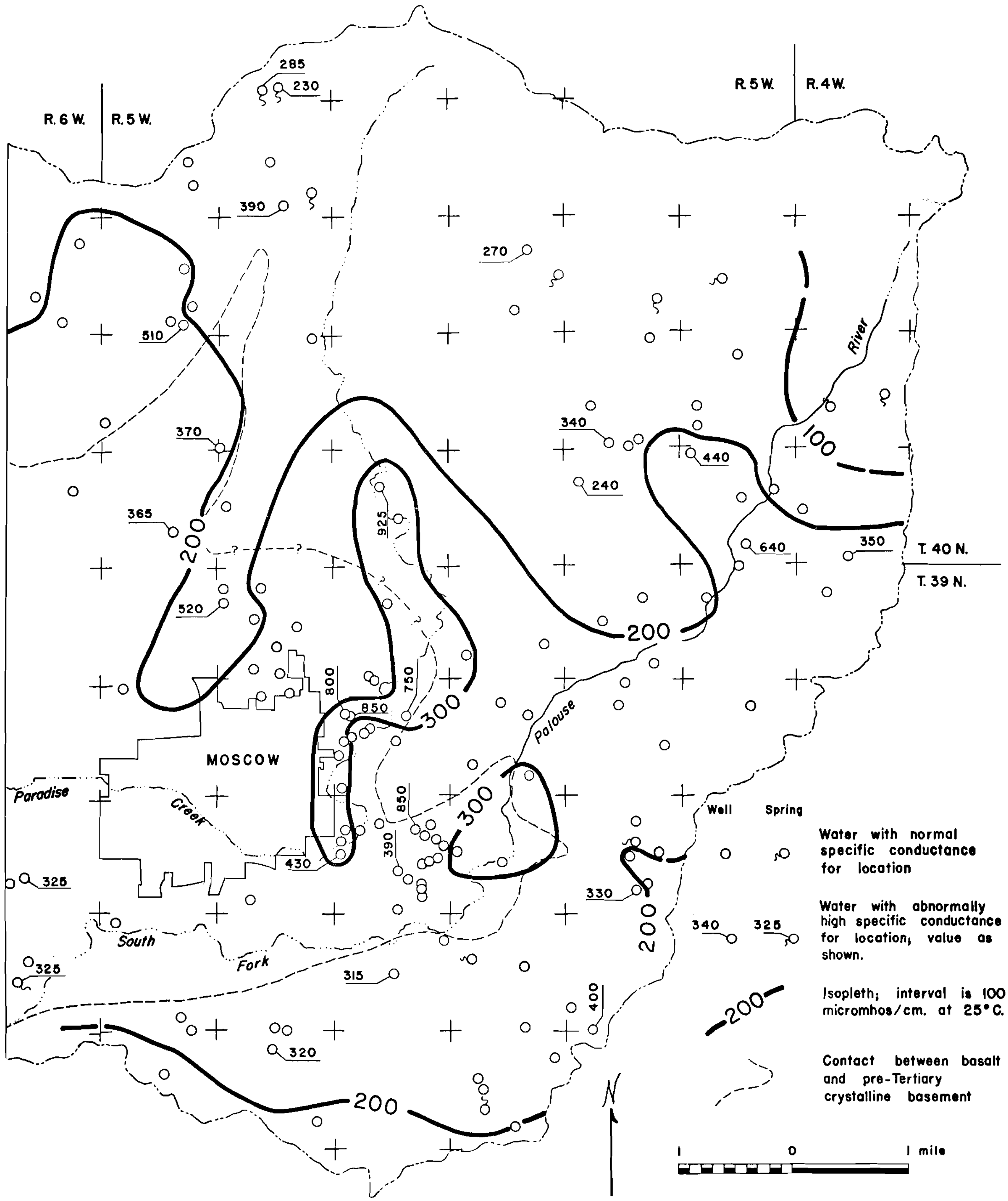


Figure 23. Specific conductance of water in surficial aquifers, Moscow basin, Idaho.

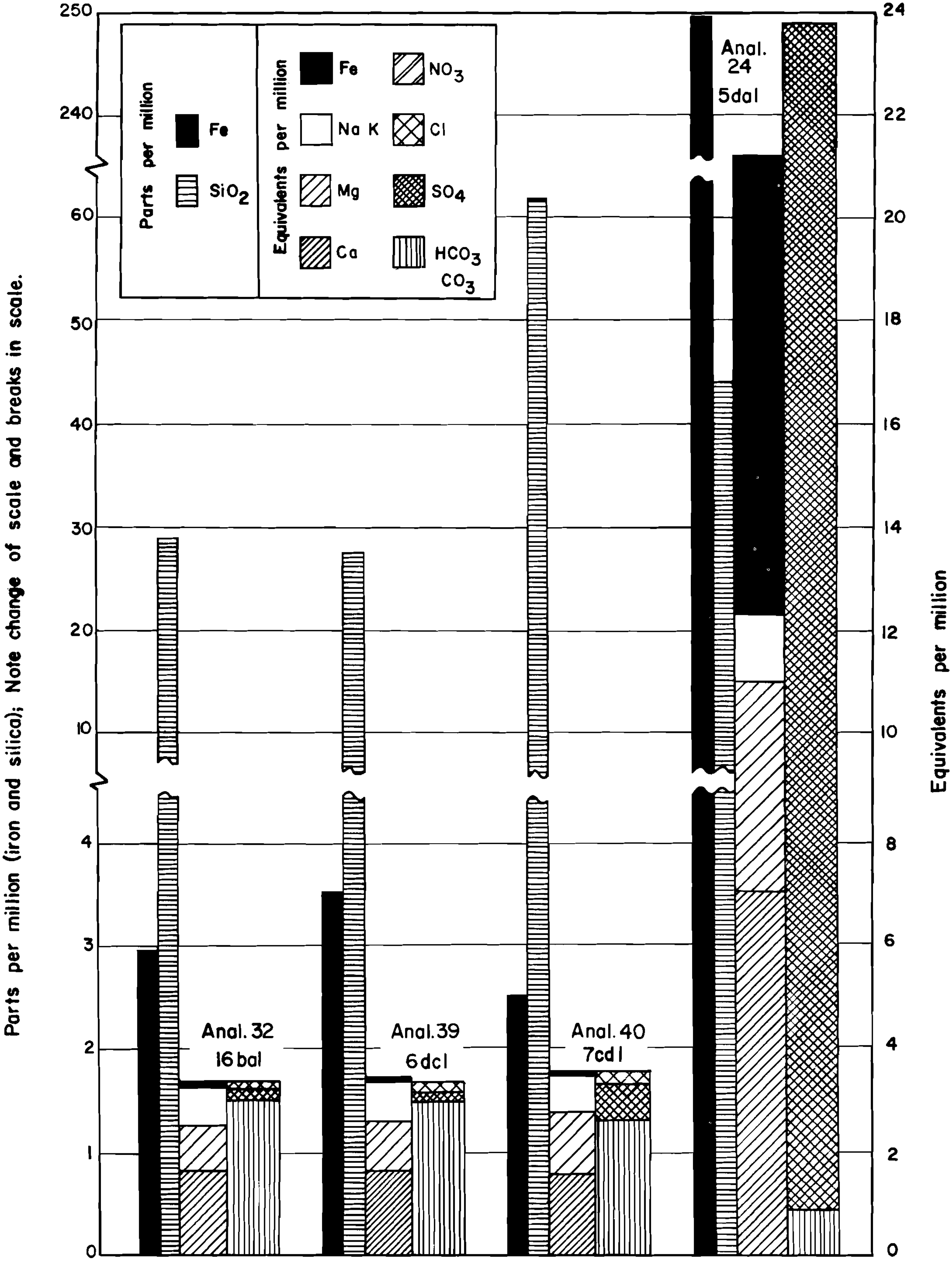
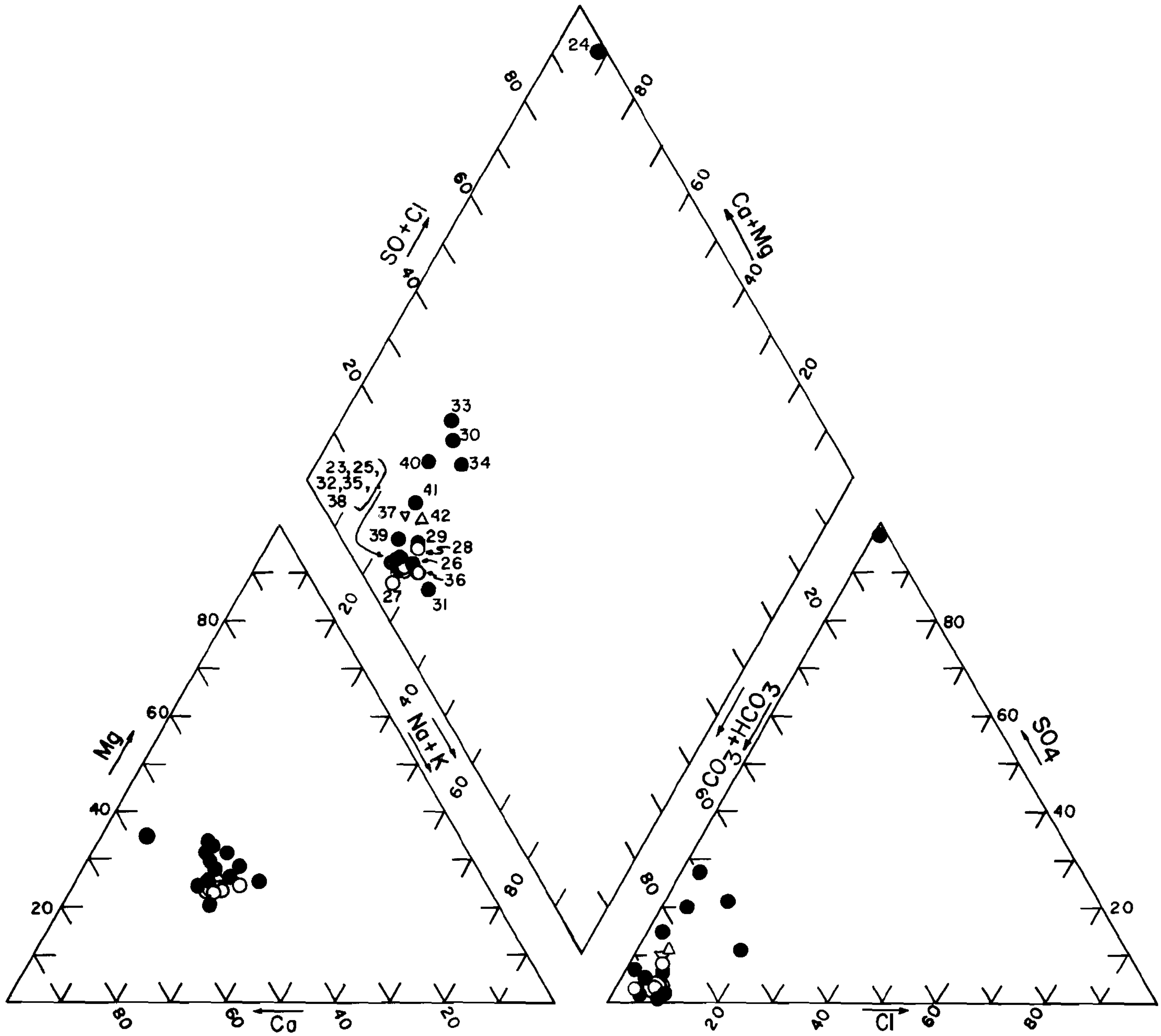


Figure 24. Chemical character of water from several wells in the upper artesian zone, Moscow Basin, Idaho.



○ Analysis; number refers to number of analysis on Table 10.

● Analysis, more than 0.3ppm Fe.

◎ Analysis, more than 10ppm NO_3.

△ Average of 13 analyses of waters from upper basalt aquifers.

▽ Average of 4 analyses of waters from upper interbed aquifers.

FIGURE 25.

CHEMICAL CHARACTER OF WATERS FROM WELLS IN THE UPPER ARTESIAN ZONE, Moscow Basin, Idaho.

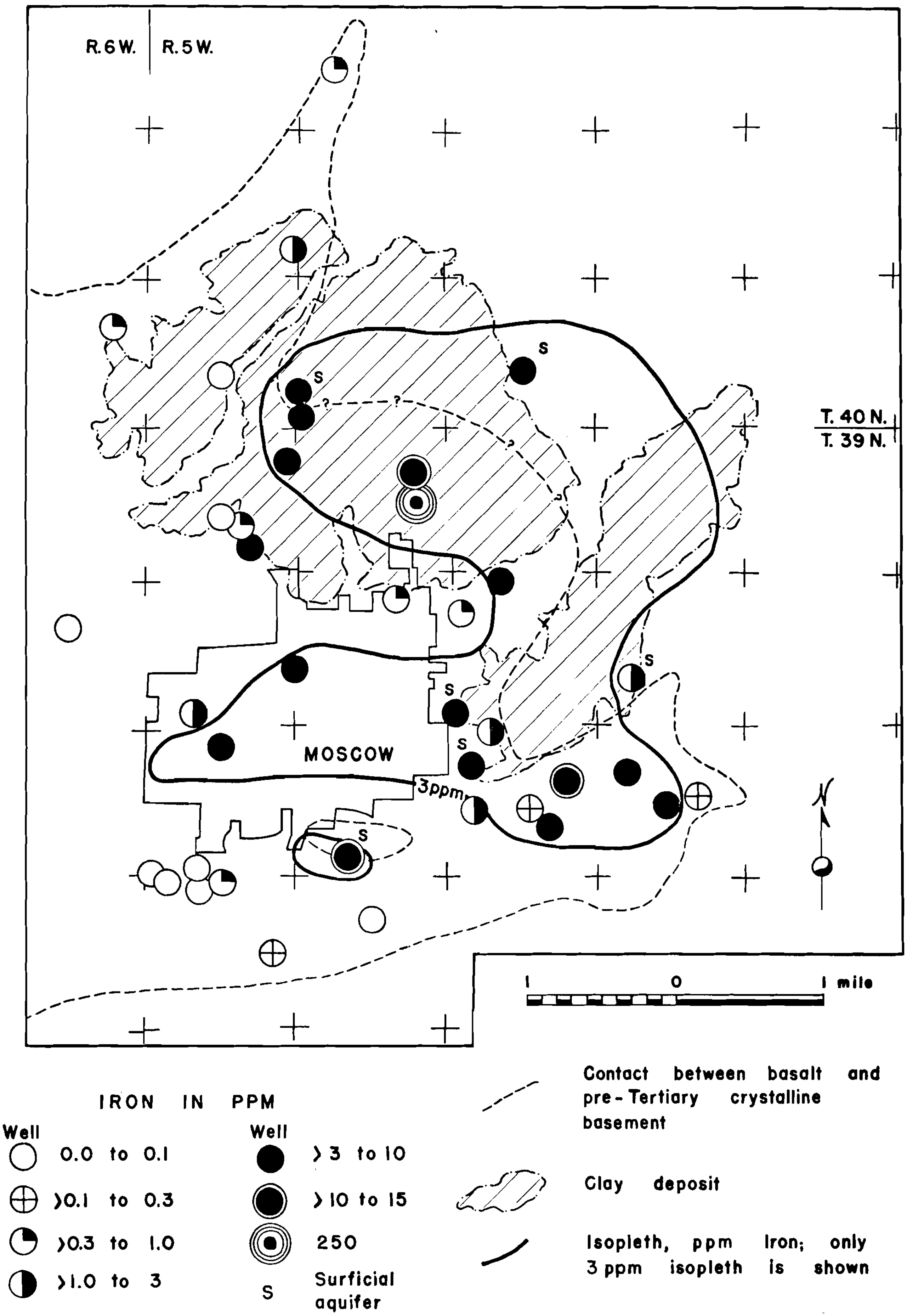


Figure 26. Iron in water in upper artesian zone and part of surficial aquifers, Moscow basin, Idaho

contents that are relatively higher than in any of the normal waters of the surficial aquifers. Analysis 24 plots near the upper vertices of the anion field and central field because of the dominance of the sulfate ion. When analysis 24 is excluded, the two average analyses (37, average for waters from upper basalt and 42, average for waters from upper interbed) plot very close together in all fields.

Of the 18 analyses plotted in Figure 25, 14 (78 percent) show more than 0.3 ppm iron. Iron contents of waters from 33 wells in the upper artesian zone (Ross, 1965) are shown in Figure 26. Distribution of the iron contents is:

Iron ppm	Number of Analyses	Iron ppm	Number of Analyses
0.0-0.1	8	3-10	8
0.1-0.3	3	10-15	2
0.3-1.0	7	250	1
1.0-3	4		

These data show that excessive iron is very common in waters from the upper artesian zone; 22 of the 33 analyses (66 percent) show iron in excess of the 0.3 ppm maximum recommended by the U.S. Public Health Service. The somewhat higher percentage of high-iron waters shown on Figure 25 is the result of selective sampling of high-iron waters. The 3 ppm isopleth on Figure 26 encloses an area in which most of the waters analyzed contain more than 3 ppm iron. This area seems to have a close relation to the high-iron clay deposits.

Distribution of specific conductivity of waters from about 25 wells in the upper artesian zone (Ross, 1965) is shown in Figure 27. Values of specific conductivity that seemed anomalous for their position in the basin were ignored in drawing the isopleths. It should be noted that the unusual waters shown in analysis 24 indicate that at least some of the anomalous values are natural.

Ross (1965, p. 71) noted areas of consistently high specific conductivity along the lower reaches of the South Fork Palouse River and suggested that the higher values are the result of high specific conductivity waters percolating down from the river. These areas are shown by the 300 and 400 micromho isopleths on Figure 27. Infiltration studies by Williams and Allman (1969) show that recharge does take place along the lower reaches of South Fork Palouse River during the spring runoff. A closure of the 300 micromho isopleth in the western part of the City of Moscow is based on calculated specific conductivities for University Wells 1 and 2. If the data are correct, the higher specific conductivity could be the result of longer time underground and greater length of travel of the water.

Quality of Water from the Middle and Lower Artesian Zones

Only one analysis is available of water from the middle artesian zone and only two are available from the lower artesian zone. The three analyses are shown on a Collins diagram in Figure 28 and on a Piper diagram in Figure 29.

Analysis 43 is from well 39N-5W-7bc1 (University 3), a drilled well 1336 feet deep that obtains water from the middle artesian zone through perforations in the casing between depths of 661 and 776 feet. Total dissolved solids is 170 ppm, somewhat lower than the average for the upper artesian zone. The Collins diagram shows a somewhat different cation distribution compared to the upper artesian zone and surficial aquifers: sodium is relatively more abundant and calcium less abundant; magnesium is about the same. The differences in sodium and calcium cause the analysis to plot further towards the sodium plus potassium vertex of the cation field of the Piper diagram and to plot below the area of the upper artesian zone analyses in the central diamond; analysis 43 is just within the lower part of the area of the surficial aquifer analyses in the central diamond field.

Analysis 44 is from well 39N-5W-7ba1 (City Well 8), a drilled well producing from depths of 1047 to 1263 feet in the lower artesian zone. Analysis 45 is from well 39N-5W-8ba1 (City Well 6), a drilled well producing from depths of 982 to 1308 feet in the lower artesian zone. Total dissolved solids are 317 ppm in analysis 44 and 284 ppm in analysis 45, about 100 ppm higher than the averages from the upper artesian zone. On the Collins diagram, the two analyses show cation patterns very similar to each other but very different from any other Moscow basin ground water: sodium is the dominant cation. Bicarbonate is still the dominant anion. On the Piper diagram, analyses 44 and 45 plot almost at the same point. In the cation field, they plot much further towards the sodium vertex than any other analysis because of the dominance of sodium. In the central field, they plot well below the areas in which the other analyses plot.

Analyses of water from Pullman, Washington wells (Foxworthy and Washburn, 1963, Fig. 11, p. 30) plot in about the same place on a Piper diagram as analysis 43 (middle artesian zone). None of the waters analyzed in the vicinity of Pullman plot in the same area as analyses 44 and 45 (lower artesian zone).

Geochemical Evolution of Ground Water

Hem (195, 201-230) presents a relatively extensive discussion of the relations of quality of water to geologic conditions. Relations of water quality to mineralogical composition of the rocks in an aquifer depend upon the stability toward water or solubility in water of the individual mineral species present in the rock. A very soluble mineral present only in trace amounts may have profound influence on the quality of ground water. Differences in climate, or in weathering can produce very different types of waters from essentially similar original rocks. Because of these complications, quality of water can be related to rock types only in a general way.

Water which falls as precipitation along the upper parts of Moscow basin can move through a variety of rock types as it migrates down into the artesian aquifers in the central part of the basin. Some of the water passes through a sequence of:

1. Soil derived from crystalline rocks or from loess.
2. Weathered crystalline rocks--granite or quartzite.
3. Unweathered crystalline rocks.
4. Surficial materials--loess; surficial sediments derived from crystalline rocks or from loess.

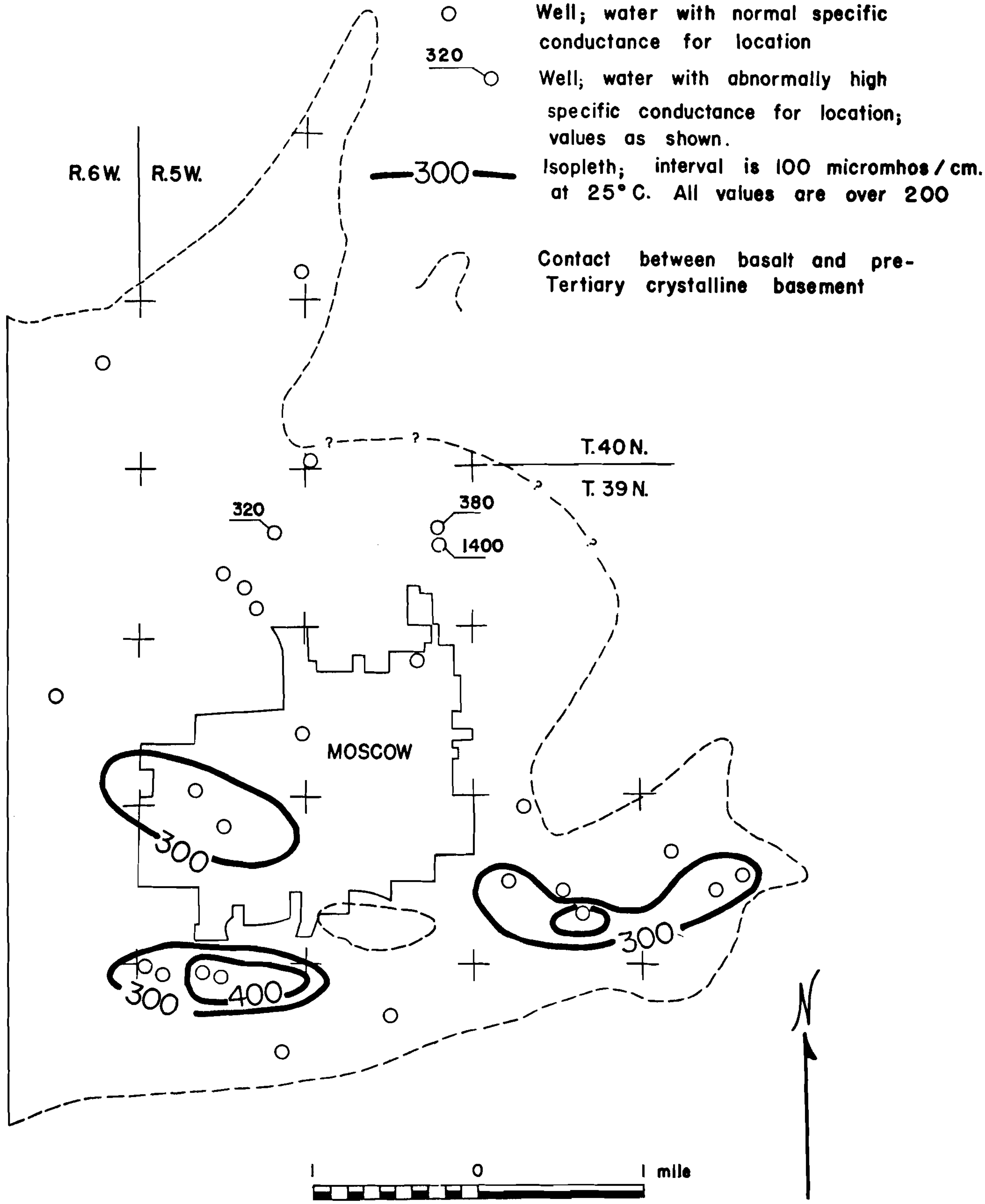


Figure 27. Specific conductance of water in upper artesian aquifer, Moscow basin, Idaho.

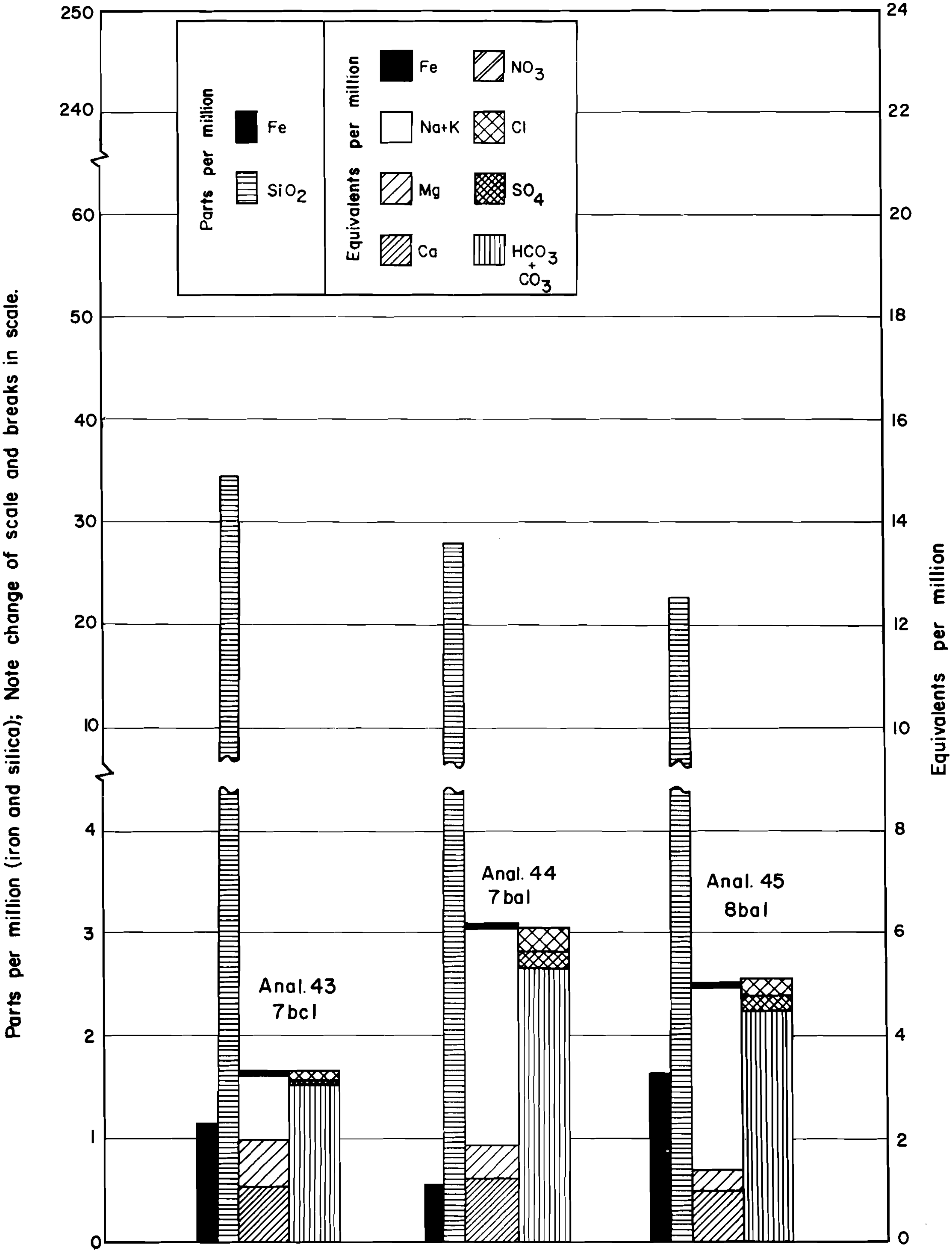
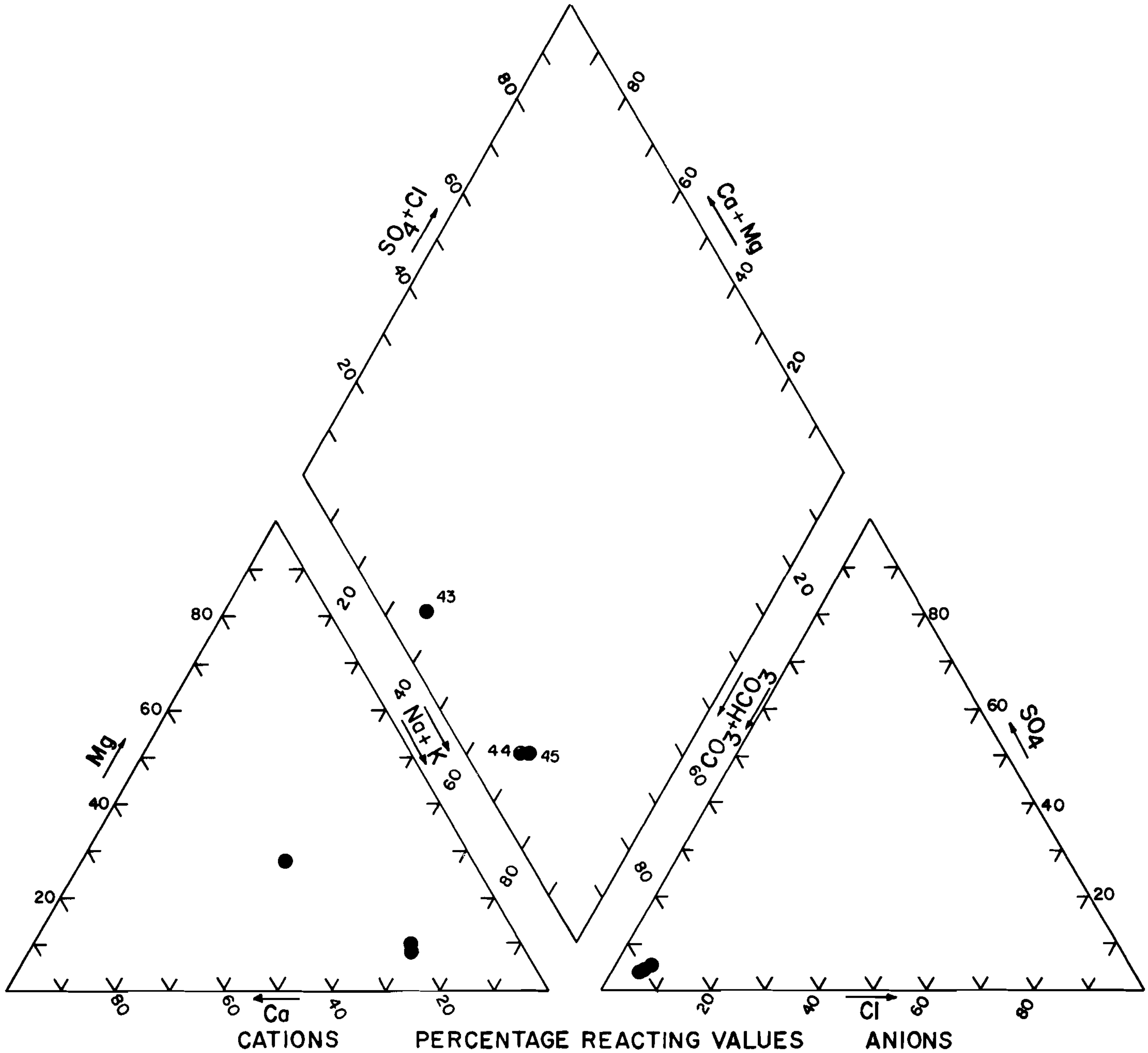


Figure 28. Chemical character of water from several wells in the middle and lower artesian zones, Moscow Basin, Idaho.



● Analysis, more than 0.3ppm iron.
Number refers to number of analysis
on Table 10.

FIGURE 29.

CHEMICAL CHARACTER OF WATER FROM MIDDLE AND LOWER ARTESIAN ZONES, Moscow Basin, Idaho

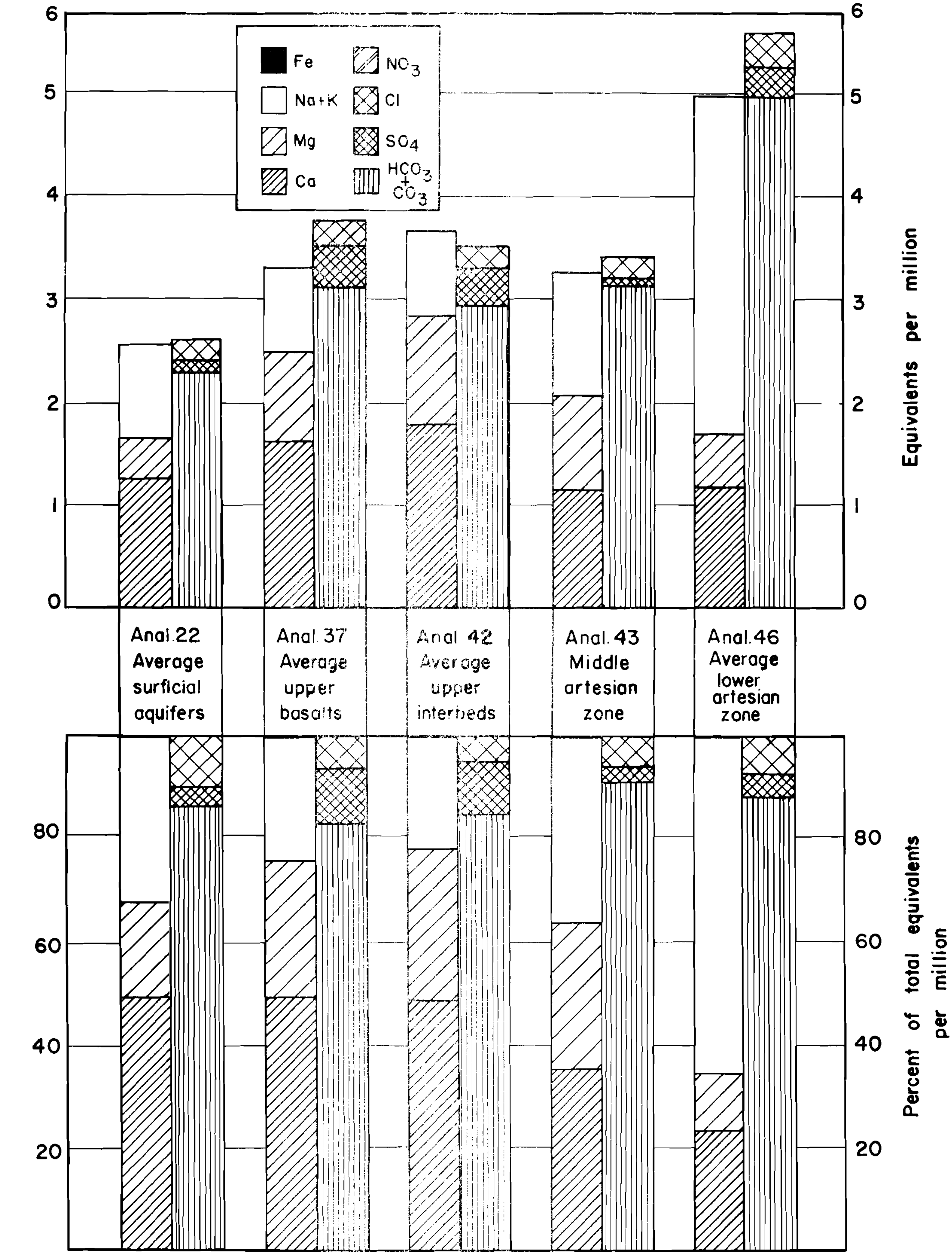


Figure 30. Relations of chemical character of water from different aquifers Moscow Basin, Idaho

5. Interbeds--sedimentary materials derived from crystalline rocks.

6. Basalt--both weathered and unweathered.

This sequence is further complicated by the influence of past climates. The climate was warmer and wetter when the basalts were laid down and an event of tropical weathering was responsible for the clay deposits east of Moscow (Hosterman and others, 1960, p. 11). In Moscow basin, relations between quality of water and mineralogy of aquifers are complex.

Origin of Principal Ions

The four principal ions of Moscow basin ground waters are calcium, magnesium, sodium, and bicarbonate. The most logical origin for calcium, magnesium, and sodium is the weathering of feldspars and mafic minerals. The bicarbonate probably came from carbon dioxide picked up as the water fell through the atmosphere or moved down through the zone of areation above the water table. Calcium and bicarbonate also could have come from weathering of calcite veins in the crystalline rocks.

Granitic rocks and feldspathic sedimentary rocks derived from granitic rocks have relatively high sodium content because of the presence of large amounts of feldspars in which sodium is more abundant than calcium and of presence of only small amounts of calcium-or magnesium-rich minerals such as biotite, hornblende, or pyroxene. Weathering of the feldspar tends to make a great deal of sodium available to be picked up by circulating ground waters. Lesser amounts of calcium and magnesium are made available from the feldspars and the mafic minerals.

Basalts have relatively high calcium and magnesium content because of the presence of large amounts of feldspars in which calcium is more abundant than sodium and of large amounts of calcium-and/or magnesium-rich minerals such as pyroxene and olivine. Weathering of basalt tends to make a great deal of calcium and magnesium available to be picked up by circulating ground waters, along with lesser amounts of sodium.

Relations between the chemical quality of water of each of the aquifers are shown in two kinds of Collins diagrams in Figure 30. The upper diagram is in weight units (epm) and the lower diagram is in percent by weight cations and anions. Comparisons of differences between average analyses from each aquifer are shown in Table 13.

The average analysis of waters from the surficial aquifers (22) contains only a little more sodium by weight than the average analysis of waters from the upper basalts (37). The water from the upper basalt contains substantially more magnesium, calcium, bicarbonate, sulfate, and total solids. The difference in sodium between the two analyses probably is not significant; the actual difference in weight units is only 0.05 epm even though it makes a difference of 11 percent. The increase in weight units of calcium and magnesium is significant. The differences suggest that relatively sodium-rich waters with rather low total solids moved down into the upper basalt and picked up additional dissolved solids. The cations picked up are calcium and magnesium, which are the cations that should be available in basalt. The increase in bicarbonate could be the result of a change in the carbon dioxide-bicarbonate-carbonate equilibrium when the positively-charged magnesium and calcium ions became available to the system. The increase in sulfate could be from the alteration of pyrite; pyrite is present in the basalts and clayey interbeds of the upper artesian zone.

Figure 30 and Table 13 show that water moving down from the upper artesian zone through the middle artesian zone to the lower artesian undergoes changes that cannot be explained in terms of the major minerals present in the rocks. Total solids go up, as is to be expected with increasing distance of travel as ground water. However, calcium and magnesium, the two cations that would be expected in increase in basalt aquifers, actually decrease, both in percent and in weight units.

The data show no change in magnesium between the upper and middle artesian zones, but a decrease between the middle and lower artesian zones. Calcium decreases between the upper and middle artesian zones and decreases again between the middle and lower artesian zones whereas sodium increases between each of the zones. Bicarbonate is about the same in the upper and middle artesian zones, but increases substantially in the lower artesian zone. Total solids decreases somewhat between the upper and middle artesian zones, but then increases substantially in the lower artesian zone. The anomalous distribution of cations probably is the result of cation exchange accompanied by precipitation and solution of ions.

Cation Exchange; Precipitation and Solution of Ions

Cation exchange, also known as base exchange, is the exchange of cations in solution in water for cations loosely held in the crystal lattices of the minerals in the aquifer. The exchange process is reversible and follows the law of mass action; however, divalent cations are held more tightly than monovalent cations in the crystal lattices. If a water containing abundant divalent cations passes through an aquifer containing an abundance of mineral grains with loosely-held monovalent cations, the divalent cations will be exchanged for monovalent cations. Principles of cation exchange are discussed by Hem (1959, p. 219-223).

In Moscow basin, the divalent cations calcium and magnesium are exchanged by the monovalent cation sodium as the water moves down through the middle and lower artesian zones. The minerals involved in the exchange probably are clay minerals. In the process, the waters are naturally softened.

The exchange processes and associated precipitation and solution of ions are shown on a Piper diagram in Figure 31. The changes are indicated by arrows connecting the plots of the average analyses in the cation field and central field. Because the Piper diagram is based on percentages and does not show changes in dissolved solids, special symbols are used in the central field to show the changes in total dissolved solids and in hardness.

Beginning with the plot of analysis 22, the average analysis of waters from the surficial aquifers, an increase in relative amount of magnesium as the water passes down into the upper artesian zone is shown by the average analyses of the upper artesian zone (analyses 37 and 42) plotting on the same percent calcium line as analysis 22. Although the calcium by weight increases by 0.26 epm, the total dissolved solids also increases so that percentage calcium

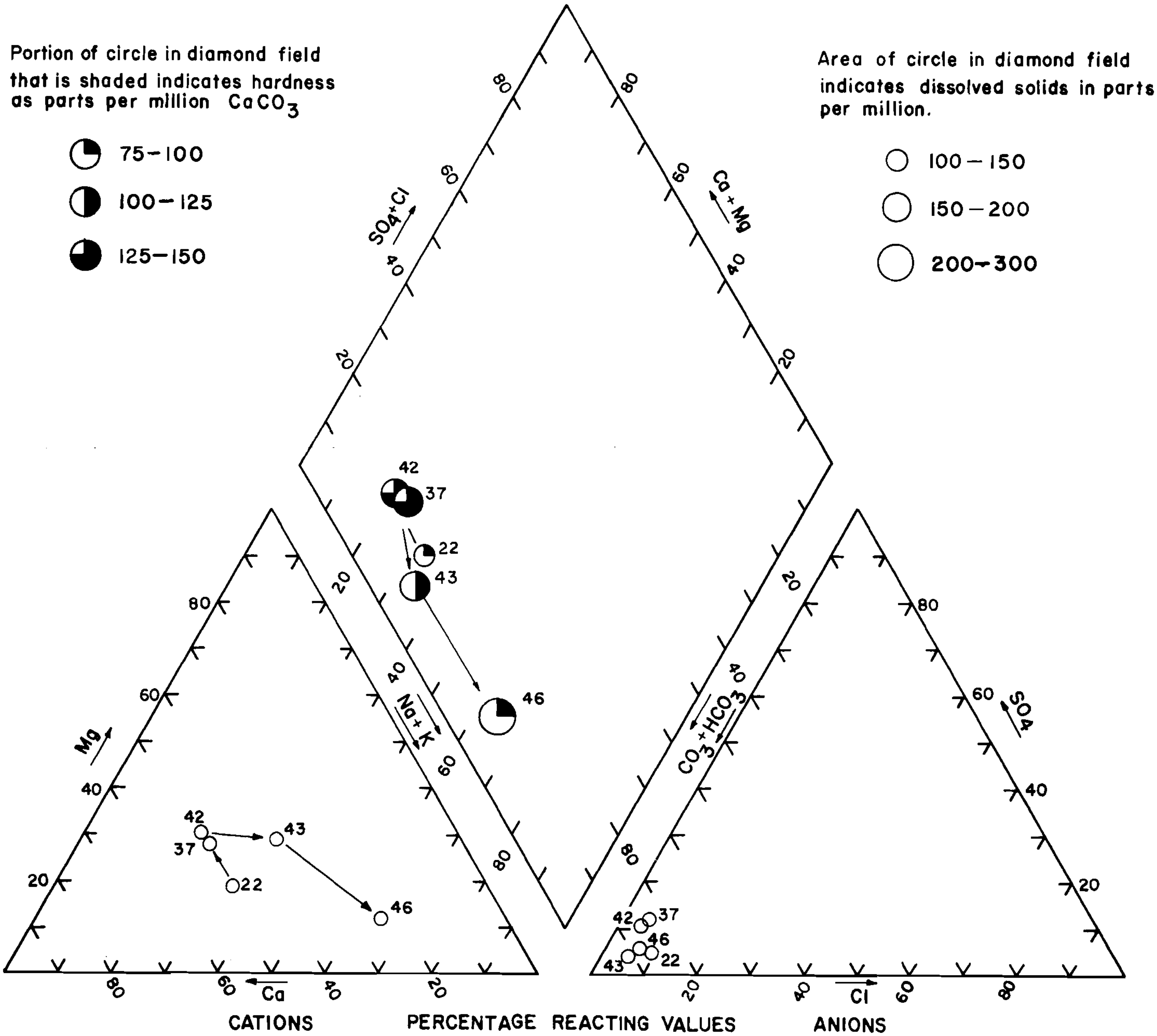


FIGURE 31.

GEOCHEMICAL EVOLUTION OF GROUND WATERS, Moscow Basin, Idaho

Numbers refer to number of analysis on Table 10.

22. Surficial aquifers, average of 16 analyses.

37. Upper basalt aquifers of upper artesian zone; average of 13 analyses.

42. Upper interbed aquifers of upper artesian zone; average of 4 analyses.

43. Middle artesian zone, 1 analysis.

46. Lower artesian zone, average of 2 analyses.

Table 13. Comparisons of Differences Between Average Analyses of Waters from the Different Aquifers, Moscow Basin, Idaho

A. Differences between surficial aquifers and upper basalt aquifers.

	Mg		Ca		Na + K		HCO_3		SO_4		Cl		Total solids
Analysis	epm	percent cation	epm	percent cation	epm	percent cation	epm	percent anion	epm	percent anion	epm	percent anion	epm
22 Surficial	0.48	18	1.24	48	0.87	32	2.28	86	0.11	4	0.26	10	5.24
37 Upper	0.89	27	1.60	48	0.82	22	3.12	83	0.40	11	0.24	6	7.07
	+0.41	+9	+0.26	00	-0.05	-11	+0.29	-3	+0.29	+7	-0.02	-4	+1.83

B. Differences between upper basalts and middle artesian zone.

	Mg		Ca		Na + K		HCO_3		SO_4		Cl		Total solids
Analysis	epm	percent cation	epm	percent cation	epm	percent cation	epm	percent anion	epm	percent anion	epm	percent anion	epm
37 Upper	0.89	27	1.60	48	0.82	22	3.12	83	0.40	11	0.24	6	7.07
43 Middle	0.90	28	1.12	35	1.20	37	3.11	91	0.09	3	0.20	6	6.62
	+0.01	+1	-0.48	-7	+0.38	+15	-0.01	+8	-0.31	-8	-0.04	0	-0.45

C. Differences between middle artesian zone and lower artesian zone.

	Mg		Ca		Na + K		HCO_3		SO_4		Cl		Total solids
Analysis	epm	percent cation	epm	percent cation	epm	percent cation	epm	percent anion	epm	percent anion	epm	percent anion	epm
43 Middle	0.90	28	1.12	35	1.20	37	3.11	91	0.09	3	0.20	6	6.62
46 Lower	0.52	11	1.15	23	3.29	65	4.95	88	0.29	5	0.39	7	10.59
	-0.38	-27	+0.03	-12	+2.09	+28	+1.84	-3	+0.20	+2	+0.19	+1	+3.97

remains the same. The increase in total dissolved solids is from 127 to 191 and 193 ppm and the hardness increases from 87 to 125 and 142 ppm as calcium carbonate. The increase in hardness reflects the increase in calcium and magnesium by weight, probably by a solutional process. Lack of change in sodium by weight suggests that no base exchange processes are involved.

Cation exchange and/or precipitation as the water passes into the middle artesian zone are shown by the analysis from the middle artesian zone, analysis 43, plotting very nearly on the same percent magnesium line as the analyses from the upper artesian zone. Calcium decreases and sodium increases, both by weight and by percent, but the change in sodium is not as large as the change in calcium. Much of the decrease in calcium is the result of exchange for sodium. The rest of the decrease in calcium might be linked with the decrease in sulfate, suggesting precipitation of calcium sulfate-gypsum. Precipitation is suggested by the decrease in total dissolved solids from 191 and 193 ppm to 170 ppm. Exchange of calcium by sodium has naturally softened the water, reducing the hardness from 125-142 ppm to 101 ppm as calcium carbonate.

Further cation exchange plus solution of additional ions as the water passes into the lower artesian zone is shown by the average analysis of the lower artesian zone, analysis 46, plotting near the sodium vertex of the cation field and having a higher total dissolved solids--but lower hardness--compared to the middle artesian zone. Calcium by weight is essentially the same as in the middle artesian zone but magnesium by weight is substantially reduced. Sodium by weight and bicarbonate by weight are substantially increased. Total dissolved solids increased to 286 ppm. Removal of magnesium could be the result of a base exchange process that produced part of the sodium and caused the reduction in hardness to 84 ppm as calcium carbonate. When the reduction in equivalent weight of magnesium is subtracted from the increase in equivalent weight of sodium, the remaining increase in sodium is 1.71 epm, which nearly balances the increase of 1.84 epm of bicarbonate.

The increase by weight of sodium and bicarbonate might seem to be the result of direct solution of sodium bicarbonate in the lower artesian zone or of the solution of sodium ions accompanied by a further shift in the carbon dioxide-bicarbonate-carbonate equilibrium system. However, no source of sodium bicarbonate is known in the lower artesian aquifers of Moscow basin; mineralogy and geology seem to be very similar to the upper and middle artesian zones. The higher total dissolved solids probably is simply the result of greater length of time in contact with rocks as ground water. Perhaps the increase in sodium is a two-step process. At some level between the middle and lower artesian zones, the descending water picked up readily available calcium and magnesium ions that caused a shift in the carbon dioxide-bicarbonate-carbonate equilibrium that produced more bicarbonate ions and, for the first time in Moscow basin, carbonate ions (analysis 45). Then, as the waters continued to descend, calcium and magnesium were replaced by sodium through cation exchange processes.

Origin and Distribution of Iron and Silica

Origin and distribution of iron in the ground water was one of the original goals of our study. We hoped that we would be able to relate iron contents to some simple chemical or physical-chemical factors in the ground waters. We did not succeed.

Ross (1965, p. 83-88) determined the geohydrologic factors in the origin and distribution of iron in the surficial aquifers and upper artesian zone and showed that the somewhat higher than normal silica is related, in part, to origin of the iron.

Relation of Iron Content to Chemistry of Water

At the time that we did our field work, relations of solubilities of various species of iron minerals had been worked out in relation to pH and oxidation potential (Garrels and Christ, 1965, p. 178-223). Back and Barnes (1965) reported on a field study where iron concentration varied systematically with oxidation potential but showed little correlation to pH.

We had a portable pH meter, but we did not have the time to develop techniques for measuring oxidation potentials of ground water in the field under the conditions that prevail in Moscow basin. Our pH data at least are conformable with the results of Back and Barnes. No relation to iron content is apparent.

We attempted to relate iron content to pH, to concentrations of other ions, to ratios of ions, and to temperature of ground water. Relations that were examined included field iron to pH (both laboratory and field), to bicarbonate (both laboratory and field), to sulfate, to the ratio of bicarbonate to sulfate, to chloride, and to temperature of ground water. On all plots, the points were scattered and no relations were apparent between the factors plotted.

Origin of Iron in Ground Waters of Moscow Basin

When the areal distribution of iron is plotted, however, it becomes evident that high iron occurs either in water from granitic rocks or in water from wells that are along the basalt-granitic rock contact. Almost all high iron wells, whether in granitic rock or basalt, are in an area between the South Fork of the Palouse River and Paradise Creek. In addition, the highest amounts of iron occur in the upper artesian zone (Fig. 26).

Much of this area that is completely encircled by a zone of anomalously high iron content is underlain by the Canfield-Rodgers clay deposit (Hite, 1930; Schied, 1940' Hosterman and others, 1960). Most other high iron areas are also in proximity to clay deposits. The eastern portion of the upper artesian zone is recharged almost exclusively through these deposits.

Shallow wells and deep wells in areas away from clay or highly weathered granitic rock generally are relatively iron free. Thus, it seems that iron is related to clay deposits.

According to Hosterman and others, (1960, p. 11-20), clay deposits in northern Idaho and eastern Washington can be divided, by origin, into three types: (1) residual clay formed from basalt, (2) residual clay from granitic rocks, and (3) transported clay formed from granitic rocks. Table 14 compares the chemical composition of the three types. The percentages of total ferric oxide (Fe_2O_3) and available ferric oxide are of special importance in determining the origin of iron in the ground water. According to Hosterman and others (1960, p. 10-11):

The ferric oxide (Fe_2O_3) in a clay deposit may be locked in unweathered pyroxene, amphibole, biotite, magnetite, and ilmenite, but it may also be in such weathering products as geothite, hematite, limonite, and iron-bearing clay minerals such as nontronite. Therefore, some of the total ferric oxide is available. By definition, available ferric oxide is the percentage by weight of the ferric oxide in the calcined clay that is soluble in a 20 percent solution of sulfuric acid after boiling for 1 hour...In general, the good drainage conditions that help to produce kaolinite flush out the iron oxides. The iron oxides in residual clay deposits usually increase with depth as weathering effects become less pronounced.

The largest portion of the Canfield-Rodgers deposit is transported clay derived from granitic rock; some of the clay is residual from granitic rock, and a very small portion is residual from basalt. In the transported clay, almost all of the ferric oxide is classified as available. Much of this available ferric oxide should be soluble by percolating ground water, although the exact mechanism of solution is not known.

Water in the surficial aquifers moves relatively slowly, and high iron water cannot move far from the source. Normally, water high in iron in the upper confined zone does not move far from the clay deposits because of the piezometric surface depression near the aquifer boundary.

However, during periods of heavy pumping of this aquifer, the piezometric depressions are enlarged, or deepened, or both and are displaced farther west of the aquifer boundary. Thus, ground water high in iron from the clay deposit moves away from the deposit, and is pumped from wells farther west in Moscow basin. The effect was particularly noticeable in the University of Idaho wells during the heavy pumping period of the 1950's.

Table 14

Average Chemical Properties of the Three Types of Clay in Northern Idaho Clay Deposits

	Residual Clays		Transported Clays
Property	Basaltic (percent)	Granitic (percent)	Granitic (percent)
Ignition loss (700° C)	10.6	6.1	8.6
Total Al_2O_3	30.1	21.3	24.8
Available Al_2O_3	28.5	16.6	22.3
Recovery Al_2O_3	94.9	77.9	89.9
Total Fe_2O_3	9.4	4.0	4.9
Available Fe_2O_3	5.4	3.1	4.2
Recovery Fe_2O_3	57.4	77.5	85.7
Total SiO_2	42.4	68.7	58.5
Total TiO_2	6.4	0.4	1.0
Moisture at 130° C	29.0	14.0	24.0

1 After Hosterman and others, 1960, p. 19.

Although large amounts of iron from surficial clay deposits do not now seem to affect the middle and lower artesian zones, it is probable that iron content in water from these zones will increase at some future date. At least some of the recharge to these lower aquifers must come through the clay deposits at the eastern edge of the aquifers; increased lowering of the piezometric surfaces should increase the amount of recharge from this area.

In addition, some silica undoubtedly also comes from weathering processes that affect the basalt. This silica would be in water that is relatively iron free, such as that from the middle artesian zone. Hem (1970. Table 12, p. 106) cites two analyses of water from wells in Columbia River basalt that are high in silica, yet low in iron.

Origin of Silica in Ground Water in Moscow Basin

Silica (SiO_2) in almost all Moscow basin water is higher than the median value of 17 mg/l for ground water quoted by Davis (1964). However, most Moscow basin analyses are in the commonly observed range of 1 to around 30 mg/l (Hem, 1970, p. 109). The U.S. Public Health Service does not place a limit on silica in drinking water because it is not considered a health hazard in quantities found in most natural waters.

Silica, both as quartz and in non-crystalline forms, is a major constituent of most rock types. Quartz, which is almost pure crystalline silica, has a solubility at normal ground water temperatures and pH's of about 6 mg/l. However, the solubility of silica in amorphous form has been reported to be as high as 115 mg/l at 25° C (Morey and others, 1964).

Hem (1970, p. 104-105) points out that "it seems probable...that most of the dissolved silica...in natural water results originally from the chemical breakdown of silicate minerals in processes of weathering." The igneous rocks of Moscow basin, formed at temperatures and pressures much different from those to which they are now subjected, are relatively prone to chemical change. Our studies of well logs and cuttings show that weathered granitic rocks may extend more than 250 feet below ground surface.

Uniformly high silica content occur in water from all artesian zones as well as from the surficial aquifers. Even where iron content is relatively low, silica is still high.

The weathering of granitic rock and clay minerals probably contributes the greatest amount of silica to the water. Altschuler and others (1963, p. 148-152) show that low temperature conversion of iron-and alumina-rich montmorillonite to kaolinite releases more than enough silica to account for all of the silica in Moscow basin waters. Clays in the interbed aquifers between basalt flows, as well as clay from the Canfield-Rodgers deposit, would serve as a source of the silica.

RELATIONS OF MOSCOW BASIN STUDIES TO MOSCOW-PULLMAN GROUND WATER PROBLEMS

Our studies have been restricted to the Moscow sub-basin of the larger Moscow-Pullman ground-water basin of Idaho and Washington. The Moscow-Pullman basin can be considered a hydrologic system in which two centers of pumpage are present--the Moscow center caused by pumpage at the City of Moscow and the University of Idaho and the Pullman center caused by pumpage at the City of Pullman and Washington State University. The results of our studies of the Moscow sub-basin must be evaluated in terms of the entire basin. In particular, we must examine the effect of pumpage at the Pullman center on water levels at the Moscow center and on interpretations of model aquifer studies of the Moscow sub-basin. In addition, we will describe a model that could be used to study the complete Moscow-Pullman basin.

Effect of Pumpage at Pullman on Water Levels at Moscow

The apparently very large aquifer constants of the Columbia River Basalt at Moscow indicate that radii of cones of influence of wells are very large (Table 4). Mutual interference may exist between the well fields at the two pumping centers and also between the image well array for the two centers. Therefore, water level fluctuations in the Moscow sub-basin may include a component that is the result of pumpage at the Pullman center. The interference probably is on a braod scale related to the overall operation of the two well fields rather than to starting and stopping of pumps on individual wells. During our studies, continuous water level recording gages were operated for at least short periods of time on wells in each of the three artesian zones of the Moscow sub-basin. We did not see any water level fluctuations that we are able to relate to operation of individual wells at Pullman (See also Sokol, 1966).

In our section on water level fluctuations in the Moscow sub-basin, we concluded that the water table in the surficial aquifers of Moscow basin has not declined in spite of the large decline of water levels in the upper artesian zone. We interpreted this to mean that recharge is at least sufficient to balance the pumpage in the Moscow sub-basin. The same interpretation can be applied to the Moscow-Pullman basin: lack of decline of water levels in the surficial aquifers through 1965 indicates that recharge was sufficient to balance the combined pumpage of Moscow and Pullman through 1965. The only available estimate for recharge to the Moscow-Pullman basin that utilizes the equation of hydrologic equilibrium is that of Packer (1955). His figure of 3.5 billion gallons annually is greater than the anticipated demand for the Moscow-Pullman basin in the year 2000. If Packer's estimate is reasonably correct, recharge should balance pumpage through 2000 for the entire basin.

Effect of Pumpage at Pullman on Interpretations of Model Aquifer Studies of the Moscow Sub-basin

In our model aquifer studies, we compared model aquifer drawdowns with the actual drawdown in the real aquifers. We assumed that the actual drawdown in the real aquifer was solely the result of the pumpage at the Moscow center. Possibly, the actual drawdown in the real aquifer includes a component that is the result of pumpage at the Pullman center. If we could prove this hypothesis we could quantify the Pullman center component, and then remove it from the total drawdown at the Moscow center and the drawdown at Moscow would be less than the drawdowns used in our studies.

We interpreted the results of our model aquifer studies to indicate that drawdowns at Moscow are much less than would occur if aquifer constants, especially coefficient of storage, were in the "normal" range of values. We concluded that either aquifer constants are very large or a great deal of recharge has taken place. Removal of any Pullman center component of drawdown would mean that drawdowns are even less than were to be expected and leads to the conclusion that aquifer constants are even higher or that recharge is even greater than our Moscow sub-basin studies indicate.

A Model of the Moscow-Pullman Basin

A model of Moscow-Pullman basin can be designed that utilizes the same general principles that we used for the Moscow sub-basin study, but the model would be more complex. Two centers of pumpage would be necessary, one at Moscow and one at Pullman. The boundaries would be generalized as a wedge shape, rather than as a rectilinear shape. The north boundary would be the partially buried crystalline high that trends east-west from the Palouse Range through Albion; the south boundary would be the partially buried crystalline high that extends S70W through Paradise Ridge and Bald Butte to Granite Point. The Snake River cuts obliquely across part of the open edge of the wedge and probably constitutes a discharge boundary. The number of image wells and the problem of balancing them would be much more complex than our Moscow sub-basin model. A simple two-boundary, wedge-shaped aquifer model is discussed by Ferris and others (1962, p. 154).

Using data on pumpage and water levels at both Moscow and Pullman in the model would permit comparison of actual drawdowns with model drawdowns that result from various combinations of assumptions as to the properties of the aquifers. Assumptions that result in model drawdowns that match real drawdowns could then be evaluated as to the probability of their actually existing in the Moscow-Pullman basin.

CONCLUSIONS

General Statement

The ground-water supply in Moscow basin is adequate to meet the expected demands through the year 2000 and perhaps longer. The long-continued decline of water levels in the upper artesian zone is not necessarily the result of depletion of the entire ground-water supply. At least part of the decline is caused by continually increasing rates of pumpage in a small basin in which short-term barrier boundary effects are pronounced. Such drawdowns are normal in the hydrologic system of Moscow basin and are to be expected in the future in the wells in the middle and lower artesian zones without necessarily indicating that pumping is in excess of recharge.

The results of our model aquifer studies can be interpreted either to mean that no recharge takes place and the ground water in storage is sufficient to meet the needs of the basin until the year 2000 or to mean that considerable recharge does take place in the basin. The apparent increase in model coefficient of storage of the upper artesian zone during each of the successive periods of major increase of pumpage suggests that recharge was available to meet each increase in pumpage. The records of water levels in observation wells in the surficial aquifers show no long-term declines and constitute further evidence that recharge was sufficient to balance pumpage through 1965. Thus, two independent lines of evidence, one based on mathematical studies and the other on field data, indicate that recharge was sufficient to balance pumpage through 1965. Estimates of water available for recharge indicate that sufficient amounts will be available through the year 2000, both to the Moscow sub-basin and to the larger Moscow-Pullman basin.

The natural recharge can be augmented by artificial recharge through wells, utilizing seasonal surface-water runoff and effluents of waste-water treatment plants. Such a program would provide sufficient additional water to meet the needs of Moscow basin well into the 21st century.

The excessive amounts of iron in the waters of the upper artesian zone are caused by lateral migration of high-iron waters from iron-rich clay deposits along the margins of the recharge area of the artesian aquifers. Pumpage of the deeper artesian aquifers may someday induce high-iron waters to move downwards and increase the iron content of the deeper aquifers. Except for the excessive iron in parts of the basin, the ground waters contain no unusual or undesirable chemical constituents. The shallower waters are relatively rich in calcium which is replaced by sodium by base exchange processes as the waters move to the deeper aquifers.

Predicted Depths to Water and Estimates of Water in Storage

If we assume that no recharge takes place, then the best-fit model aquifer data for recent years of pumping are the best indication of the future properties of the Moscow basin aquifers. Under this assumption, the amount of water in usable storage in Moscow basin as of 1965, the amount of water that will remain in usable storage in the year 2000 after meeting anticipated demands, and the depths to water in the year 2000 are estimated to be:

Artesian Zone	In Storage 1965 (Billions of Gallons)	Consumption 1965-2000 (Billions of Gallons)	In Storage 2000 (Billions of Gallons)	Depth to Water 1965	Depth to Water 2000
Upper	9.2	?	?	120	?
Middle	86.7	10.5	76.2	250	305
Lower					
Sub-minimum	39.6	39.6	0	308	894
Optimum	262.8		233.2	308	391
Totals, Middle and Lower Zones only					
Sub-minimum	126.2	50:1	76.2		
Optimum	349.5		299.4		

If no recharge does take place and all water pumped comes from storage, the water in ground-water storage in Moscow basin is adequate to meet the needs of the basin past the year 2000. The 76 to 299 billion gallons that should remain in storage in the year 2000 should meet the needs of the basin at least until 2030 and perhaps until 2100.

These figures are based on the probably invalid assumption that no recharge takes place but are useful in managing a basin in which recharge does take place and is concealed as an unusually high value of coefficient of storage. If recharge is large, then the depths to water at any given time in the future will be less than those predicted in the no-recharge model. Should pumpage increase to an amount notably greater than the recharge, depths to water will become greater than those predicted in this particular no-recharge model.

RECOMMENDATIONS

The results of our studies convince us that the Moscow basin ground-water supply will meet the demand until at least the year 2000. Adequate time is available to make further studies and to evaluate the various alternatives before making a decision on the best way to meet the future water needs of the basin. This decision should be based on scientific, engineering, and economic analysis of data that are not yet available. We recommend a program consisting of:

1. Appointment of a Pullman-Moscow Hydrologist to supervise a continuing program of study and evaluation of the water resources of the basin;

2. Study of the engineering design and economics of artificial recharge and of recycling of waste water;

3. Further study of simple models based on existing data;

4. Implementation of a long-term program of basic data collection;

5. More sophisticated model aquifer studies of the basin as the basic data become available.

This program should continue for about 20 years and would permit a final decision as to the future sources of water for the basin around the year 1990, leaving at least 10 years to design and build the necessary physical plant should importation of water into the basin prove to be the only feasible alternative.

Pullman-Moscow Hydrologist

Study of the Pullman-Moscow basin lacks coordination. Over the years, different persons affiliated with various federal, state, and local agencies have studied unrelated aspects of the hydrology of the basin. Many of these studies have recommended continuation of collection of basic data relating to the aspect of the basin hydrology that was covered in that particular study. None of these data-collection programs has continued to any significant extent. The person making the recommendation either is not a resident of the area, or if he is a resident, he turns to other duties and does not have the time himself to continue the data collection program. As an example, none of our project observation wells has been measured since 1966; all water-level recording gages have been removed from wells. We simply have too many other things to do to take the time to continue the data collection program.

Continuity of data collection could be assured by appointment of a Pullman-Moscow Hydrologist whose duties would include:

1. Supervision of resumption and continuation of basic data collection;

2. Continuing evaluation of the usefulness of the various data being collected and interim interpretations of the significance of the data;

3. Supervision and coordination of design, implementation, and continuation of such new programs that he or other workers consider necessary to the continuing study of the basin;

4. Advising the cities and universities as to the interim status of the basin water supply and as to new studies that should be implemented.

We suggest that a half-time position be created to be housed at one of the universities or in one of the cities and that the cost of the position be shared by the four agencies. Costs would include salary, expenses, and fringe benefits of the position itself plus additional costs of new programs. Some of the new programs might receive support from other governmental agencies.

Artificial Recharge and Recycling Waste Water

The artificial recharge plan that we proposed should be cheaper than long-distance importation of water. However, we have only made very preliminary cost estimates. Tentative designs of several alternative physical plants should be worked out and the capital investment and operating costs of each should be analyzed. The concept of recycling effluent of the Moscow Waste Water Treatment Plant should be included in the alternatives under study. A similar program should be evaluated for Pullman which also has a wastewater treatment plant and a stream with seasonal runoff.

Additional Studies of Simple Models

In studying our Moscow basin aquifer model we were able to examine only a limited number of possible combinations of data because of the time consumed in doing computations with graphs and a hand calculator. Adapted to the computer, our model could be used to predict the drawdowns at the end of each year from 1965 to 2000 for several sets of assumptions as to the properties of the aquifers. Our Moscow basin aquifer model can be modified to include the effect of recharge. A series of studies should be run in which various percentages of the water pumped are assumed to come from recharge. Comparison of the drawdowns predicted from these studies with the actual drawdowns would lead to a better understanding of the hydrologic properties of the aquifers and of the actual recharge as well as providing a basis for determining the rate of drawdown that should be considered "normal" in the future operations in the basin,.

The model studies should be supported by a theoretical study of the problem of change of coefficient of storage with time and of the range of possible values of coefficient of storage in an artesian aquifer. A better theoretical knowledge of long-term coefficient of storage would permit better evaluation of the amount of recharge to Moscow basin.

Additional Basic Data Collection

Systematic basic data collection should be greatly expanded and the expanded program should continue at least until 1990. Nearly all studies to date have been short-term investigations, generally as part of a current research project of a faculty member or graduate student at one of the two universities. The duty of long-term basic data collection should be assumed jointly by the Cities of Moscow and Pullman and the Physical Plant Divisions of the University of Idaho and Washington State University. Assignment of personnel funding of the data collection program should become part of the regular operations of these organizations. Development of some special study techniques can be considered research for which funding from outside sources is possible, but once a technique has been developed, routine operation should be supported by local funds. Routine programs should involve collection of data on surface-water runoff, microclimate, water-level fluctuations in observation wells, and hydrologic properties of aquifers.

The number of stream-gaging stations should be increased. Surface-water studies should be expanded to include regular collection of data on suspended sediment and chemical quality of water. These data are particularly needed as a means of further evaluation of artificial recharge. Basic research also is needed on stream gaging methods in streams with small discharge and wide variation in flow. (See Chang-Lu, 1967, p. 69-70 for a discussion of problems in stream-gaging of Paradise Creek downstream of the City of Moscow Waste-Water Treatment Plant.)

The number of climatological stations should be increased. In particular, more precipitation and evaporation data are needed in the catchment area in the Palouse Range.

Systematic water-level measurements should be resumed in observation wells. Although data are being collected in some of the pumped wells in the basin, data are needed from observation wells in the three artesian zones and from wells in the surficial sediments and in the exposed crystalline basement. It may be necessary to drill special wells solely for observation purposes.

Pumping tests to determine aquifer constants have not proved satisfactory in Moscow basin because of the rapidity with which boundary effects influence water levels in the wells. Basic research should be done on boundary effects in small aquifers with large aquifer constants. The theoretical study should be followed by pumping tests in Moscow basin that should lead to better field data on the properties of the aquifers.

Studies of More Sophisticated Model Aquifers

Availability of a large amount of basic hydrologic data would justify use of aquifer models that are more sophisticated than those we used. According to R.E. Williams (oral communication), existing models that could be used if sufficient data were available include unsteady state (transient) finite difference or finite element models (Pender and Bredhoeft, 1969; Javandel and Witherspoon, 1969). A steady state, finite element model also is available that could be used to determine the steady state safe yield of the basin and also to evaluate the proposed artificial recharge program (Freeze and Witherspoon, 1966, 1967, and 1968; Freeze, 1969). Presumably, other sophisticated methods will be developed in the future that could be used in Moscow basin; however, almost any sophisticated method will require far better basic data than are now available for Moscow basin.

The 1970-1990 Program

Our studies indicate that the natural ground-water supply will meet the needs of Moscow-Pullman basin past the year 2000. If the basic data collection program and model studies that we recommend begin immediately, twenty years of observation and testing and twenty years of basic data will be available by 1990. A comprehensive review of the hydrology of the basin in the year 1990 would serve as the basis of a decision as to whether to continue to utilize the resources of the basin only or to import water from outside of the basin. If the decision is to import water, at least ten years will be available to design and build the system before the demands of the basin exceed the amount of recharge that is available as indicated by the present study.

If during the period 1970-1990, pumpage does exceed recharge, the first reliable indication will be excessive, long-term decline in water levels in observation wells, especially in the surficial aquifers, plus an increase in drawdown in production wells. It should be emphasized that these declines would be in excess of declines predicted from model aquifer studies. Declining water levels are to be expected in the artesian aquifers during the period 1970-1990 because of increased pumping rates and the inevitable increased effect of the barrier boundaries. As long as the decline is no greater than that predicted by the most reliable model of the basin, then no immediate danger of depletion exists. Should the actual decline exceed predicted decline consistently, then the model being used should be re-evaluated and the possibility of pumpage in excess or recharge should be reviewed. A program of continuous monitoring of water levels in unpumped observation wells and of comparing predicted and actual drawdowns of supply wells should furnish ample warning if pumpage begin to exceed recharge. Sufficient time would be available to design and build a water imporation system before the ground-water supply becomes fully depleted.

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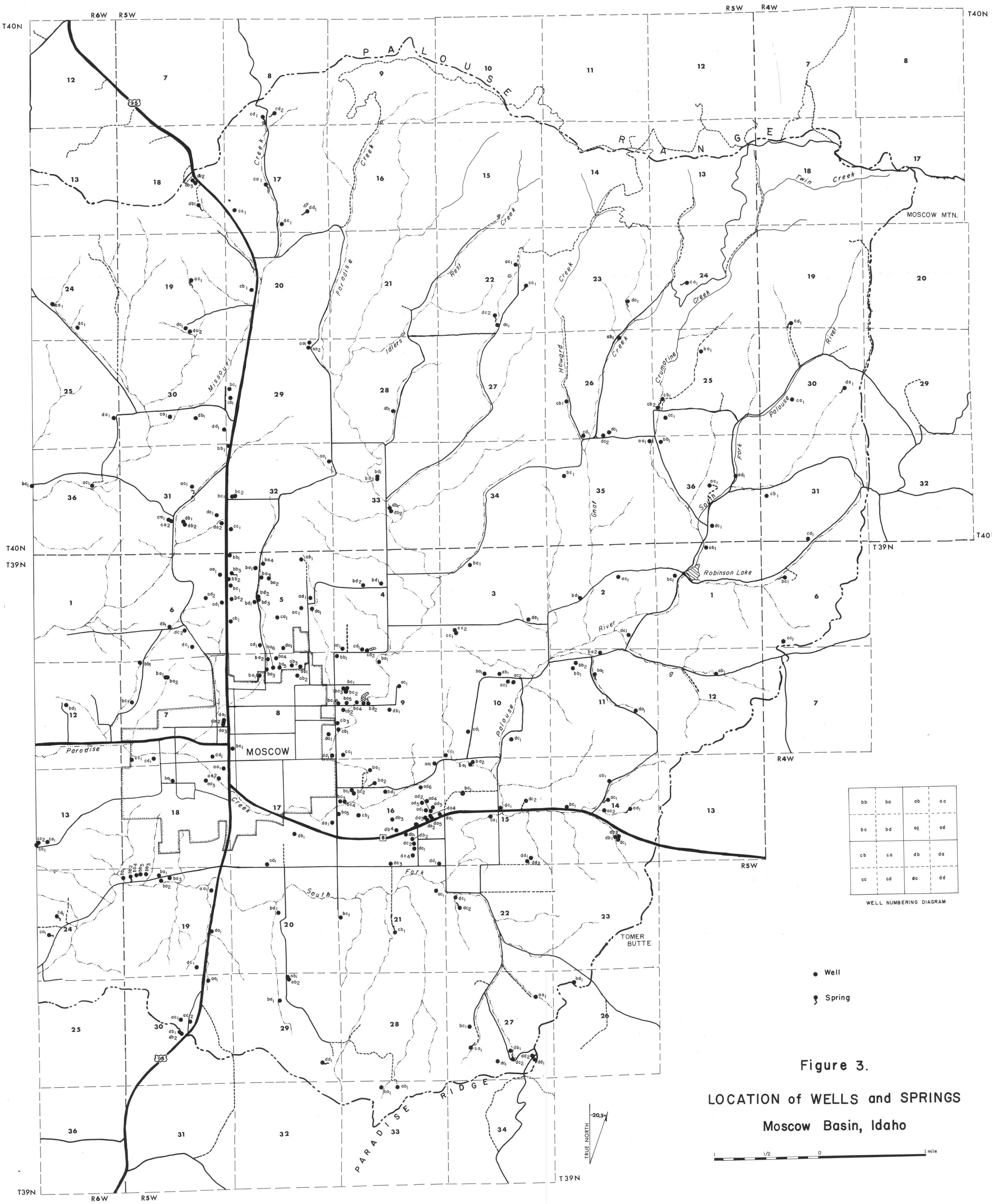
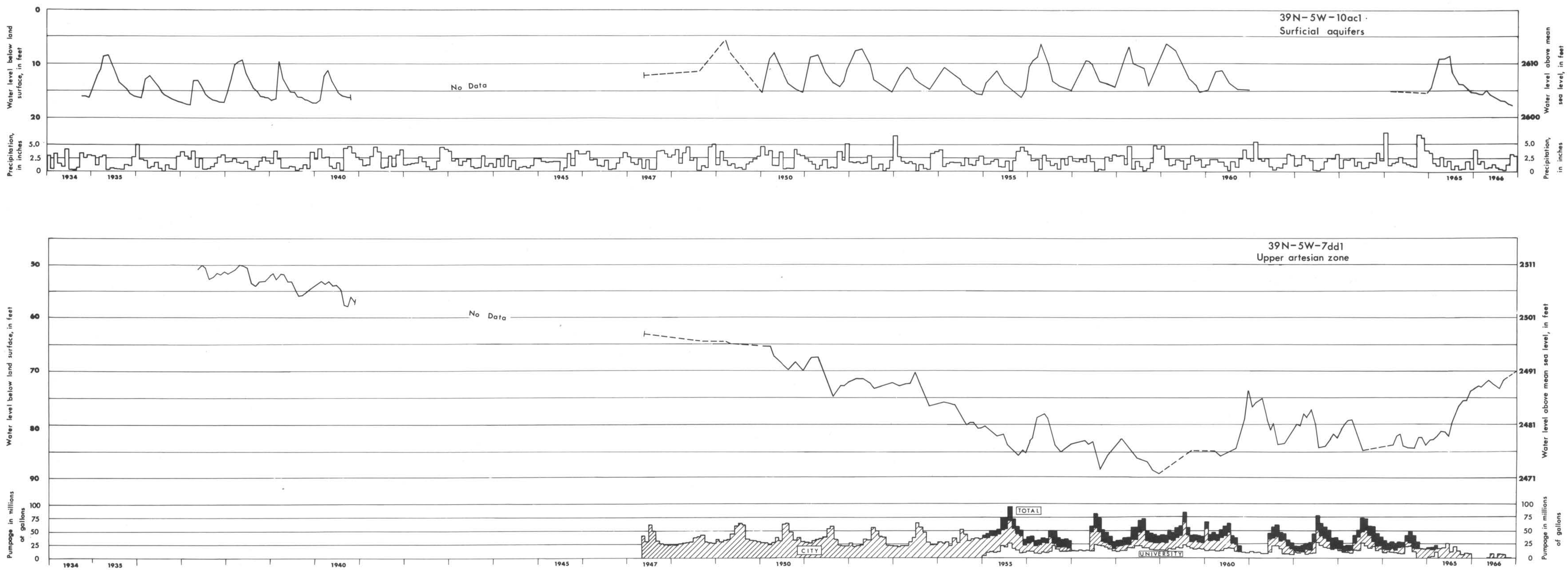


Figure 3.

LOCATION of WELLS and SPRINGS

Moscow Basin, Idaho

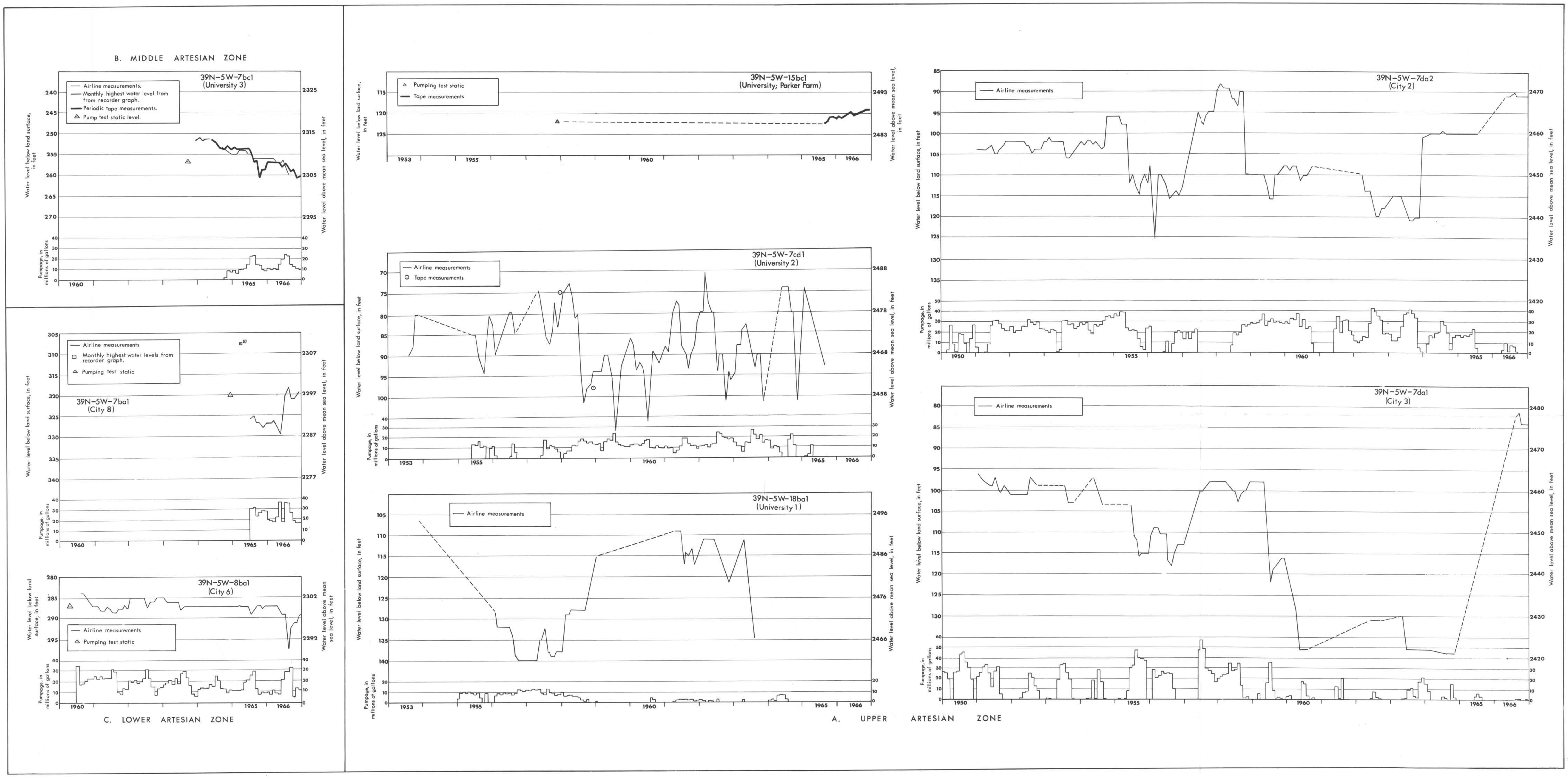
FIGURE 15.



WATER LEVEL FLUCTUATIONS IN TWO OBSERVATION WELLS WITH LONG PERIODS OF RECORD (1934-1966), MOSCOW BASIN, IDAHO

Precipitation from U.S. Weather Bureau records, (1934-1964), and College of Mines Meteorological Station, (1965-1966), Pumpage from records of the City of Moscow, (1947-1966), University of Idaho, (1955-1966)

FIGURE 17.



WATER LEVEL FLUCTUATIONS IN PUBLIC SUPPLY WELLS, CITY OF MOSCOW AND UNIVERSITY OF IDAHO, 1951-1966; Moscow Basin, Idaho