File Copy AW

15

IDO-22034-USGS - PT 3

ADMINISTRATIVE USE ONLY

UNITED STATES DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY
WATER RESOURCES DIVISION

B. Baca 3-22-85

GEOGRAPHY, GEOLOGY AND WATER RESOURCES OF THE NATIONAL REACTOR TESTING STATION, IDAHO

PART 3. HYDROLOGY AND WATER RESOURCES

By R. L. Nace, J. W. Stewart, W. C. Walton and others

Report not reviewed for conformance with
Geological Survey editorial standards
and usage of geologic names.

Prepared for the U. S. Atomic Energy Commission

Boise, Idaho
1959

Administrative Report
Not for Public Release

UNITED STATES DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY
WATER RESOURCES DIVISION

GEOGRAPHY, GEOLOGY AND WATER RESOURCES OF THE NATIONAL REACTOR TESTING STATION, IDAHO

PART 3. HYDROLOGY AND WATER RESOURCES

By R. L. Nace, J. W. Stewart, W. C. Walton and others

Report not reviewed for conformance with
Geological Survey editorial standards
and usage of geologic names.

Prepared for the U. S. Atomic Energy Commission

Boise, Idaho
1959

Administrative Report
Not for Public Release

PREFACE

This report embodies results of the work of many people in addition to the authors. They contributed much field and laboratory work, preliminary notes, suggestions and, most important of all, ideas. Also, they helped with the writing of earlier special reports which have been used as reference material. Special recognition is due to the following members of the Geological Survey: J. T. Barraclough, Morris Deutsch, F. E. Fennerty, K. H. Fowler, J. R. Jones, I. S. McQueen, Alan E. Peckham, Rex O. Smith, H. E. Skibitzke, C. V. Theis, W. I. Travis, and P. T. Voegeli. The authors are grateful to these geologists and engineers for their help. Some of them are co-authors of parts of this report, as indicated in the table of contents. All sections not otherwise credited, were written by the senior author.

CONTENTS

Contents—Continued

Page

Contents—Continued

Page

Contents--Continued | Page

Contents—Continued Page

Contents--Continued Page

ILLUSTRATIONS

PLATES
(in pocket)

FIGURES

Figures--Continued

Page

TABLES

GEOGRAPHY, GEOLOGY AND WATER RESOURCES OF THE NATIONAL REACTOR TESTING STATION, IDAHO

PART 3. HYDROLOGY AND WATER RESOURCES

By R. L. Nace, J. W. Stewart, W. C. Walton and others

INTRODUCTION

PURPOSE AND SCOPE OF REPORT

This report describes the hydrology and evaluates the water resources of the NRTS, sets these in their regional perspective, and forecasts conditions and effects during future years.

The ground-water yield of the Snake River Plain from the area east of the Hagerman Valley exceeds 5,000 cfs. This water is discharged through large springs in the valley of the Snake River, of which the Hagerman Valley is a part. Water wells on the NRTS tap the same body of ground water that feeds the springs. Pumpage from the wells is a very small percentage of the water yield of the plain, and the aquifers beneath the NRTS are capable of yielding many times the amount of water now pumped. Nevertheless, the water demand of the NRTS is bound to increase and the increase may be large. Therefore, it is desirable to know the potential perennial yield of aquifers beneath the NRTS. An estimate is derived in this report.

Yearly withdrawals of ground water are large in many parts of the Snake River Plain and in adjacent valleys, chiefly for irrigation. Pumpage for industrial and public water supplies and for stock and rural domestic use also is substantial. In the Hagerman Valley, water is diverted directly from springs to generate hydroelectric power. Use of ground-water on the Snake River Plain has increased markedly since 1946, and by 1965 developments on the plain will require pumpage of about 25 pct of the perennial ground-water yield of the area. Consumptive depletion, chiefly by irrigated crops, may be on the order of 15 pct of the total supply. If the water requirement of the NRTS increases greatly, the Station will compete directly with other water users within 20 years, if not sooner. In some parts of the plain, water developments already have created local conflicts of interest.

These few facts about the water outlook are sufficient reason for the interest of the Commission in the quantity, occurrence, source, movement, and chemical quality of the ground water. The regional hydrologic relations of the water beneath the NRTS, the mutual effects of water use by the Commission and by others, and the ultimate or stabilized water regimen that may be expected on the Snake River Plain are among the more important factors in the water situation. The public-health hazard that may be created or threatened by routine waste disposal or by an accident with radioactive materials is a problem of related importance. The hazard would apply to a large segment of the Snake River Plain because the ground-water body is essentially plains-wide. The problem, therefore, is regional, and certain sections of this report deal with the large plains area bounded approximately by the Snake River on the south and west from Blackfoot to Bliss, by the Mud Lake basin on the east, and by the mountain complex of central Idaho on the north (fig. 1; see also pl. 2 of pt. 2). The land area so circumscribed comprises about 8,000 sq mi.

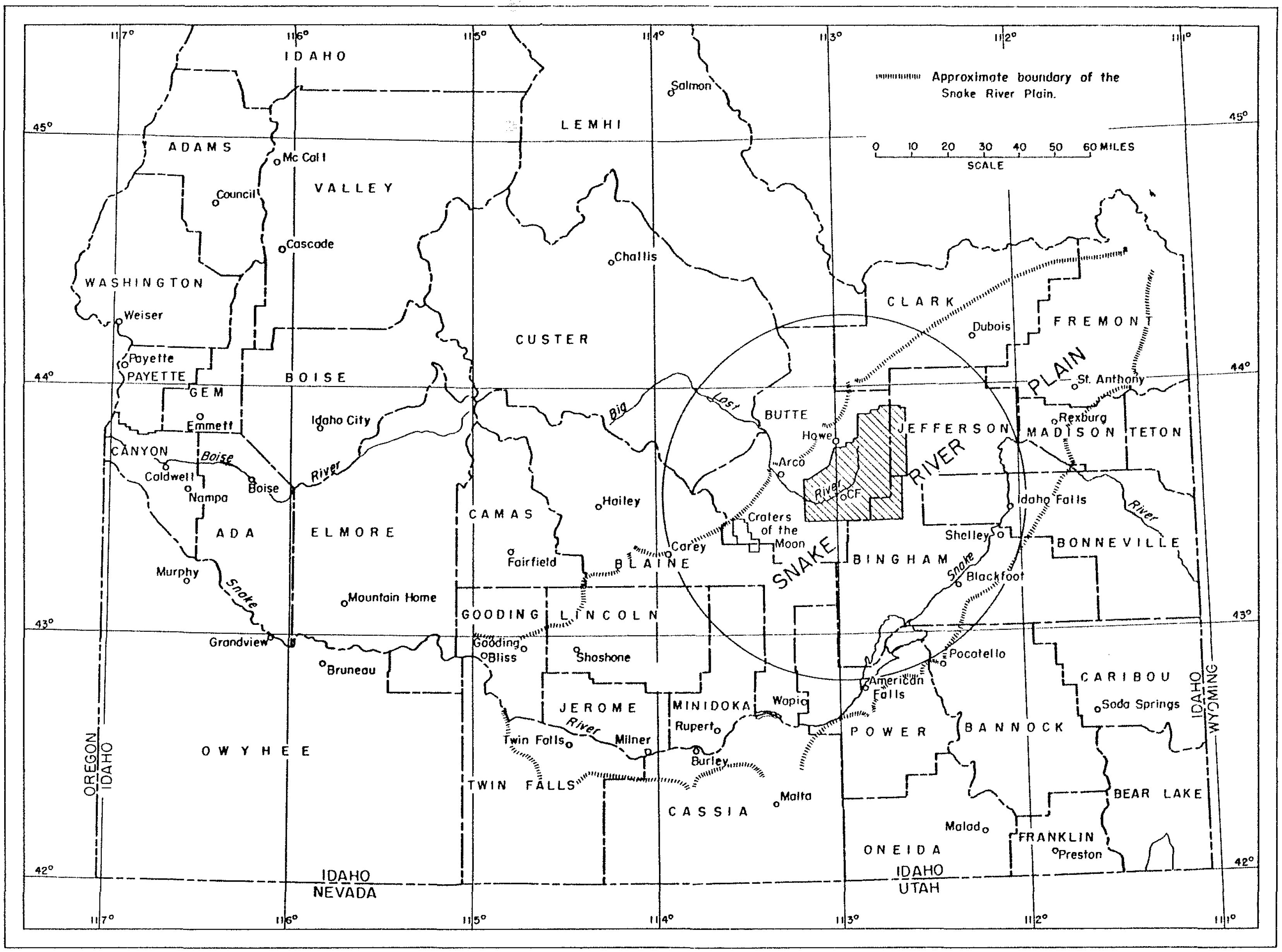


Figure 1.—Index map of southern Idaho showing location of National Reactor Testing Station on the Snake River Plain. (Circle, radius 50 miles, centers on CF area.)

GPO 988997

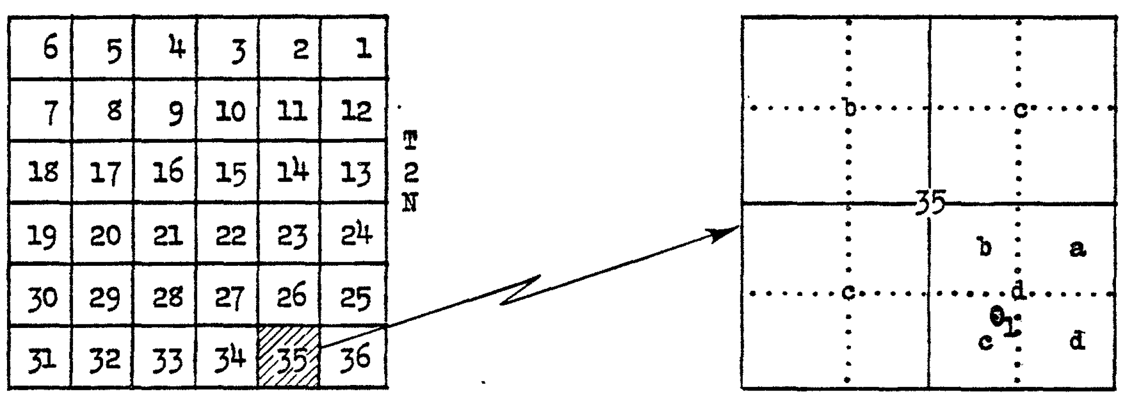
FIELD AND LABORATORY WORK

The ground-water geology and hydrology of the NRTS was studied concurrently with the geologic work described in part 2. Geologic mapping and study of basalt outcrops helped to delineate the water-bearing properties of the basalt. Test drilling aided the delineation, yielded information about the occurrence of ground water and the position and configuration of the water table, and provided a network of observation wells.

Study and description of the general ground-water geology of the NRTS was assisted by Morris Deutsch, J. R. Jones, R. O. Smith, P. T. Voegeli, and S. W. West. R. C. Carson, Eugene Shuter, and H. G. Sisco made most of the water-level measurements. J. T. Barraclough studied the discharge of the Big Lost River and made most of the direct measurements of discharge. A recording gage near old Pioneer station is operated by the Geological Survey, and annual reports on the stream-gaging have been prepared by Wayne I. Travis. Laboratory work in Idaho was chiefly by I. S. McQueen and J. W. Stewart, assisted by other members of the Geological Survey. Work at the Denver Hydrologic Laboratory was supervised by A. I. Johnson, engineer in charge. Geochemical and geophysical field experiments on ground-water behavior were by H. E. Skibitzke, assisted by A. E. Robinson, both from Phoenix, Arizona. Chemical analyses of water were made in Geological Survey laboratories at Salt Lake City, Utah, and Portland, Oreg. Radiometric analyses of water samples were made in the Washington, D. C. laboratories of the Geological Survey.

WELL-NUMBERING SYSTEM

The well-numbering system used in Idaho indicates the locations of wells within the official rectangular subdivisions of the public lands, with reference to the Boise baseline and meridian. The first two segments of a number designate the township and range. The third segment gives the section number and is followed by two letters and a numeral, which indicate the quarter-section, the 40-acre tract, and the serial number of the well within the tract. Quarter sections are lettered a, b, c and d in counterclockwise order, from the northeast quarter of each section (see diagram). Within the quarter-sections 40-acre tracts are lettered in the same manner. Well 2N-31E-35dc1 is in the SW$\frac{1}{4}$SE$\frac{1}{4}$ sec. 35, T. 2 N., R. 3 E., and is the well first visited in that tract.



R. 31 E.

2N-31E-35dc1

THE HYDROLOGIC CYCLE IN THE SNAKE RIVER PLAIN

Hydrology, the science dealing with water, is concerned chiefly with water in its course from the place where it falls on the land to where it reaches the sea or is returned to the atmosphere. The hydrologic cycle consists of a general circulation of water from the sea into the atmosphere, onto the land, and back to the sea. The major cycle contains many subcycles. Geologic and physiographic factors are two of the principal natural environmental factors that affect local and regional subcycles.

NATURE OF THE CYCLE

Some precipitation is intercepted by vegetation and other obstacles and is re-evaporated directly; some evaporates from the land surface and from lakes and ponds; some runs off at the surface; and some become soil moisture and is transpired by plants. Ground-water recharge occurs only where and when the supply of water exceeds the amount disposed of by these processes, which have, in effect, a prior call. Water in the zone of saturation does not remain there forever but migrates inexorably through various environments back to the sea.

The water supply on the Snake River Plain is perennially dependable only because it is replenished, directly and indirectly, by perennial precipitation. Precipitation on the plain and its tributary areas is derived largely from moist air masses that move in from the Pacific and Arctic regions. Mountain barriers between the plain and those regions wring most of the water from the air masses before they reach the plain, which is in their rain shadow. Water that is not consumed on the plain leaves the area through the Snake River and by ground-water underflow.

The Snake River is one of the most thoroughly regulated streams in North America, but additional regulating works are under construction, are authorized, or are planned. The surface-water supply upstream from Bliss probably will be in full use within 20 years or less.

The Snake River Plain is the gathering ground for surface water and ground water that originate within an area much larger than the plain. The ground water, most of which is in the Snake River basalt aquifer, is the only abundant and thoroughly dependable water supply that is directly available to the NRTS. Some surface water runs into the Station and contributes soil moisture and ground-water recharge by seeping into the ground, but there is no "through-going" surface drainage. Relatively little ground-water recharge occurs from precipitation directly on the plain, and underflow from more distant sources is the chief means of replenishment.

PRECIPITATION AND EVAPORATION

Precipitation ranges between 6 and 8 in. on the central part of the Snake River Plain and between 10 and 12 in. in some border areas (see part 2, p. 26). Precipitation is greater in the east than in the west, being about 16.2 in. at Ashton and 8.8 in. at Bliss. Precipitation in much of the area barely suffices to supply the soil moisture needed for survival by the sparse growth of "desert" shrubs and grasses. Most of the water precipitated on the plain is more or less continuously evaporated and transpired from the land where it falls.

Parts of the bordering intermountain valleys receive somewhat more precipitation than does the plain, but some valleys are nearly as dry as the plain. For example, precipitation is 15.3 in. at Hailey and 14.1 in. at Hill City, but it is only about 7.3 in. at Challis and 9.3 in. at Mackay Ranger Station. On the higher mountains in areas tributary to the Snake River Plain, precipitation falls largely as snow during September through April and the water equivalent probably ranges up to 35 in. or more. The seasonal pattern of precipitation on the plain differs from that in the mountains, midsummer being the time of heaviest precipitation. Summer is also the time of higher temperature and lower relative humidity, which cause relatively rapid evaporation and transpiration.

RUNOFF AND INFILTRATION

Certain regional peculiarities about runoff are decisive factors in the hydrology of the Snake River Plain, especially the ground-water hydrology. Precipitation directly on the plain yields practically no exterior runoff, and nearly all the water evaporates or enters the ground. Precipitation generates considerable runoff in mountains north of the plain and some of the water later sinks into the ground and reaches the plain by underflow. Water that reaches the plain at the surface sinks into the ground around the edges of the plain. Thus, unconsumed surface water from much of the tributary area is a source of replenishment for ground water beneath the plain.

The Snake River, the mainstem surface drainageway of the plain, derives some millions of acre-feet of runoff yearly from the mountains of eastern Idaho and northwestern Wyoming. In recent years, however, owing to regulation and use of the river water, the average discharge past Milner Dam (the irrigation-diversion structure farthest downstream in Idaho) was only about 1.5 million acre-feet yearly, and in a few of the years no water spilled past Milner Dam. The water diverted for irrigation is used and reused one to several times on the plain. The net consumptive use by evapotranspiration on crop lands was not calculated, but a large unconsumed residual volume of the diverted surface water enters the ground and recharges the ground water.

The water-bearing, permeable Snake River basalt rests on nonpermeable older rocks which crop out along the east wall of the valley of the Snake River between Milner and Bliss at many places above the level of the river. Ground water is discharged at the surface in springs along the top of the basement rock, which forms a natural dam over which the water spills. The yield of the springs is an approximate measure of the unused ground-water yield of the basin to the east--currently about 4,000,000 ac-ft a year, which is about 30 pct greater than the estimated yield before irrigation was begun. Thus, about 30 pct (1,200,000 ac-ft) of the water now discharged by the springs probably is unconsumed irrigation water. In the natural water regimen, all surface water yielded by valleys tributary to the north side of the plain formerly reached the plain and became ground water. Now, however, much of the tributary surface water is consumptively used by irrigated crops. Ground-water recharge from those sources decreased concurrently with the increase from Snake River water. The regimen is complicated also by large withdrawals of ground water from wells east of the springs. For these reasons the aggregate discharge of the springs does not represent the perennial gross ground-water yield of the basin nor the gross amount of ground-water recharge. The discharge is a measure only of the unused ground-water supply.

SURFACE WATER

Very little water enters the Station by surface inflow, and no water leaves the NRTS by surface outflow except for minor local slope runoff. The named drainageways into the NRTS are the Big Lost River, the Little Lost River, and Birch Creek. Ephemeral drainage from foothills to the north occasionally reaches the Station. Only the Big Lost River regularly delivers water within the Station boundaries.

BIG LOST RIVER

The recording-gage station on the Big Lost River below Arco is about 4 mi southeast of the town in the NW$\frac{1}{4}$SW$\frac{1}{4}$ sec. 17, T. 3 N., R. 27 E. The station designation is "Big Lost River near Arco, Idaho". Special study of river discharge was made at intervals from late in 1951 through the early part of 1953. Several series of direct measurements were made at the gaging station and downstream at ten temporary measuring stations along a 40-mile reach of the river channel. Their locations were shown on the geologic map (pl. 1, pt. 2.)

Regimen of the River

The reach of the river in the NRTS flows intermittently. Within historic time, the river has discharged a considerable volume of water onto the plain, but for many years upstream storage and diversion of water for irrigation have drastically curtailed runoff to the plain. Mackay Reservoir regulates the river above Arco, and irrigation diversions above and below the reservoir deplete the flow.

"Old-timer" inhabitants of the area report that during seasons of high runoff in former times the river commonly reached the first and second Big Lost River playas east of Howe. Few people have seen ponded water in the third and fourth playas. The late George Walker,1/ a resident of the area from 1882 to 1951, saw "the waters of Birch Creek and the Big Lost River mingle" several times in the fourth playa, but the Big Lost River probably has not reached the fourth playa since the turn of the century. Water has reached the first playa in substantial amount in many years—1921, 1922, 1923, 1927, 1938, 1943, 1944, 1947, and perhaps others2/. In 1943 Mr. Crandall observed water extending along the outlet channel from playa 1 toward the second playa, but he believed that no water had reached the second playa since 1918. The first and second playas contained ponded water during part of 1952. Water reached the first playa again in 1953.

The four playas in the northern part of the NRTS are connected by shallow channels. Playas 1 and 2 are in the northeastern part of T. 5 N., R. 30 E.; playa 3 is in the south-central part of T. 6 N., R. 31 E., and playa 4 is in the northeastern part of T. 6 N., R. 31 E. (pl. 1). The

1/ Oral communication, 1951. According to Lynn Crandall, Snake River Watermaster at Idaho Falls, Idaho (written communication, May 1, 1952) it is said that Birch Creek and the Big Lost River mingled in the "lower sinks" in 1894.

2/ Lynn Crandall, written communication, May 1, 1952.

playas were flooded repeatedly during recent geologic time, and they contained ponded water during periods sufficiently long that distinctive shoreline and offshore physiographic features developed. Climatic change, probably including reduced precipitation, may have reduced runoff during relatively recent geologic times.

Nowadays, the small amount of water discharged past the gaging station near Arco during most of the year is dissipated by evaporation and seepage within a few miles below the station. The channel ordinarily is dry below the old diversion dam in the NRTS in sec. 5, T. 2 N., R. 29 E. During periods of raised runoff, water sometimes reaches the crossing of Highway 20 in sec. 33, T. 3 N., R. 29 E. At still higher runoff stages, the river extends several miles farther downstream, as it did late in 1951, 1952, and 1953.

Discharge at Station Below Arco

The records of daily discharge past the gaging station are published annually[1/]. High wind, channel moss, and ice were minor disturbing factors at times during the period of record. In general, however, the daily records probably conform to usual standards of accuracy — that is, within 3 to 5 pct. Table 1 shows the maximum, minimum and mean yearly discharge at the station during 9 water years[2/] of record. The mean

1/ U. S. Geol. Survey Water-Supply Papers 1093 (1947), 1123 (1948), 1153 (1949), 1183 (1950), 1217 (1951), 1247 (1952), 1287 (1953), 1347 (1954), and 1397 (1955).

2/ By convention, any water year begins on October 1 of the preceding year. The 1947 water year began on Oct. 1, 1946.

discharge during the 9-year period was 58.4 cfs. Hydrographs for each year of record, tabulation of mean daily and monthly discharge, and records of miscellaneous measurements of discharge at the temporary measuring stations below the gage are contained in figures 1-9, and tables 1 and 2 of Appendix 2. The mean annual and monthly discharge of the river during the period of record is represented graphically in figure 2.

Table 1.--Maximum and minimum instantaneous and yearly mean discharge of Big Lost River below Arco, 1947-55

	Maximum		Minimum		Mean yearly	
Year	Date	Cfs	Date	Cfs	Cfs	Acre-feet
1947	6- 6	285	8-12, 13, 19, 20	28	83.3	60,260
1948	6-15	171	5-20, 21	3.0	37.0	26,870
1949	6- 3	237	5- 7	9.9	32.6	23,620
1950	4- 3	102	5-22	5.1	25.0	18,127
1951	8- 6	272	-	a/	39.6	28,649
1952	6-11	698	8-25	36	171	124,310
1953	6-17	251	5-14, 16, 18	16	80.8	58,470
1954	1-23	b/ 85	5- 9, 10	6.0	39.4	28,530
1955	10-31	b/ 52	6- 2	c/ .5	16.9	6,178
1947-55	6-11-52	698	6- 2-55	2.7	58.4	41,669

a/ Not determined. Estimated minimum daily flow of 10 cfs Jan. 28 to Feb. 2.
b/ Maximum mean daily discharge.
c/ Minimum mean daily discharge.

GPO 991857

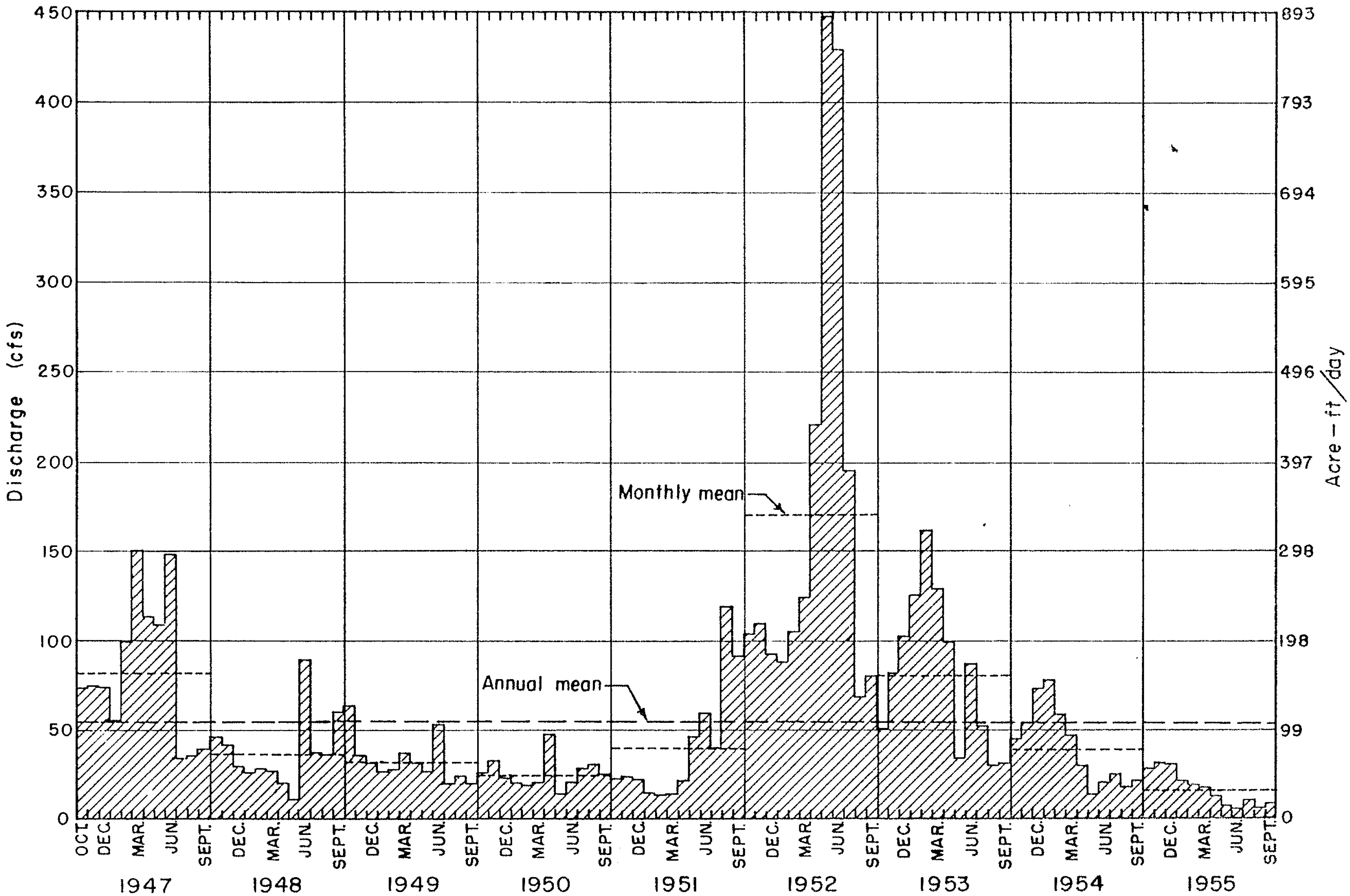


Figure 2.—Annual and monthly mean discharge of Big Lost River below Arco, water years 1947-1955.

Discharge in 1951

Precipitation in July and August 1951 and 1952 exceeded the normal amount in much of southeastern Idaho (table 2). During those months in 1951, precipitation occurred chiefly during a two-week period. About two inches of rain fell on August 3 in less than an hour during a cloudburst on hills near Arco. The main irrigation-diversion canal below Mackay Dam was overloaded and breached after the cloudburst; several lateral canals and ditches failed, and the canal headgate had to be closed. Water from the breached canals, undiverted water in the Big Lost River, and flash runoff flooded thousands of acres, including the town of Arco. At the gaging station below Arco, the river rose sharply on August 3 (fig. 3), reached a peak on August 6, and returned to normal on August 19. The storage level in Mackay Reservoir, about 20 mi. above Arco, was high, owing to unusually high runoff late in July and early in August. Because of disrupted diversion works upstream, an unusually large amount of water was released from Mackay Dam, and the discharge below Arco reached a secondary peak about August 28. After September the discharge rate fluctuated in a gradually rising trend until early November, after which it leveled off until early in the spring of 1952. During several months thereafter the discharge rates and fluctuations were high.

Table 2.—Departure from normal precipitation at stations in southeastern Idaho, July and August 1951 and 1952

[Inches of water. From published records of the U. S. Weather Bureau]

(T, total; D, departure from normal; N, normal)

	1951						1952					
	July			August			July			August		
	T	N	D	T	N	D	T	N	D	T	N	D
Aberdeen Experiment Station	0.53	0.50	0.03	1.10	0.42	0.68	0.08	0.50	-0.42	1.23	0.42	0.81
Arco	-	-	-	-	-	-	1.24	.55	.69	.28	.60	-.32
Blackfoot	.25	.66	-.41	1.11	.64	.47	.28	.66	-.38	.29	.64	-.35
Chilly-Barton Flat	2.27	.60	1.67	3.17	.95	2.22	1.30	.60	.70	.45	.95	-.50
Grouse	1.24	.77	.47	.86	.82	.04	.79	.77	.02	1.67	.82	.85
Hamer	1.84	.47	1.37	4.22	.80	3.42	1.17	.47	.70	.51	.80	-.29
Idaho Falls	.22	.62	-.40	2.43	.59	1.84	.07	.62	-.55	.53	.59	-.06
Mackay RS	2.31	.85	1.46	2.71	.78	1.93	1.83	.85	.98	1.69	.78	.91
Reactor Testing Station	-	-	-	-	-	-	.38	-	-	.29	-	-

GPO 991857

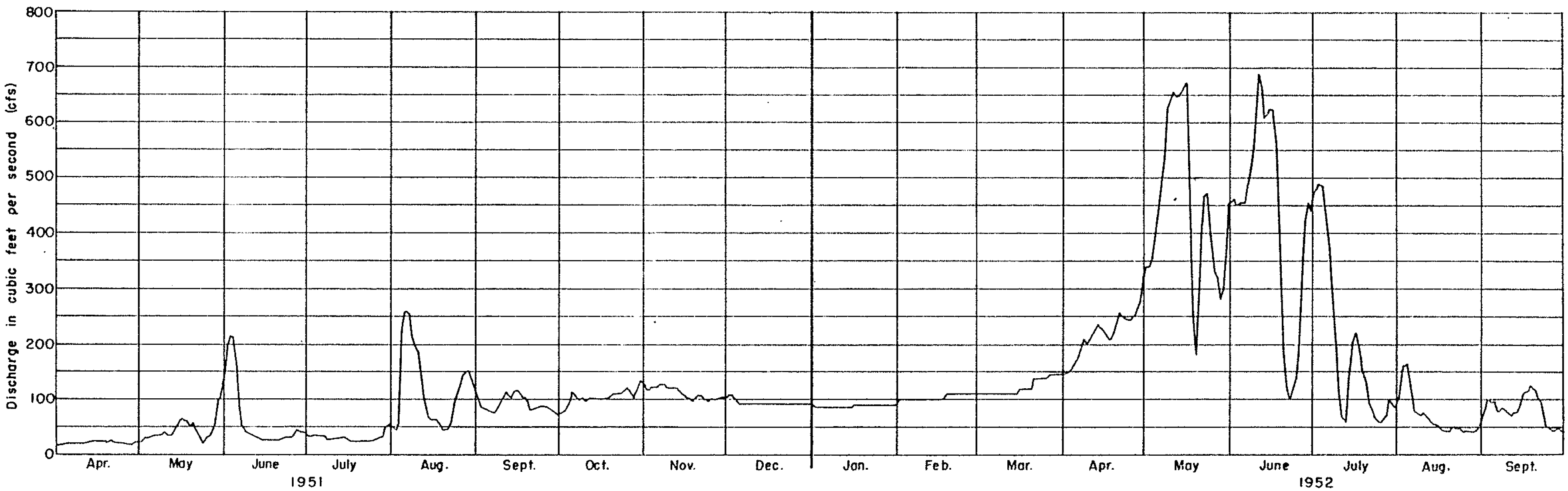


Figure 3.—Mean daily discharge of Big Lost River near Arco, Idaho, April 1951 to September 1952.

By the time repairs were made to breached canals in the Big Lost River Valley, the main irrigation season was past, the storage level was still high in Mackay Reservoir, and there was no need to divert or store much water for irrigation. The river below Arco continued to discharge more than the normal amount of water during the rest of 1951 and the first half of 1952, and water extended far down the channel through the NRTS.

Discharge in 1952

The discharge of the river below Arco was above normal during most of 1952, the year having the highest mean daily discharge rate in the 9 years of record. Before 1952 the maximum discharge was 285 cfs, on June 6, 1947. That amount was exceeded on 58 days in 1952 and a new maximum of 681 cfs occurred on June 11. The high runoff was derived from the melting of unusually heavy snow that accumulated on the entire watershed from December 1951 through February 1952.

Percolation from the River

The relation of the Big Lost River to the water table in the vicinity of Arco and upstream from there changes from time to time and place to place, but the river gains in aggregate flow from effluent ground water. Downstream from Arco the water table drops some hundreds of feet and the river is perched far above it. Practically nothing is known about percolation losses between Arco and the gaging station below Arco. Below the gage, losses were measured while the discharge was high in 1951-52. The losses measured indicate the magnitude of infiltration rates in materials like those in the channel of the Big Lost River, and information about the rates may be useful in planning the disposition of liquid waste in the ground.

From 1951 to 1953 nine sets of discharge measurements were made along the main channel of the Big Lost River, and inflow and outflow in the playa areas were measured. Water was ponded in three of the four playas in 1952, and the seepage losses in the playas were computed. Throughout the gravelly part of the flood plain, the river bed is permeable and the loss of water by infiltration is proportionately large. The flowing reach of the river extends far into the NRTS only when the discharge exceeds the seepage rate through the channel deposits. Also, the river is able to extend its flowing reach, when runoff persists long enough for seepage water to saturate the sediments beneath the channel. On August 2, 1951 water extended along the channel to a point about midway between the abandoned diversion dam on the NRTS and the bridge on Highway 20. Following the cloudburst and canal failures of August 3, water extended a short distance below the highway bridge on August 4; beyond the Lincoln Boulevard bridge in the NE$\frac{1}{4}$ sec. 24, T. 3 N., R. 29 E. on August 6; and to a ford on the West Monument Road in the SW$\frac{1}{4}$ sec. 3, T. 4 N., R. 30 E. on August 9. A trickle of water flowed into playa no. 1 on August 31.

Channel Losses

The surface areas of measured reaches in the river channel (table 3) were computed from stadia measurements of stream-channel cross sections at half-mile intervals. Linear channel distances were scaled from aerial photographs. The rates of infiltration into the ground (table 4 and 5) were computed from discharge measurements and area computatións. No correction was made for evaporation from the water surface because, in the measured short narrow segments of the stream channel, evaporation was negligible in proportion to other sources of error in the data. A correction for evaporation loss in the playa area was made because the loss was appreciable there.

Table 3.—Dimensions and areas of measured reaches of the Big Lost River

Station	River miles from reference point a/	Length		Average width (ft)	b/ Area (sq ft)
		Miles	Feet		
1	0.8	-	-	-	-
2	14.3	13.5	71,280	48.0	3,420,000
3	19.7	5.4	28,510	39.4	1,112,000
4	22.0	2.3	12,140	36.4	442,000
5	26.7	4.7	24,820	33.0	820,000
6	30.9	4.2	22,180	23.1	512,000
7	34.5	3.6	19,000	35.6	676,000
8	40.2	5.7	30,100	38.0	1,140,000
9	42.6	2.4	12,670	29.9	374,000
1-9		41.8	220,700	-	8,496,000

a/ Reference point chosen arbitrarily, 0.8 mi above gaging station.

b/ Horizontal area of water surface; not the slope area of the channel.

Table 4.—Average infiltration rates in Big Lost River below station 1, 1951-52

Date	Seepage rate: cfs	Seepage rate: cfd (thousands)	Seepage rate: cfd/ft^2	Seepage rate: gpd/ft^2
Aug. 16-17, 1951	63.9	5,521	0.677	5.1
Aug. 30-31, 1951	70.3	6,070	.746	5.6
Sept. 13-14, 1951	96.3	8,320	.980	7.4
Nov. 2-3, 1951	65.8	5,690	.697	5.2
Apr. 30-May 1, 1952	111	9,600	1.18	8.8
May 22-24, 1952	130	11,200	1.38	10.3
June 30, 1952	90	7,780	.955	7.2
Sept. 15-17, 1952	50.4	4,350	.534	4.0
Apr. 8-10, 1953	22.8	1,970	.242	1.8
Average			0.820	6.2

Table 5.--Average infiltration rates in measured reaches of Big Lost River, 1951-53

Reach	Length		Wetted area (thousands of sq ft)	Average seepage rate		a/ Total seepage (thousands of gpd)
	Miles	Feet		cfd/ft²	gpd/ft²	
Sta. 1-2	13.5	71,280	3,420	0.414	3.1	10,580
Sta. 2-3	5.4	28,510	1,112	.710	5.3	5,920
Sta. 3-4	2.3	12,140	442	1.20	9.0	3,960
Sta. 4-5	4.7	24,820	820	1.24	9.3	7,600
Sta. 5-6	4.2	22,180	512	2.52	18.9	9,630
Sta. 6-7	3.6	19,000	676	.289	2.2	1,460
Sta. 7-8	5.7	30,100	1,140	.984	7.4	8,380
Sta. 8-9	2.4	12,670	374	1.11	8.3	3,095
Sta. 2-6	16.6	87,650	2,886	1.15	8.6	24,800
Sta. 6-9	11.7	61,770	2,190	.966	7.2	15,600
Sta. 1-8	39.5	208,500	8,147	.820	6.1	49,800
Sta. 1-9	41.8	220,700	8,504	.691	5.2	44,100

a/ The accumulative totals of lines 1 to 8 do not agree with the compound totals in lines 9 to 12 because no measurements were simultaneous and the compound totals are from averages on days when the stage of the river differed from that on days when single segments were measured. Nevertheless the totals correctly show the general magnitude of seepage losses.

The rates of infiltration along the river channel were quite uniform until the time of the measurements on April 6-10, 1953, when the discharge rate was moderate. At that time the infiltration rate was much smaller than it was during earlier periods of moderate discharge. The change in the infiltration rates occurred after the river had maintained a rather steady flow at medium stage throughout the winter of 1952-53, following the higher stages in the summer of 1952. Silting of the channel may have reduced its permeability, and prolonged seepage may have saturated a considerable volume of gravel beneath the river bed and decreased the subsurface hydraulic gradient away from the river. The most permeable reach for which the infiltration rate was determined was that between stations 5 and 6, where the rate averaged 2.52 cfd/ft^2 (18.9 gpd/ft^2) (see table 5). A general direct relationship between river stage and rate of infiltration is shown by a plot of the data in table 4 (fig. 4). The maximum observed aggregate rate in the entire reach between stations 1 and 9 was 10.3 gpd/ft^2 on May 22-24, 1952, when the discharge at the gaging station was 130 cfs. The minimum was 1.8 gpd/ft^2 on April 8-10, 1953, when the discharge was 22.8 cfs.

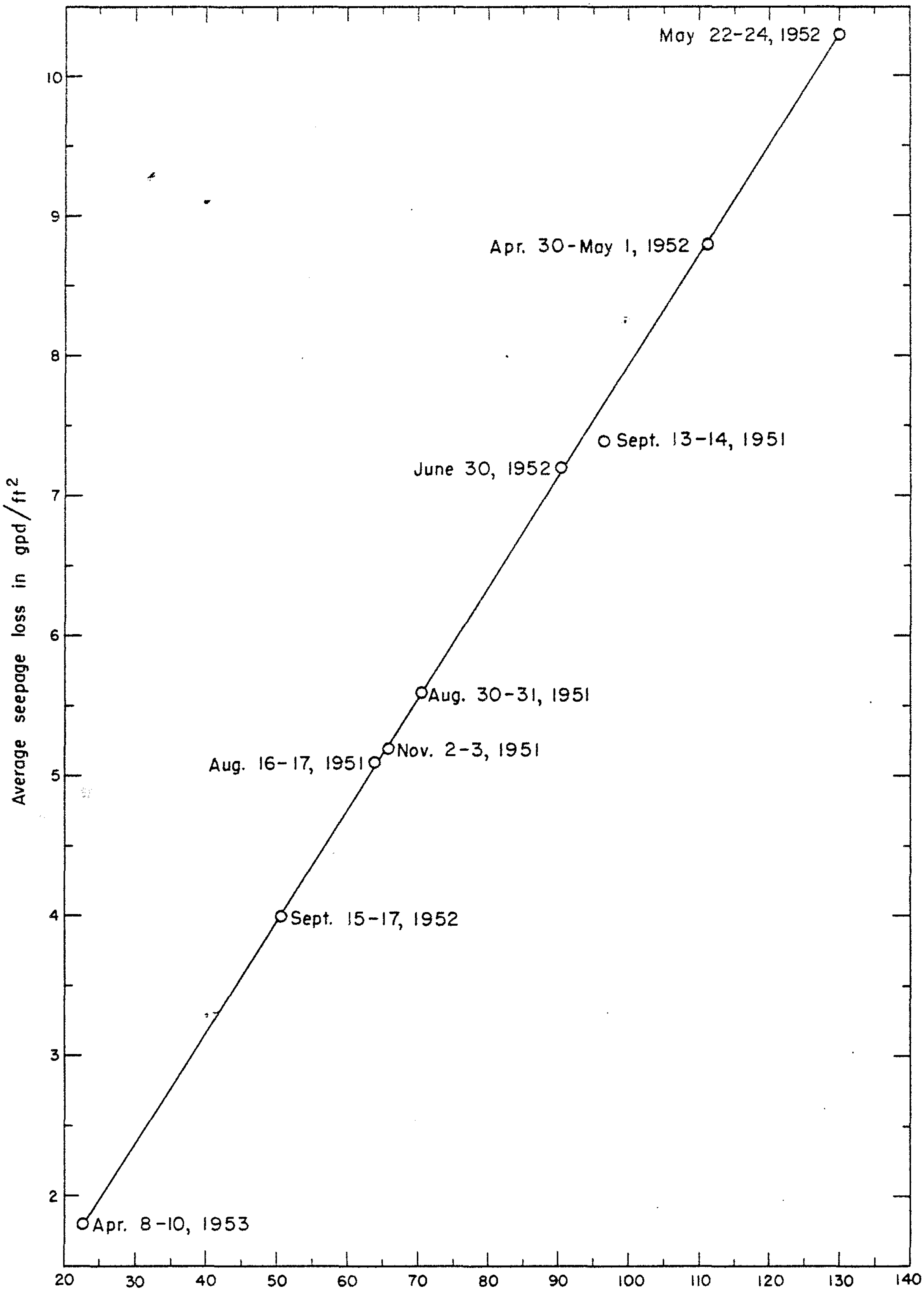


Figure 4.— Relation of discharge to average rate of infiltration between stations 1 and 9, Big Lost River.

The average depth of water over the infiltration area in the channel, approximated from inspection of plotted cross-sections, was about 1.3 ft. Upstream from station 2 the channel of the river is cut chiefly in basalt; between stations 2 and 6 it is chiefly in gravel; downstream from station 6 the bed is sand and silt. A correlative change is apparent in the grade of the channel (fig. 5). The average grade is 1.5 pct above station 2, 2.1 pct, between stations 2 and 7, and 1.1 pct below station 7. The steeper slopes induce higher water velocity and greater load-carrying capacity. The average rate of infiltration between stations 2 and 6 was much higher than the rate below station 6, and of course higher than the average rate in the entire segment, stations 1 to 9 (table 5).

Summary

The average rates of infiltration in single reaches ranged from 0.289 to 2.52 cfd/ft^2 (2.2 to 18.9 gpd/ft^2). During the different periods of measurement, the average infiltration rate in the whole 40-mile gaged segment varied from 0.242 to 1.38 cfd/ft^2 (1.8 to 10.3 gpd/ft^2). The average of the averages for the nine series of measurements was 0.82 cfd/ft^2 (6.2 gpd/ft^2), which is more representative of the average rate of infiltration during the period.

Increasing discharge caused a proportional increase in the total volume of seepage and in the average rate of infiltration per unit area of river reach, because both the hydraulic head over the infiltration area and the surface area itself increased with rising stage. Continued infiltration probably induced temporary perched zones of saturation which were continuous with the river at places in the ground, which tended to reduce the losses. Entrapped air in the ground also impeded infiltration. The water was comparatively shallow in most sections of the channel; higher rates of infiltration would occur with greater depth of water.

Percolation in Playa Basins

The rates of water loss in the Big Lost River playas were measured when they contained ponded water in 1952 and 1953. The discharge through inflow and outflow channels of three playas was measured, the total water loss in each playa was computed, and the wetted areas were mapped on aerial photographs. The loss by evapotranspiration was estimated by computations based on records of the mean monthly evaporation at three weather stations on the Snake River Plain at Aberdeen, Jerome, and Milner Dam. The infiltration rates ranged from 0.719 cfd/ft^2 (5.4 gpd/ft^2) to 0.058 cfd/ft^2 (0.4 gpd/ft^2), and the average loss in the three playas was 0.308 cfd/ft^2 (2.3 gpd/ft^2) (table 6).

The ratio of the infiltration rate to the depth of ponded water in the playas varied considerably between playas and between the different times of measurement. The depth of water directly affects the rate of infiltration, but other factors also are important. Some of these factors were the length of time the playas were wetted, the nature of the underlying sediments, the time of year, the temperature, and the relative humidity.

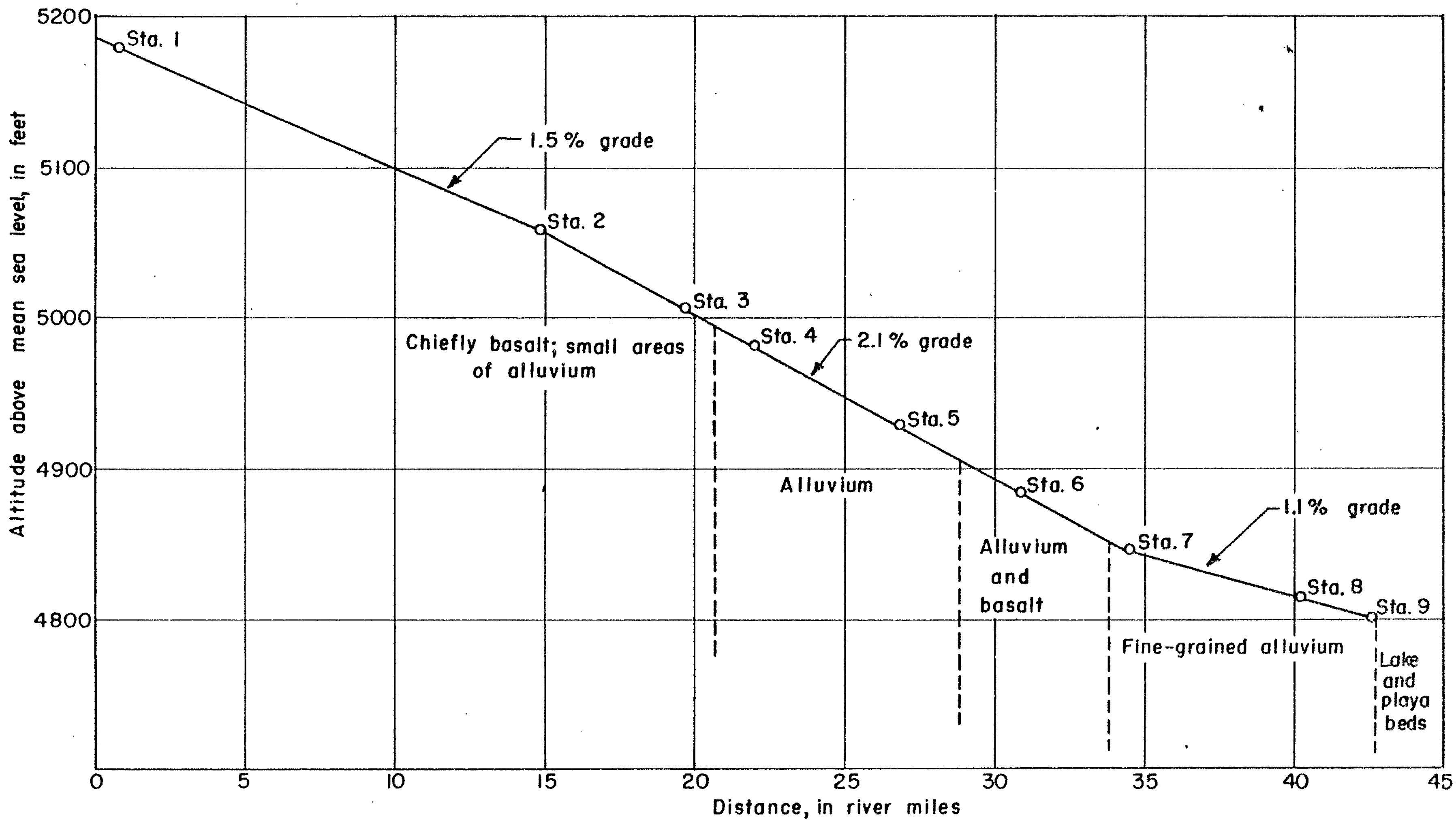


Figure 5.—Grade of channel of Big Lost River through the NRTS, Idaho.

Table 6.--Infiltration rates in the Big Lost River playas, 1952-1953

(Adjusted for evaporation loss)

Playa	Date	Discharge (cfs) Inflow channel	Discharge (cfs) Outflow channel	Wetted playa area Square miles	Wetted playa area Acres	Wetted playa area Square feet (thousands)	Total loss of water Cfs	Total loss of water Cfd (thousands)	Total loss of water Cfd/ft^2	Average mean daily evaporation (feet of water)	Infiltration rate Cfd/ft^2	Infiltration rate Gpd/ft^2
	1952											
1	May 1	20.3	-	0.084	54.4	2,370	20.3	1,750	0.738	0.019	0.719	5.4
1	May 24	84.2	-	.42	269	11,700	84.2	7,270	.621	.019	.602	4.5
1	May 26	68.2	-	.42	269	11,700	68.2	5,890	.503	.019	.484	3.5
1	June 27	83.5	37.3	.448	287	12,500	46.2	3,990	.319	.022	.297	2.2
1	Sept. [illegible]	7.1	-	.046	29.4	1,280	7.1	613	.479	.014	.465	3.5
	1953											
1	April 10	26.4	-	.258	165	7,190	26.4	2,280	.317	.016	.301	2.3
1	June 26	4.7	-	.034	21.7	945	4.7	406	.430	.022	.408	3.1
										Average	.469	3.5
	1952											
2	May 1	6.1	-	.100	64.0	2,790	6.1	[illegible]	.189	.019	.170	1.3
2	May 2	1.1	-	.033	21.1	919	1.1	95	.103	.019	.084	0.6
2	May 24	36.[illegible]	13.1	.144	92.2	4,020	23.1	2,000	.498	.019	.479	3.6
2	May 26	23.9	13.1	.144	92.2	4,020	10.8	933	.232	.019	.213	1.6
2	June 27	63.2	31.4	.144	92.2	4,020	31.8	2,750	.684	.022	.662	5.0
	1953											
2	April [illegible]	17.4	0.1	.144	92.2	4,020	17.3	1,490	.371	.016	.355	2.7
2	April 10	10.9	0.7	.144	92.2	4,020	10.2	881	.219	.016	.203	1.5
										Average	.309	2.3
	1952											
3	May 24	4.6	-	.068	43.5	1,890	4.6	397	.210	.019	.191	1.4
3	May 26	4.6	-	.068	43.5	1,890	4.6	397	.210	.019	.191	1.4
3	June 27	24.0	-	.430	595	25,900	24.0	2,070	.080	.022	.058	.4
										Average	.147	1.1
										Average of 3 playas	.308	2.3

Playas 1 to 3 are underlain by alternate layers, a few inches to several feet thick, of fine sand, silty sand, and clayey silt. Pebble gravel occurs in the central part of playa 1 beneath a thin cover of fine sediment. The uppermost layer of sediment beneath the playas at many places is a crusted layer of silt and clayey silt. The vegetal cover is sparse, consisting of scrubby desert shrubs and grasses. It was not practical to measure the depth of ponded water in the playas during the periods for which water losses were computed, but the depth in parts of playa 1 was 5 to 6 ft. In other playas the water at most places was only a few inches deep.

Most of playa 1 is in secs. 2 and 3, T. 5 N., R. 30 E., but a small part is in sec. 34, T. 6 N., R. 30 E. The maximum area of ponded water in the playa was about 287 ac (about 12,500,000 sq ft), at which stage some of the water spilled eastward into playa 2. The gross rate of water loss varied from 0.317 cfd/ft^2 (2.3 gpd/ft^2) to 0.738 cfd/ft^2 (5.4 gpd/ft^2), and the average was 0.469 cfd/ft^2 (3.5 gpd/ft^2).

Playa 2 is in sec. 2, T. 5 N., R. 30 E., and a small part of sec. 35, T. 6 N., R. 30 E. The maximum area of ponded water in the playa was about 92 ac (about 4,020,000 sq ft), at which stage water spilled eastward in an overflow channel toward playa 3. The gross rate of water loss varied from 0.103 cfd/ft^2 (0.6 gpd/ft^2) to 0.684 cfd/ft^2 (5.0 gpd/ft^2), and the average was 0.309 cfd/ft^2 (2.3 gpd/ft^2).

Playa 3 is very irregular in form and lies in sec's 4 and 5, T. 5 N., R. 31 E. and sec's 26, 27, 28, 32, 33, and 34, T. 6 N., R. 31 E. The total playa area is about 800 ac, but the maximum area under water in 1952 was about 595 ac. An outlet channel extends northeastward to playa 4, which is the terminus also of the Birch Creek drainage from the north. Water did not spread into playa 4. The infiltration rate in playa 3 was less than in the other two because the materials in its floor are less permeable. The gross rate of loss ranged from 0.080 cfd/ft^2 (0.4 gpd/ft^2) to 0.210 cfd/ft^2 (1.4 gpd/ft^2), and the average was 0.147 cfd/ft^2 (1.1 gpd/ft^2).

LITTLE LOST RIVER AND BIRCH CREEK

The Little Lost River spills into its playa basin adjacent to the NRTS each spring, except in very dry years, but no surface flow reaches the Station. Practically all the water in Birch Creek is diverted for irrigation in upstream areas and very little runoff reaches the Station.

LOCAL RUNOFF

Ephemeral runoff from the foothills bordering the NRTS on the northwest reaches the Station at times but rarely extends beyond the edges of the alluvial fans at the foot of the mountain slopes. Numerous short, ephemeral drainageways are scattered over the rest of the Station, where they drain from local slopes onto flats and small playas.

Many small playas are scattered through the NRTS (pl. 1, part 2), ranging in area from less than an acre to a few tens of acres. These receive ephemeral local runoff and occasionally contain ponded water. Many of them contained ponded water during thawing of an unusually heavy snow cover late in February 1952. Rye-Grass Flat, the playa just east of the junction of U. S. Highway 20 and 26, was covered by about 5 ft of water, and a playa 2 mi north of the highway junction was covered by about 2 ft of water.

FLOOD AND EROSION HAZARD

The flood hazard on the NRTS is relatively small. Birch Creek reportedly has not discharged water onto the Station since about 1894; it is not likely that much flood or freshet water originating in the upstream area ever will reach the NRTS. A local cloudburst might produce heavy flash discharge and that possibility should be considered in construction plans. Cloudburst drainage across the alluvial fan of Birch Creek, which extends into the northeastern part of the NRTS, would spread out in distributary channels and might form a sheet flood. A structure, such as an air strip with its long axis transverse to the direction of flow of the water, might be inundated and littered with debris at places.

The Little Lost River terminates in a small playa near Howe, outside the NRTS. That area is topographically low, and the possibility of flood water spreading beyond the playa is remote. In the event of a catastrophic flood, spill from the filled playa would be toward Big Lost River playa 1.

The Big Lost River is a more imminent flood and erosion hazard. Although the river seldom discharges much water as far as the bridge on Highway 20, it has reached there and beyond to the first playa in at least 11 of the years since 1920, and perhaps in more. The combination of circumstances that increased the discharge greatly in 1951-1953 may recur. During the high runoff period of 1952 the river reached nearly full-channel capacity at some places. Failure of Mackay Dam would unleash a serious flood.

From the rate of infiltration above the smaller channel sections that have been surveyed, we estimate that discharge at the gaging station probably would rise to 700 cfs or more before the constricted lower reaches of the channel and its distributaries would be bankfull. The two canals leading from the river at the old diversion dam (station 3) are complicating factors. At high river stages water might enter the canals, escape from them where the canal banks are breached by new road cuts, and spread over the flood plain. The minimum discharge capacity of the river channel could be increased greatly by modifications of critical sections of the channel and banks. Repair of diversion works and canals would help to control large floods because, above the diversion point, the river channel is incised at most places and is several times as large as the section of the channel that ordinarily is occupied by water. Controlled canal diversion of water during a flood might reduce the discharge below the diversion dam to a volume within the capacity of the channel. The possibility and probable extent of flood damage, and reasonable preventive measures, could be estimated only after further field study.1/

1/ After the field work for this report was completed, modifications were made of the river channel and old canals. These changes materially alter the runoff situation.

DEPTH AND VELOCITY OF WATER IN THE BIG LOST RIVER

No special study was made of channel erosion or of potential threats to over-channel structures, such as bridges. Inasmuch as the depth and velocity of water are functions of erosion, however, some data on the depth and velocity of water in the Big Lost River channel are summarized in Table 7.

The maximum depths of water listed in Table 7 are the maxima in cross-sectional areas where discharge rates were measured. They are not necessarily actual maximum depths in the stream reaches. Criteria for the selection of measuring stations commonly operate to avoid the deepest channel areas.

The "average" velocity of the water in a stream section where a discharge measurement is made is computed by dividing the rate of discharge by the cross-sectional area of the water in the measured section. The maximum measured velocity is a "representative velocity", rather than a true maximum, because the true maximum is not determined for discharge computations.

The velocities along the bottom and sides of a channel largely determine the erosive power, while the variation of velocity in the vertical column generally determines the characteristics and amount of suspended load that can be carried.

Table 7.—Representative velocities, depths and rates of discharge of water in the Big Lost River, 1951-53

[Station numbers refer to stations shown on pl. 1, Part 2]

Station	Date	Velocity (ft sec) Average (computed)	Velocity (ft sec) Maximum (measured)	Maximum depth (ft)	Discharge (cfs)
1	9-14-51	1.99	3.28	1.75	119
	5-22-52	3.16	4.29	3.20	483
	6-24-53	1.52	2.22	2.50	120
2	5-23-52	4.12	6.46	2.50	449
	6-25-53	1.92	2.46	1.45	79.5
3	9-14-51	2.70	4.15	1.24	64.5
	5-23-52	4.22	6.34	2.50	438
	6-25-53	2.36	2.88	1.80	71.2
4	9-13-51	1.75	2.15	1.36	56.1
	6-28-52	4.06	5.88	2.35	399
5	9-13-51	2.98	4.06	1.36	49.4
	5-23-52	3.99	5.99	2.30	415
6	8-17-51	1.60	2.45	1.20	19.4
	4-30-52	2.72	-	3.00	236
7	8-17-51	1.34	2.25	.80	14.3
	5-23-52	2.92	4.40	2.85	376
	9-17-52	1.62	1.89	1.18	63.6
8	8-16-51	.92	1.38	.61	6.86
	6-30-52	2.94	4.29	3.14	308
9	8-17-51	.62	.76	.62	4.77
	5- 1-52	3.20	-	3.20	209

USE AND QUALITY OF SURFACE WATER

Surface water is used on the NRTS only sporadically, as for road construction. The chemical quality of water in the Big Lost River is of interest chiefly because the river contributes to ground-water recharge. Table 8 shows the results of analyses of three samples from the Big Lost River. No analyses were made of water from the Little Lost River or Birch Creek. The principal dissolved ions in the water from the Big Lost River are silica, calcium, magnesium, sodium, potassium, bicarbonate, sulfate, and chloride. The water contains small amounts of iron and fluoride. It is moderately hard but is suitable in quality for all ordinary uses. No determinations were made of the suspended load of sediment in the water.

GROUND WATER

The ground water beneath the NRTS is part of the great regional body of water that underlies the entire Snake River Plain east of Bliss. The water occupies pores and other voids in the basalt and in sedimentary interflow beds associated with the basalt. It is replenished by several processes from numerous sources, but underflow from adjacent areas to the north, northeast, and northwest is the chief source of replenishment beneath the NRTS. A small amount of recharge occurs directly from precipitation on the station and from infiltration along the channel of the Big Lost River.

Theis, C. V., 1957. The spacing of pumped wells: U. S. Geol. Survey Open-file report (duplicated), 4 p.

U. S. Coast and Geodetic Survey: Annual Reports of United States earthquakes.

Wenzel, L. K., 1942, Methods for determining permeability of water-bearing materials, with special reference to discharging-well methods: U. S. Geol. Survey Water-Sup. Paper 887.

Nace, R. L., 1956, Ground water in the water economy of Idaho: U. S. Geol. Survey report (mimeographed), 23 p., January.

Parker, G. G. and Stringfield, V. T., 1950, Effects of earthquakes, trains, tides, winds, and atmospheric pressure changes on water in the geologic formations of southern Florida: Econ. Geology, vol. 45, no. 5, p. 450-452, August.

Shuter, Eugene, 1953. Records of wells and water-level fluctuations in western Bingham County, Idaho: U. S. Geol. Survey dupl. report, 96 p.

Shuter, Eugene, and Brandvold, G. E., 1952, Water levels in wells in Bingham, Bonneville, Butte, and Jefferson Counties, Idaho: U. S. Geol. Survey dupl. report, 99 p.

Slichter, C. S., 1905, Field measurements of the rate of movement of underground waters: U. S. Geol. Survey Water-Sup. Paper 140, 125 p.

Slichter, C. S., and Wolff, H. C., 1906, The underflow of the South Platte Valley: U. S. Geol. Survey Water-Sup. Paper 184, 42 p.

Stearns, H. T., 1928, Record of earthquake made by automatic recorders on wells in California: Seismol. Soc. America Bull., vol. 18, p. 9-15.

Stearns, H. T., Bryan, L. L., and Crandall, Lynn, 1939, Geology and water resources of the Mud Lake Region, Idaho, including the Island Park area: U. S. Geol. Survey Water-Sup. Paper 818, 125 p.

Stearns, H. T., Crandall, Lynn, and Steward, W. G., 1936, Records of wells in the Snake River Plain, southeastern Idaho: U. S. Geol. Survey Water-Sup. Paper 775, 139 p.

Stearns, H. T., Crandall, Lynn, and Steward, W. G., 1938, Geology and ground-water resources of the Snake River Plain in southeastern Idaho: U. S. Geol. Water-Sup. Paper 774, p. 163.

Stearns, H. T., Robinson, T. W., and Tayler, G. H., 1930, Geology and water resources of the Mokelumne area, California: U. S. Geol. Survey Water-Sup. Paper 619, p. 145-151.

Tayler, G. H., and Leggette, R. M., 1949, Ground Water in the Jordan Valley, Utah: U. S. Geol. Survey Water-Sup. Paper 1029, p. 76-81.

Theis, C. V., 1935, The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using ground-water storage: Am. Geophys, Union Trans., pt. 2, p. 519-524.

Theis, C. V., and others, 1954, Estimating transmissibility from specific capacity: U. S. Geol. Survey, Ground Water Branch (unpublished) note 24, July.

REFERENCES

Barraclough, J. T., 1952, Records of wells in western Jefferson County, Idaho: U. S. Geological Survey mimeographed report, June.

California State Water Pollution Control Board, 1952, Water quality criteria: Publication no. 3, 512 p., 4 figs, with Addendum no. 1, 164 p.

Cooper, H. H., and Jacob, C. E., 1946, A generalized graphical method for evaluating formation constants and summarizing well-field history: Am. Geophys. Union Trans., vol. 27, no. 4, p. 526-534.

Crosthwaite, E. G., and Scott, R. C., 1955, Ground water in the North Side Pumping Division of the Minidoka Project, Minidoka Co., Idaho: U. S. Geol. Survey Circ. 371, 20 p.

Fancher, G. H., Lewis, J. A., and Barnes, K. B., 1933, Some physical characteristics of oil sands: Proc. Third Pennsylvania Mineral Industries Conference, sec. 6: Pennsylvania State College Mineral Industries Exp. Station Bull. 12, p. 159.

Ferris, J. G., 1949, "Ground water," chap. VII, in Wisler, C. O. and Brater, E. F., Hydrology: John Wiley and Sons, Inc., New York, p. 198-272.

Humphrey, P. A., and Wilkins, E. M., 1953, Engineering climatology of the ANP site, National Reactor Testing Station: U. S. Weather Bur. report no. IDO-10021 (mimeographed).

Jacob, C. E., 1947, Drawdown test to determine effective radius of artesian well: Am. Soc. Civil Engineers Trans., vol. 112, p. 1047-1070.

__________, 1950, "Flow of ground water", chap. V, in Rouse, Hunter (ed.) Engineering hydraulics: John Wiley and Sons, Inc., New York, p. 321-386.

Knowles, D. B., 1955, Ground-water hydraulics: U. S. Geol. Survey open-file report, Ground-Water note 28, 105 p.

Leggette, R. M., and Taylor, G. H., 1937, Geology and ground-water resources of Ogden Valley, Utah: U. S. Geol. Survey Water-Sup. Paper 796-D, p. 128.

Meinzer, O. E., 1928, Compressibility and elasticity of artesian aquifers: Econ. Geol., vol. 23, no. 3, p. 263-291, May.

Mower, R. W., 1953, Records of wells and ground-water levels in eastern Jerome County, Idaho: (mimeographed), June.

The bulk of the determinations of beta-gamma activity were made in the laboratory of the Atomic Energy Commission, and this laboratory, taking into account its procedures and the sensitivity of its counting apparatus, considers that the background (natural activity) for beta-gamma activity is 200×10^{-12} c/L (curies per liter).

Applying these criteria, we find that about ten pct of the sampled sources have had at one time or another, somewhat more than the background level of beta-gamma activity. Less than 5 pct of the sources have had values of 400×10^{-12} curies per liter, which is only twice the background value. The values above background are not only few, but nonpersistent, for they occur only occasionally in a series of determinations from a given source.

From the foregoing it appears that the radioactivity of ground waters in the Snake River Plain is generally low, and commonly it is less than or barely exceeds the limit of detectability against the background radioactivity. A few samples have appreciable activity but it does not appear to be significantly higher than that of other natural waters from many other parts of the west; however, valid comparisons with waters elsewhere cannot be made at present because of the differences in sampling and counting methods used by the several organizations and laboratories that are now making measurements of radioactivity of waters.

No pattern of levels of activity has yet been detected, though it would be normal for one to exist. If so, it will be detected eventually.

to the west and southwest. Temperature readings are not available for all seasons of the year, but those recorded probably are representative, because seasonal variations are small.

Some pumped samples of water contained mixtures of water from several productive zones in which the temperatures probably differ somewhat. The temperatures of those samples would not reliably show the geothermal gradient, but from the standpoint of plant operation they are the effective temperatures. In general, higher temperatures occur at greater depth, but the trend is not very systematic. As was mentioned earlier, waters from certain wells near the mountain front — for example, wells 3N-27E-9ab1 and 3N-27E-9ab2 — have unusually high temperatures. This may be caused by hot waters rising along faults. Also, the Snake River Plain was the scene of comparatively recent volcanic activity, and the geothermal gradient probably is steep. Local temperature anomalies may arise from residual heat of volcanism.

RADIOACTIVITY IN THE WATER

Since the establishment of the NRTS, the Atomic Energy Commission has made periodic determinations of the radioactivity of the waters of the Snake River Plain. The Geological Survey has made a small number of check determinations on samples from the same sources. The purpose of this sampling was to determine the normal values of radioactivity in these waters and to detect whatever changes occur in time. The surface and ground waters have been sampled at many points and repeatedly at a number of chosen points. A compilation of the results is given in Table 12, appendix 2.

The chemical characteristics of these two waters probably vary seasonally, especially in the shallower aquifer tapped by well -7adl, owing to seasonal variations in the relative proportions of recharge by underflow from distant sources and recharge from nearer sources. Much more study is needed of the chemical characteristics of the waters.

TEMPERATURE OF THE GROUND WATER

Diurnal changes in soil temperature extend to depths of only a few inches or feet, and seasonal changes extend but little deeper. In ordinary rock and soil, the temperature a few feet below the land surface is about the same as the mean annual air temperature in that area. The temperature of ground water in very shallow aquifers also is near the mean annual air temperature, but some fluctuation is caused by variations in the temperature of recharge water.

Beneath the upper few feet of earth or soil, the earth temperature increases downward, so that ground water from successively deeper zones is apt to be increasingly warm. Moreover, deep water is apt to be more nearly constant in temperature throughout the year than is shallow ground water. The temperature in wells may be increased very slightly during prolonged heavy pumping, because of the heat of friction in a pumped aquifer.

Water from wells and test holes on the NRTS ranges in temperature from 50 to 66°F, and the arithmetic mean of the temperatures is 56. The temperature in wells nearby on the plain ranges from 45 to 89°, but most of the higher temperatures are in wells near the flanks of the mountains

Chemical analyses show certain anomalies in the chemical character of waters from different parts of the NRTS. For example, water from test-hole 3N-29E-19cd1 differs in chemical properties from the water from other wells in that general area. Noteworthy is the presence of chlorides, 47 ppm; nitrate 4.1 ppm; noncarbonate hardness, 46 ppm; and dissolved solids 228 ppm. The water from wells in that general vicinity is somewhat warmer than at most locations on the NRTS. Some privately owned deep wells to the northwest, near the flank of the Big Lost River Range, disclose occurrences of high chloride and warm temperature in that area. The water may be associated with a fault along the flank of the mountains. Similar conditions may prevail along the mountain flank northeastward from Arco to Howe. Warm mineralized water thus could be moving from the border of the plain southward and southwestward under the NRTS. If flow of the water in the aquifer is streamlined, marked differences in the chemical quality of water may occur in relatively short distances.

A marked difference in the chemical quality of the water samples collected from two wells in the central part of the NRTS is related directly to their origin from different depths (table 10, Appendix 2). Well 4N-30E-6ab1 derives water from basalt at a depth of 1,407 to 1,480 ft, and well -7ad1 taps water in basalt from a depth of 387 to 518 ft. The water from well -7ad1 does not differ greatly from that in other wells which tap the basalt aquifers, but the sample from well -6ab1 differs markedly from all others collected on the Station. The water is very soft, low in silica, calcium, magnesium, and sulfate, and high in sodium, fluoride, and boron (?). The temperature of 66°F is the highest recorded on the Station.

partly because the basaltic aquifer is a "fast-circulating" system which does not retain water in storage as long as is common in more slowly circulating system.

Local Variations in Quality

With a few exceptions, the ground-water samples tested are generally similar in chemical composition. Waters from wells near the irrigated areas to the northeast, north, and northwest of the NRTS tend to contain somewhat more dissolved solids than water elsewhere. Unconsumed irrigation water from the Mud Lake basin leashes mineral matter and agricultural chemicals from the soil as it percolates downward, and increments from that source change the chemical composition of the water as it moves down-gradient. The greatest increase in concentration in the wells adjacent to the irrigated areas commonly is in Ca and Cl.

In some wells scattered through the central part of the NRTS the percent sodium is slightly higher than that in most other wells on the Station but much lower than in most wells in the southern and western parts of the plain. The cause of the difference was not determined, but it existed before the beginning of large operations on the Station.

Three wells in the northeast quarter of section 9, T. 3 N., R. 27 E., yield waters having appreciably higher temperatures than in those of water from other wells in that general area. The amount of dissolved solids in the wells also is comparatively high and is probably related to the temperature, because of the higher solubility of some minerals in waters of higher temperature. The temperature and quality may make the water less desirable for some purposes or necessitate more treatment.

area in the Mud Lake basin. Therefore, one might expect to find some chemical differences between the water coming from the north and west and that from the northeast.

The information now available discloses no obvious correlation of the chemical quality of the water with the stratigraphy, depth, or mineralogy of the aquifer. The proportionately large amount of bicarbonate (doubtless chiefly of Ca and Mg) is characteristic of waters from the basalt generally (see table 11, Appendix 2). Calcium and magnesium are essential constituents of the principal minerals in the Snake River basalt. Much of the water is derived by underflow from the adjacent mountainous area, where limestone ($CaCO_3$) and dolomitic limestone ($Ca,MgCO_3$) are common rocks. The soil and gravel through which the water passes in intermontane valleys also contains abundant calcium carbonate. Undoubtedly, ground water that is recharged directly from precipitation on the plain also must be charged with $CaCO_3$, because the plains soils are rich in that mineral. The total dissolved solids and the carbonate-bicarbonate content increase westward in the plain, probably from two causes: (1) longer contact with the aquifer rock, and (2) increased contributions of recharge from irrigated areas where percolating water is more mineralized. The basalt minerals are not readily soluble in water, but some leaching occurs of silica, iron, manganese, calcium, magnesium, sodium and potassium, all of which are present in minerals in the basalt.

The generally low amount of dissolved solids in most of the samples is indicative partly of the low solubility of minerals in relatively fresh and unaltered basalt; partly of a lack of opportunity for solution to occur, owing to the nearness of some recharge areas to the NRTS; and

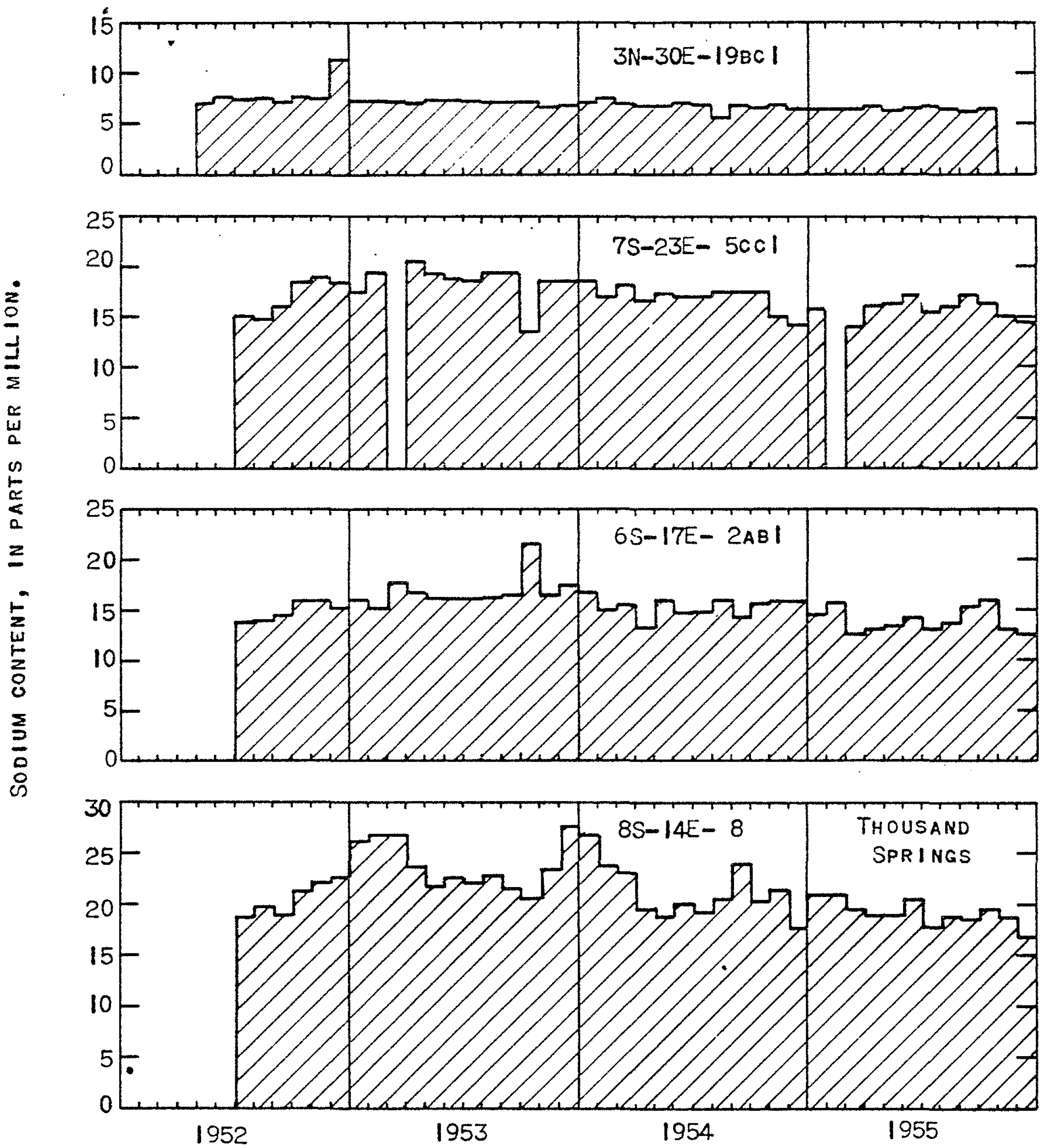


FIGURE 31.--SODIUM CONTENT OF GROUND WATER AT LOCALITIES ON THE SNAKE RIVER PLAIN.

The graphs in figure 31 also disclose certain changes in sodium content with time. All four graphs indicate a slight decrease in sodium content during the period of sampling, from early in 1952 to the end of 1955. This decrease may be correlated with greater recharge during this period, when precipitation was above average. The graph of sodium content of the water from the Thousand Springs shows some evidence of a seasonal cycle in which the sodium content is greatest in winter, least in summer.

Geologic Significance of the Chemical Quality

Water is a geochemical agent and its chemical character is a result of the geologic environment through which it has passed and the length of time spent in that environment. The dissolved constituents in the water are derived from minerals with which it has come in contact (ignoring, for the moment, artificial contaminants introduced in the water). The concentration of minerals in the water depends on their solubility, the temperature of the water, the chemical composition of the water at the time of contact, and geochemical changes, such as ion-exchange, that occur as the water moves through the aquifer.

The ground water on the NRTS is derived by recharge from several distinct sources (see p. 7-8). Water from the north and northwest is derived from a terrane of Paleozoic sedimentary rocks, Mesozoic intrusive rocks, and Tertiary volcanic and pyroclastic rocks. Dolomitic limestone is abundant in parts of that terrane. Water entering the area from the northeast has had a long and complicated history, including passage through a wide range of types of host rocks. In a late stage before entering the NRTS, much of the water is cycled through the irrigated

The discharge of organic and industrial wastes by plants on the NRTS has had no observed effect on the quality of water pumped from wells outside the Station. Except for a few cases, the quality of water in the NRTS is suitable for domestic, agricultural, and ordinary industrial use.

Descriptions of the relationship of various chemical ions to the utility of water are available in standard works on that subject. A rather exhaustive summary of the relationship of the utility of water to its chemical constituents was published by the State of California (1952).

Special information on water quality has been collected by the Commission since 1952. Samples from wells and springs scattered throughout the Snake River Plain have been analyzed for sodium and radioactivity. Waste water discharged to the ground at the NRTS contains considerable amounts of ordinary sodium chloride, and slight amounts of radioisotopes. Presumably, continual plainswide monitoring of sodium and radioactivity in the ground water would give a check on whether contamination occurs.

The graphs in figure 31 show the sodium content of water from three representative wells on the Snake River Plain and from the Thousand Springs, a principal outlet for the ground water of the plain. They indicate that, in general, the waters acquire sodium in transit through the aquifer from northeast to southwest, or that the water from the northeast mingles, in transit with water from other sources of recharge. The water that drains from irrigated tracts and enters the ground is a principal source of dissolved sodium. Certain wells in sediments along the southern edge of the plain, especially the Rupert city well and to a lesser extent the Eden and Aberdeen city wells, contain much greater amounts of sodium than do wells in the basalt. Graphs for these and other wells are included in appendix 2.

Table 24.--Summary of chemical properties of representative ground waters from the Snake River Plain--Continued

Constituent or property	Reactor Testing Station			Central Snake River Plain		
	Minimum	Maximum	Arithmetic mean	Minimum	Maximum	Arithmetic mean
Boron (B)	0.00	0.11	0.037	0.00	0.25	0.065
Dissolved Solids:						
ppm	144	354	218.1	143	821	293.8
tons per ac-ft	.20	.50	.300	.19	1.12	.40
Hardness as $CaCO_3$:						
total	44	265	155.3	85	516	178.3
noncarbonate	0	84	22.0	0	380	25.5
Percent sodium	3	26	12.0	0	79	21.7
Specific conductance ($K \times 10^6$ at 25°C.)	182	540	345	257	1260	463.5
pH	7.2	9.5	7.9	7.1	9.4	7.83

1/ Analyses of samples of spring water are not included in this table. Some reports of analyses of water combined Na and K values; those values were not used to develop this table.

Table 24.--Summary of chemical properties of representative ground waters from the Snake River Plain 1/
[Chemical constituents in ppm; analyses by Geological Survey]

Constituent or property	Reactor Testing Station			Central Snake River Plain		
	Minimum	Maximum	Arithmetic mean	Minimum	Maximum	Arithmetic mean
Temperature (°F)	50	66	56	45	89	55.9
Silica (SiO_2)	15	39	25	8.1	50	32.1
Iron (Fe) dissolved	.00	.34	.052	.00	.38	.058
Manganese (Mn)	.00	.58	.035	.00	.93	.043
Calcium (Ca)	7.7	70	39.3	16	141	43.4
Magnesium (Mg)	2.8	22	14.05	9.3	40	16.4
Sodium (Na) 1/	2.6	20	9.62	3.3	155	28.2
Potassium (K) 1/	1.0	15	3.18	1.6	20	4.35
Bicarbonate (HCO_3)	58	226	165.9	99	422	188.8
Carbonate (CO_3)	0	18	.60	0	58	1.44
Sulfate (SO_4)	12	42	23.8	9.1	210	36.5
Chloride (Cl)	6.0	58	15.4	4.2	265	29.3
Fluoride (F)	.0	.7	.24	.0	2.0	.53
Nitrate (NO_3)	.02	5.5	2.15	0.0	72	4.41

1/ See footnote at end of table.

Differences of several degrees in temperature may be of interest in connection with plant designs.

Analytical Results

The results of chemical analyses of waters from the NRTS and other parts of the Snake River Plain are summarized in table 24, and the detailed results are given in Appendix 2, tables 10 and 11. The chemical character of some of the samples was affected by conditions during their collection. Many samples, for example, necessarily were obtained by bailing from test holes that were not equipped with pumps. In some cases the water had been in contact with clayey mud in the holes and may have increased its sodium concentration by ion-exchange for calcium in the water. In a few instances, cement grout had been used during drilling, and the cement produced an increase in the hardness of the water. In order to clarify possible discrepancies in the analyses, a complete list of the samples and descriptions of the circumstances of their collection are given in Appendix 2.

The quality of the sampled waters was generally excellent for ordinary uses. In all but two samples the amount of dissolved solids was appreciably less than 500 ppm, the upper limit specified by the U. S. Public Health Service for desirable drinking water. Softening treatment probably would be required for most industrial uses (aside from demineralizing for special use), but the water is not excessively hard. Water from well 3N-27E-9aa1 had 265 ppm of chloride, which is slightly higher than the maximum of 250 ppm recommended by the U. S. Public Health Service for drinking water.

is large, the volume of underflow from tributary areas is considerable, the seasonal variations in underflow probably are small, and the movement of ground water is slow. Because of the considerable time between periods of reduced precipitation and decline of the water table, long-term records of water-level fluctuations are essential for satisfactory interpretation of the complex interrelationships between precipitation and water-level fluctuations. The overall effect of deficient precipitation on the water table probably would be a general trend in water-level fluctuations similar to that in USGS wells 1 and 4 (figs. 16 and 28). The hydrographs of these wells show several years of declining water levels, followed by a steady rise, although precipitation during the period of record averaged below normal.

Water-level fluctuations in the area do not correlate closely with local precipitation because local replenishment by precipitation is comparatively small. Most replenishment is from sources outside the NRTS. Very little is known about the relative importance of the various sources as contributing factors in water-table fluctuations, but the relationships between them probably vary from time to time.

CHEMICAL QUALITY OF THE GROUND WATER

The ground waters tapped by wells and test holes 1,200 ft or less in depth on the NRTS are basically similar in chemical type. Some variations in chemical quality occur, both with location and with depth, and the quality at some places might influence the locations of facilities having very specialized water requirements. There is a slight regional variation in water temperature, and more noticeable variations with depth.

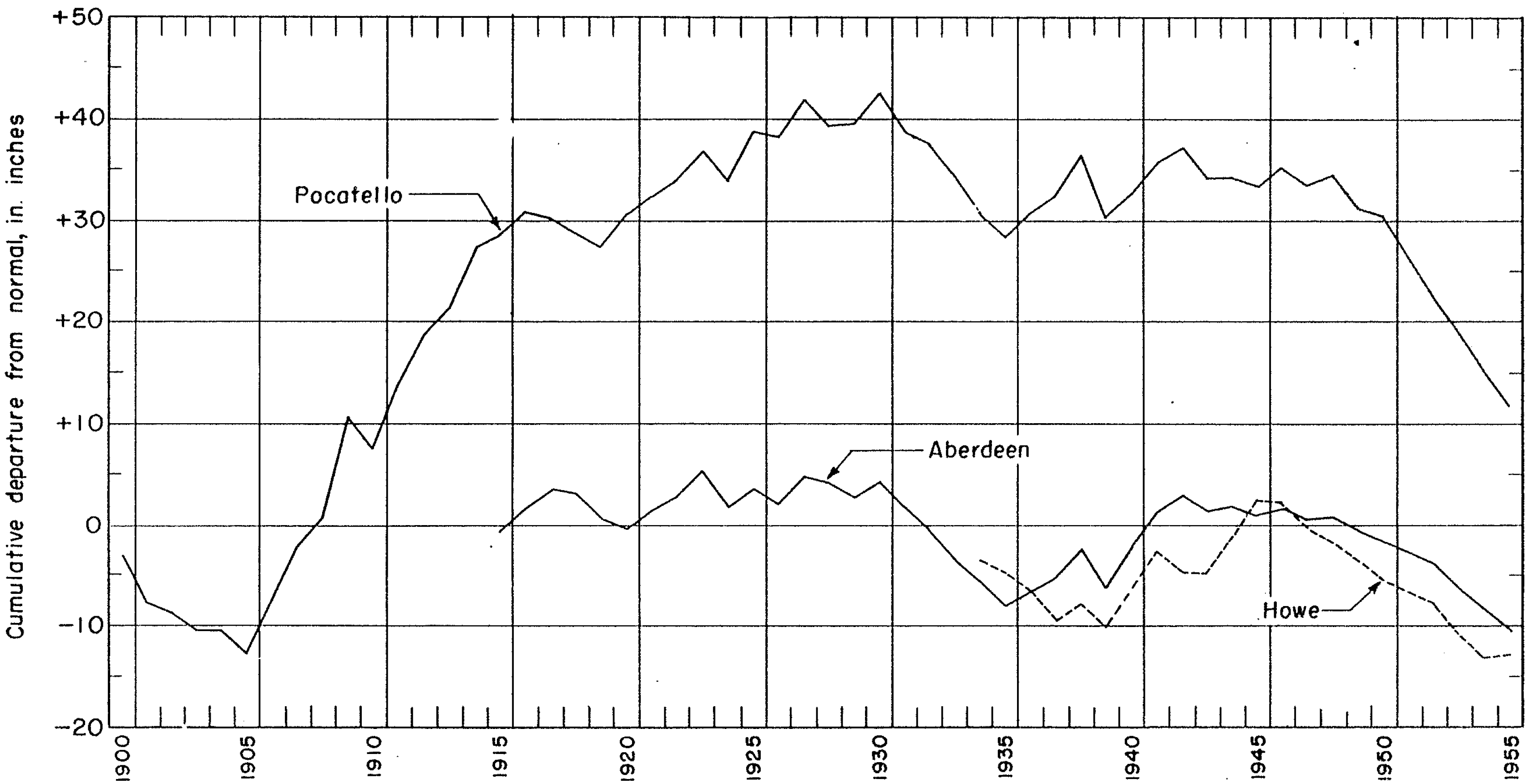


Figure 30.—Cumulative departure from normal yearly precipitation at three stations on the Snake River Plain.

The plotting of cumulative departures tends to "iron out" the fluctuations in years of short-term departures, such as single years when precipitation is deficient or excessive. A recent "wet" cycle began about 1935 and reached a climax in 1942, when the cumulative departure of precipitation was plus 3 in. at Aberdeen and 37 in. at Pocatello. A subsequent drier period continued through 1955, and the accumulated excess at Pocatello was reduced to about 12 in., and an accumulated deficiency of 10 in. developed at Aberdeen. At Howe a drier period began in 1946 and continued in 1954. At the end of 1954 the accumulated deficiency at Howe was about 13 in. Precipitation was slightly above normal at Howe in 1955 and the accumulated deficiency was slightly less than 13 in. Accumulated deficiencies prevailed in 1955 at Idaho Falls, Arco, and Spencer.

The water level in wells in the central and northern parts of the plain began to decline in 1954 and by the end of 1955 were at record low levels. Decline began in the southern part in 1953 and also reached record lows in 1955. In the Minidoka area, water levels in most wells began a general decline in 1954, and by the end of 1955 the net decline amounted to about 2 to 4 ft (see fig. 29).

Precipitation at most stations on the Snake River Plain averaged below normal during 1943-55, but water-level records are not long enough to show the effects of reduced recharge as a consequence of declining precipitation. The Snake River Plain contains a very large ground-water reservoir on which several years of subnormal precipitation probably would have little noticeable effect. Several to many years might elapse before the water table in most parts of the plain would show appreciable effects of reduced recharge, because the amount of ground-water storage currently about a foot below the 1950 levels. In wells where there are markedly regular annual fluctuations, as in Tps. 5-6 N., Rs. 33-34 E., water levels declined about a foot during 1950-52, but recovered during 1952-54. In 1955 they declined about a foot. On the other hand, water levels in wells in Tps. 2 N., to 1 S., Rs. 28-31 E. remained fairly constant during 1950-52 but declined about 1 to 2 ft from 1952 to 1955. As was noted earlier (p. 229), a decline in the southern part of the NRTS began about a year sooner than in the Minidoka area (see fig. 29). Data are not sufficient to show a relationship between the declines in water levels in the two areas. In general, water levels in most wells in the eastern Snake River plain have declined gradually since about 1953-54 and are now at record low levels in many parts of the plain.

Effects of precipitation.—In general, the water-level trends during the period of record seem inconsequential despite a steady cumulative decline in precipitation after 1943 at weather stations on the Snake River Plain. Figure 30 shows the cumulative departure from normal precipitation. The curves were obtained by algebraic addition of annual departures from normal plus the cumulative total obtained for the previous years, at three stations on the Snake River Plain. An upward slope to the right indicates a cumulative increase in precipitation and a downward slope to the right indicates a decrease. The longest continuous period of record is from the Pocatello station, beginning in 1900. Except for a few records which began about 1934, records from other stations on the plain cover periods beginning between 1905 and 1920.

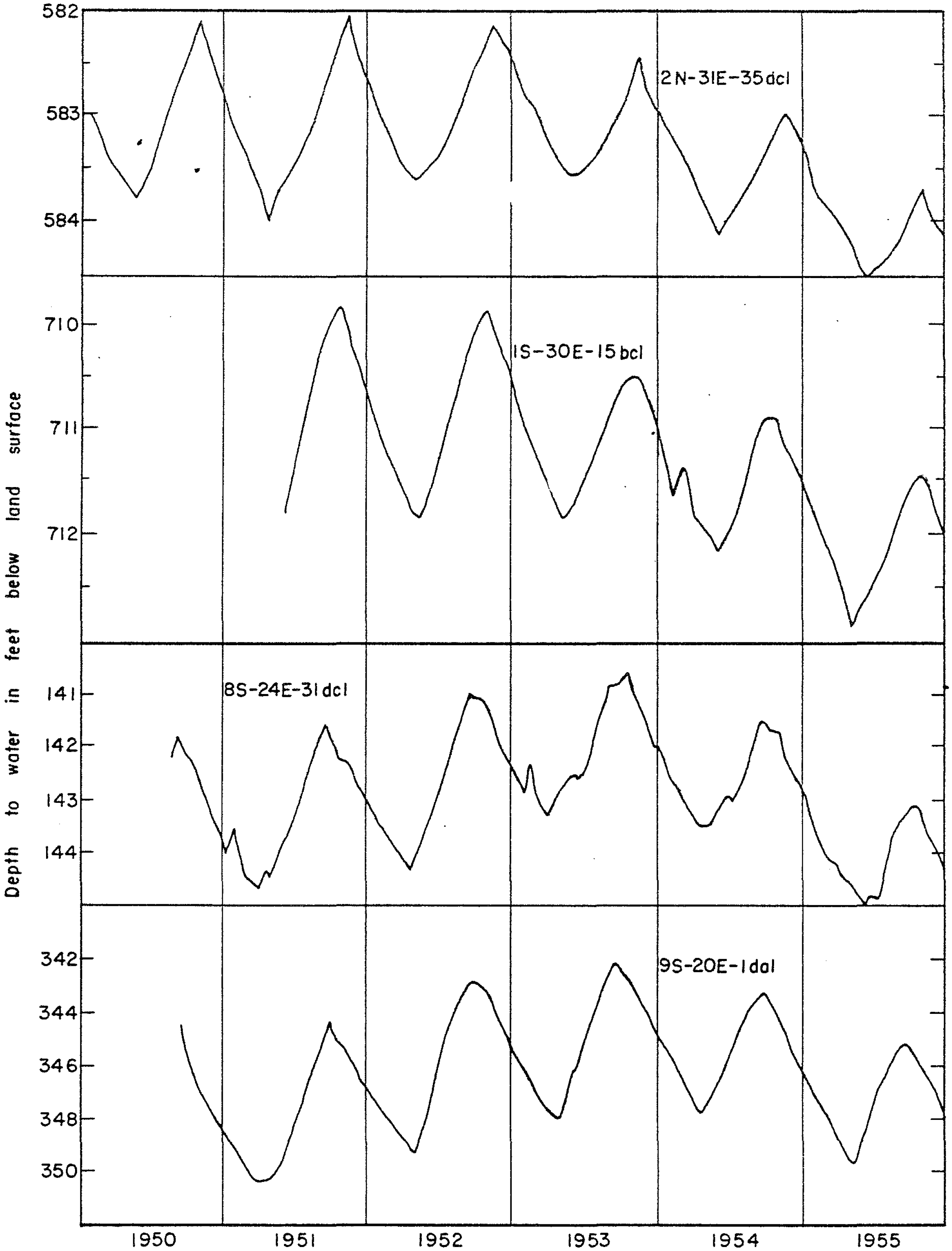


Figure 29.— Hydrographs of wells 2N-31E-35dcl and 1S-30E-15bcl near NRTS, Bingham County (above), and wells 8S-24E-31dcl and 9S-20E-1dal in Minidoka North Side Pumping Division, Jerome and Minidoka Counties (below).

and 1 to 2 ft additional in 1955. The decline began in the Minidoka area about a year later than in Bingham and Butte counties. Figure 29 shows the water-level fluctuations in several wells near the NRTS and in the Minidoka North Side Pumping Division in Jerome and Minidoka Counties. No direct relation between the declines in the two areas is known.

Long-Term Trends

Long-term trends (net changes of water levels at corresponding times in successive years) are an effect of changes in ground-water storage. Factors that influence the trends are weather variations, irrigation, pumping, and changes in the surface-water regimen. Records of water-level fluctuations in the NRTS and vicinity span only a 7-year period and good area coverage is available only for a 5-year period; these records define only short-term trends.

The trend in water levels in wells near the ANP and IET sites (T. 6 N., Rs. 31-33 E.), where the yearly range of fluctuations is small, has been slightly upward since about 1950, with a net rise of about 0.4 ft near the ANP area. About 8 mi northeast of the ANP area, in a privately owned well (7N-33E-35bb1) near the southwestern edge of the Mud Lake area, the range greatly exceeds that in wells at the ANP and IET sites; the net rise of the water table has been about 2 ft since 1950. Water levels in these wells declined slightly after July 1955. In wells affected by underflow from the valleys of the Big Lost and Little Lost Rivers and Birch Creek, the net rise from December 1951 to May 1953 was about 7 to 9 ft in the northern wells and about 4 to 6 ft in the southern wells. Water levels declined steadily after 1953 and in 1955 were

Project included 32 wells from which 54,000 ac-ft of water was pumped to irrigate about 14,200 ac; about 140 private wells withdrew about 108,000 ac-ft for 29,800 ac of land. The gross water requirement for the completed Government project will be 235,000 ac-ft yearly and that for private development may be about 325,000 ac-ft (Crosthwaite and Scott, 1955). The heavy pumping may lower the water table a few feet and bring in water by underflow from adjacent areas. Thus far there has been no appreciable interference among wells and no regional decline of water levels in the Snake River basalt attributable to pumping in the Minidoka area.

Immediately west of the Minidoka area, in eastern Jerome County, about 28,000 ac-ft of ground water was pumped from 36 irrigation wells to serve about 9,700 ac of land in 1955. A report by Mower (1953) contains records of wells and ground-water levels in the area.

In general, water levels in the Minidoka area rise and fall in an annual cycle similar to that in wells south of the NRTS in Tps. 2 N. to 1 S., R. 28 E. The downward trend usually starts in October or November and the rise usually begins in April or May. The annual range of fluctuations in the Minidoka area is about 1 to 5 ft (Crosthwaite and Scott, 1955). In wells south of the NRTS the annual range in fluctuations is 1 to 1.5 ft.

During 1950 to 1952 water levels remained rather constant in wells in Tps. 2 N., to 1 S., Rs. 28-32 E., Bingham and Butte Counties, but in 1953 the levels declined about 0.5 ft; by the end of December 1955 an additional foot of decline occurred. During 1948-53 water levels in the Minidoka area showed a net rise of 0.5 to 3 ft (Crosthwaite and Scott, 1955). Thereafter, the water levels declined about 1 to 2 ft in 1954, ranged from about 300 to 2,500 gpm. By 1955 the estimated number of wells increased to 160, which produced about 60,000 ac-ft of water and served about 18,000 ac of land.

Water levels in wells that reach the regional water table in the Roberts-Taber area ranged between about 164 and 720 ft below the land surface (see Appendix 2). The depths of the wells range from 219 to 786 ft. High water levels occur from October to November and low levels from April to June. The wells were measured bimonthly and the actual highest and lowest levels probably were not detected, but the measurements probably correctly indicate the months in which maximum and minimum levels occur. The yearly range of water-level fluctuations averaged between 2 and 4 ft, but in a few wells south of Roberts the yearly range was about 15 ft. Water levels in a few wells south and southeast of Mud Lake declined about 2 to 3 ft from 1950 to 1955.

Minidoka area.--In the Minidoka area, ground water is being developed extensively for irrigation by the U. S. Bureau of Reclamation in the North Side Pumping Division of the Minidoka Project, and by private enterprise for large adjacent tracts of land. The new developments center in southern Minidoka County and parts of Lincoln and Blaine Counties, about 55 mi southwest of the NRTS. Within a few years this area probably will be the most heavily pumped of those currently under development. The present total withdrawals are second in amount only to those in the Mud Lake basin. In 1955 about 162,000 ac-ft of ground water was pumped in the Minidoka area to irrigate 44,000 ac of land. The ground-water development began in 1947 with the drilling of three private irrigation wells. By 1955 the number increased to more than 170. In 1955 the U. S. Bureau of Reclamation the shallow wells remained fairly steady, without much change during the period of record. Water levels in the Mud Lake area rose appreciably only to the west of Mud Lake in Tps. 6 and 7 N., R. 33 E. Well 7N-33E-35bb1 is 70 ft deep and the water level is very responsive to recharge by infiltration of irrigation water. A seasonal rise in the water level begins late in each spring, shortly after the start of irrigation. The higher levels are reached in July and August and lower in May. The yearly range is between 7 and 8 ft, and the net rise since 1950 is about 3 ft (see fig. 27). Well 6N-33E-2ba1 is 245 ft deep and its water level also is affected by irrigation, but the effect occurs later in the year and is less pronounced than in the shallow well. The water level in the well has trended steadily upward since 1950 and the net rise at corresponding times of the year was about 2 ft (see fig. 27).

The water levels in wells that reach the regional water table about 6 to 10 mi south of Mud Lake, are not noticeably affected by deep seepage from irrigation. Daily water-level records for well 5N-34E-9bd1, 1950-55, show no unusual fluctuations that could be attributed to irrigation in the Mud Lake area. The long distance of the wells from Mud Lake and the large ratio of storage in the Snake River basalt to underflow from the basin probably are the reasons for the lack of noticeable effects.

Roberts-Taber area.—The Roberts-Taber area, east and south of the NRTS, extends from the vicinity of Roberts to Taber in a broad arc along the north side of the Snake River. The estimated number of irrigation wells in the area south of Taber and north of Aberdeen and Springfield was 55 in 1953 (Shuter, 1953), and the estimated pumpage was about 23,000 ac-ft of water for 7,000 ac of land. The rate of discharge from single wells discharge being by ground-water underflow to the Snake River Plain. The principal pumpage is done where the depth to water ranges from a few feet to about 50 ft below the land surface. Numerous flowing artesian wells furnish much of the irrigation water. About 99 pct of all the ground-water pumpage is for irrigation.

The estimated pumpage of ground water from 61 irrigation wells and 86 stock and domestic wells in 1952 was 83,300 ac-ft (Barraclough, 1952), and the water supplied about 58,000 acres of land. The yields of single irrigation wells ranged from 360 to 9,900 gpm and averaged about 3,000 gpm. Pumpage in 1955 was about 266,000 ac-ft, derived from 165 irrigation wells for use on about 90,000 acres (Nace, 1956). Therefore, ground-water pumpage increased by about 300 pct in a 3-year period, and the total irrigated acreage more than tripled.

Records of water levels in the Mud Lake basin have been published (Stearns and others, 1936; Shuter and Brandvold, 1952; Barraclough, 1952). Water-level records are available since 1949 for wells tapping shallow ground water northeast of Mud Lake in Tps. 7-8 N., Rs. 36-37 E. The depths of the wells range from 45 to 93 ft and the average depths to water in single wells range from 13 to 50 ft. The yearly range of fluctuation is about a foot. High levels occur in April and May and low ones in August and September. The net trend in levels after 1953 was slightly downward, amounting to about a foot at the end of 1955. Wells south of Mud Lake in Tps. 5-6 N., Rs. 34-35 E. range from 280 to 930 ft in depth and from 50 to 160 ft in depth to water (see Appendix 2). The deeper wells reportedly tap several water-bearing zones. Water levels in the deep wells declined as much as 20 ft after 1951, whereas water levels in

Water levels in the shallow wells are very responsive to infiltration of irrigation water. The higher levels are reached in July and August at the height of the irrigation season, and the low levels are reached in April and May, shortly before the irrigation season begins. Water levels in the deeper wells are not noticeably affected by recharge from infiltration of irrigation water.

Howe area.—The Howe area, around the mouth of the Little Lost River Valley in Tps. 5-6 N., Rs. 28-29 E., is near the western-central part of the NRTS. The principal uses of ground water around Howe are for domestic, stock, and irrigation supplies. Records of water levels in a few wells in the area are contained in an earlier report (Shuter and Brandvold, 1952). The estimated ground-water pumpage in 1949 was 1,300 ac-ft from 10 wells. About 3,500 additional ac-ft was pumped in the valley about 5 to 10 mi north of Howe. Pumpage in the vicinity of Howe in 1955 was about 1,500 ac-ft and that north of Howe about 4,000 ac-ft.

The average depth to water in the Howe area ranges between about 65 to 105 feet (see Appendix 2). Weekly water-level records from a few wells are available for the period December 1949 through October 1953, and monthly and quarterly records thereafter. The highest recorded water level in the area occurred in August 1952, about 7 to 9 months earlier than the peak level of that year in wells in the NRTS. The annual range in fluctuation averaged 1 to 3 ft. Water levels were fairly constant through December 1952 but declined thereafter about 1 to 2 ft.

Mud Lake basin.—Ground-water withdrawals in the Mud Lake basin, north and east of the NRTS, are the largest in Idaho and totaled about 270,000 ac-ft in 1955. There is no surface runoff from the basin, all

7 mi west of the western edge of the NRTS. About 33,000 acres of land in the Big Lost River Valley, from Mackay to about 3 mi southeast of Arco, is irrigated with surface water from Mackay Reservoir on the Big Lost River. About 9,000 additional acres is irrigated above the reservoir. Ground water in the Big Lost River Valley is used chiefly for domestic, stock, and municipal supplies. A few irrigation wells have been drilled since 1954, principally along Antelope Creek, but little is known about the depths and yields of the wells. The largest single withdrawal of ground water is for the municipal supply of Arco, which averages about 250,000 gpd. The municipal system includes three wells ranging in depth from about 28 to 60 ft. The depth to water ranges from 12 to 30 ft and the yields of the wells average about 400 gpm. The Village of Butte City, about 2.5 mi southeast of Arco, has one deep well which taps the regional water table. The depth of the well is about 443 ft and the depth to water is about 400 ft. The yield during a test in 1952 was about 60 gpm. Except for a few wells tapping the deep aquifer, most wells in the Arco area range in depth from 8 to 140 ft, and the depth to water ranges from 3 to 55 ft (see Appendix 2).

In the shallow water table in the vicinity of Arco, ranging from about 2 to 6 feet below the land surface, the highest water level of record occurred in July and August 1951, shortly after unusually heavy rains during those months. The high levels where the water table is deeper (18 to 55 ft below the land surface) were not reached until July and September 1952. In the regional water table (205 to 400 feet below the land surface) water levels remained fairly steady and showed little or no effect of the heavy runoff in 1951 and 1952.

In T. 3 N., Rs. 31-33 E. the yearly range is about 1.5 to 2 ft, with the higher levels about mid-January and the lower between mid-July and early August. In Tps. 2 N. to 1 S., Rs. 27-31 E., the yearly range is about 1.5 ft. The high levels are reached between early November and mid-December, and the lows in May and July.

The greatest yearly range in fluctuations occurred in wells in T. 5 N., Rs. 33-34 E., where the slope of the water table was about 5 to 7 ft per mi. South of T. 5 N., in Rs. 27-34 E., net water-level changes averaged 1 to 2 ft and the slope of the water table was only 2 to 3 ft per mi. The seasonal pattern of fluctuations in the northeastern area probably is influenced largely by underflow from Mud Lake, the apparent direction of which is southwestward toward the wells. However, no definite time pattern is discernible in the occurrence of high and low levels at increasing distances from Mud Lake. Probably factors other than underflow from Mud Lake strongly affect water levels in that area.

Regional Water-Level Fluctuations in the Snake River Plain

The area here treated is the part of the Snake River Plain outside the NRTS and includes several large ground-water irrigation tracts. Regional water-level fluctuations have twofold interest: their records assist interpretation of fluctuations in the NRTS, and they register changes in ground-water storage, so they are significant in problems of ground-water depletion now and in the future. Figure 27 illustrated the principal types of water-level fluctuations in wells on the Snake River Plain.

Arco area.—The Arco area, at the mouth of the Big Lost River Valley in Tps. 4-6 N., Rs. 25-27 E., is about 20 mi northwest of the CF area and

Underflow from the Mud Lake basin is an important source of ground-water replenishment to the central Snake River Plain, but direct effects of underflow have not been identified in water-level fluctuations south and southwest of Mud Lake. The apparent lack of correlation in water-levels may be partly due to inadequate records, but the lack of strong fluctuations probably is normal. Owing to the relatively long distance of the wells from the main irrigated area around Mud Lake, the recharge wave moving southward is damped before reaching the wells. Also, the Mud Lake basin itself is such a large reservoir that seasonal variations in underflow probably are relatively small.

Wells in which the seasonal rise and fall of water levels is most marked are in the path of the apparent movement of water from Mud Lake.

A few deep wells tap water under local artesian or quasi-artesian pressure, which may cause some of the differences in the behavior of the water levels in some of the wells.

The greater range in water-level fluctuations occurs in areas having the greater slopes on the water table.

Seasonal Water-Level Fluctuations

A yearly cycle of water-level fluctuations, represented by a graph having approximately sinusoidal form, occurs in wells in Bingham and Jefferson Counties, in Tps. 2-6 N., Rs. 31-34 E., and in Butte County south of U. S. Highway 20, in Tps. 2 N., to 2 S., Rs. 27-31 E. (see fig. 28, well 5N-34E-9bd1). The yearly range of fluctuation in Tps. 5-6 N., Rs. 32-34 E. averages about 3 ft. The high water levels occur between mid-January and mid-March, the lows between mid-August and mid-September.

rise of the water level in no. 18 in 1953 was less than half that in no. 12. The total rise of the water level in the MTR test well (3N-29E-14ad1) was about 4 ft in 1952 and 6 ft in 1953. In 1950-51, however, before the effects of recharge from the Big Lost River were registered, the annual water-level fluctuation in the well was about a foot. In well 3N-30E-31aa1 (field no. 20, about 3 mi southeast of the MTR site) the highest water level also was reached in mid-March, and the range in fluctuations was about the same as in the MTR test well. Water levels at the CF area in 1953-54 also were at their peak levels in the first half of March.

In general, rises of water level caused by underflow from the valleys of the Little Lost and Big Lost Rivers and Birch Creek began early in the spring of 1952, about 6 to 10 months after the unusually high runoff and recharge that began late in July and early in August 1951. Except for slight seasonal decline in several wells in 1952, water levels in all the wells rose steadily, reaching a peak in the spring of 1953. Water levels declined steadily after reaching the record high levels, and by the end of 1955 were about 2 ft lower than at the end of 1950.

Water levels in the western central part of the area, near the Little Lost River playa in T. 5 N., R. 29 E., reached peak levels early in April 1953; those in the intermediate area, near the Big Lost River playas in Tp. 4-5 N., Rs. 29-31 E., were at peaks in early and mid-May; those in the southern part, near the CF area and MTR site, were at peak levels in early and mid-March.

The greatest range in water-level fluctuations, and the most rapid seasonal rise and decline of water levels, occurs in wells near the Little Lost River playa south of the Village of Howe.

Well 5N-29E-23cd1 (field no. 19) is about 3 mi southeast of the Village of Howe and 1½ mi southeast of the Little Lost River playa. The water level in that well had the largest seasonal range of fluctuations noted in any well in the area. The range was about 8 ft in 1953 and 7 ft in 1954, with very distinct peaks about April 1 in both years, about a month earlier than the highest yearly levels in nearby wells. Water levels were at a record low in April 1955. Seepage losses from the Little Lost and Big Lost Rivers seemingly did not influence the water level in the well before the summer of 1952, several months after the water levels in other wells began to rise. Infiltration and percolation to the regional water table of ponded water from the Little Lost and Big Lost River playas, north and northeast of the well may have caused the unusually high water levels in 1953 and 1954.

In well 4N-29E-9dc1 (field no. 23), which is about 4 mi south of field no. 19, the water level reached peaks in May 1953 and 1954. The maximum rises during the period were 0.5 ft to 1.5 ft. The highest water level in well 4N-30E-7ad1 (field no. 12), about 5 mi southeast of field no. 19, occurred in May 1953 and 1954, about a month after the peak in field no. 19. The total rise of the water level in no. 12 was about 2 ft in 1953 and about a foot in 1954. In well 4N-30E-22bd1 (field no. 17), about 7 mi southeast of no. 19 and 3 mi southeast of no. 12, the highest water level in 1953 was reached early in May, and the range in fluctuation was about half that in no. 12. Well 5N-31E-14bc1 (field no. 18) is the well farthest north and east that shows effects of recharge from Birch Creek and the Big Lost River. The water level reached a peak in mid-May 1953, about concurrently with the peaks in wells 12 and 23. The total production wells near the river. For well 2N-31E-35dc1, near Atomic City and about 11 mi from the river, the hydrograph in 1952 repeated very closely the seasonal pattern of 1951.

Wells in T. 2 N., Rs. 27-28 E., 1 to 4 mi south of the river, reached peaks in November and December 1953 and perhaps were affected strongly by recharge from the river. Records are not available for those wells earlier than 1952 and there is no basis for comparison. Noticeable affects of seepage from the river seemingly did not extend to wells about 10 mi to the southeast, in Tps. 2-3 N., R. 31 E., where daily records for several wells show that peak levels were reached at about the same time each year during 1950-55.

The effect of recharge from the Big Lost River was registered earliest in wells at the MTR and CFA sites. The highest water levels in the MTR test well and in well 3N-30E-31aa1 (field no. 20) occurred during early and mid-March, about 2 to 4 weeks earlier than in other wells affected by the river.

The larger rises in water levels (7 to 10 ft) occurred in wells north of the STR site in Tps. 4-5 N., Rs. 29-31 E. The general slope of the water table in that area was about 10 to 12 ft per mi. Near the MTR and CPP sites the water levels rose about 6 ft; the slope of the water table ranged from about 5 to 7 ft per mi.

Recharge from local sources.--In wells in the western-central part of the Station, in Tps. 3-5 N., Rs. 29-31 E., water-level fluctuations follow a fairly well-defined seasonal pattern that is influenced both by local and by regional recharge (see fig. 28, well 3N-29E-14ad1).

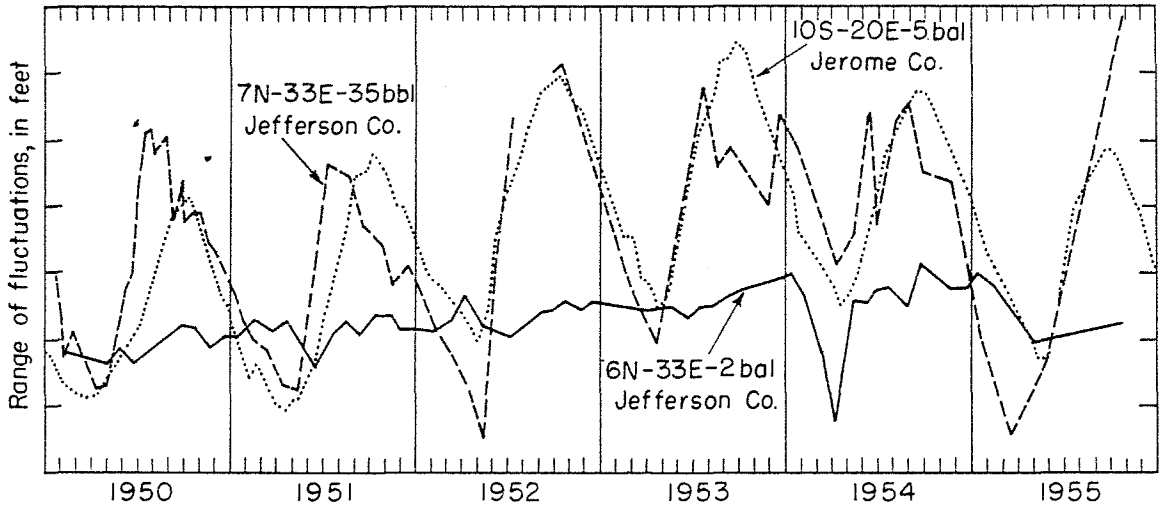


Figure 27.—Three types of water-level fluctuations in wells on the Snake River Plain: (1) relatively large annual range, with superimposed short-term fluctuations and longer-term trends in well 7N-33E-35bbl, Jefferson Co.; (2) somewhat irregular short-term fluctuations and longer term trends in well 6N-33E-2bal, Jefferson Co.; (3) small short-term fluctuations, relatively large annual range, and pronounced longer-term trend in well IOS-20E-5bal, Minidoka Co.

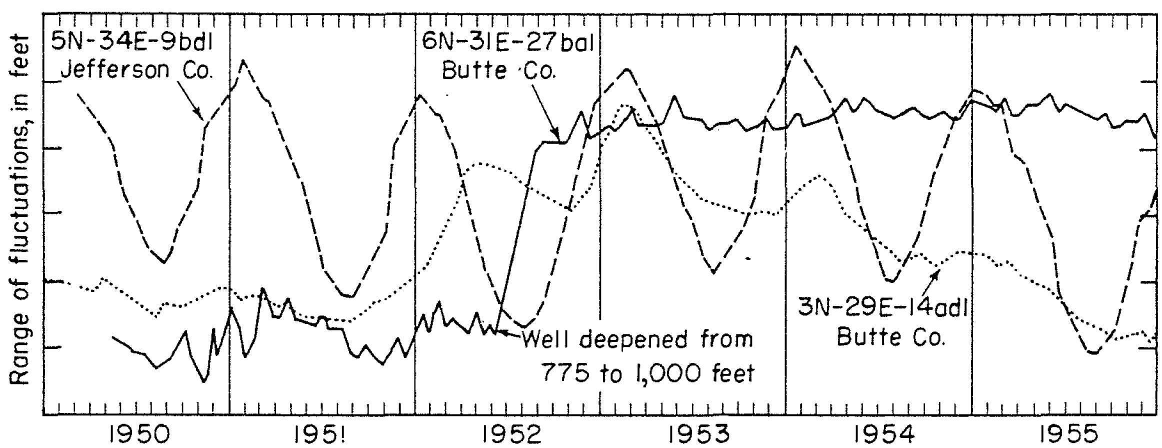


Figure 28.—Types of water-level fluctuations in wells in and near the NRTS: (1) small seasonal and annual range in well 6N-31E-27bal, Butte Co.; (2) small normal seasonal and annual range in well 3N-29E-14adl, Butte Co., modified by effects of unusual recharge from Big Lost River from late 1951 to early 1954; (3) relatively large annual range and cyclic trends in well 5N-34E-9bdl, Jefferson Co.

rose about 3 ft. Thereafter, the water level in the deepened well remained about 3 ft above the old general level, but the annual range of fluctuation after 1952 was only about 0.5 ft. The yearly range in well 6N-33E-2ba1, about 8 mi northeast of the NRTS, also was about 0.5 ft.

In the northern part of the NRTS the water table seems to be nearly flat, and at places the slope is less than a foot per mile. Pumping from the ANP wells has had no noticeable regional effect on water levels.

Fluctuations Caused by Local Recharge

Recharge from the Big Lost River.--The effects of local recharge by percolation from the Big Lost River during the high runoff period 1951-53 were considered briefly on p. 61-65, in connection with the rate of percolation through the basalt. Hydrographs show that, beginning in September 1952, a change occurred in the seasonal pattern of water-level fluctuations near the river, probably as a result of exceptional recharge. Some of the recharge may have been from deep percolation out of the river. In well 3N-29E-14ad1 (see fig. 28), a mile from the river, the seasonal trends in 1950 and 1951 were entirely different from those in 1952-55. About September 15, 1951 the water level in well -14ad1 began a general rise that continued through April 1952, when the water table reached its highest level during the year. Except for a slight decline from June through October 1952, the water table rose steadily and reached its highest level of record in March 1953. The beginning of the general rise lagged about five weeks behind the arrival of surface water in the Big Lost River channel and the beginning of seepage losses in the channel reach nearest to the well. Similar rising trends occurred in several

Fluctuations Caused by Changes in Storage

Ground-water fluctuations caused by changes in storage follow several patterns in the NRTS and vicinity. Within a single pattern-group the magnitude of fluctuations and the time of occurrence of the high and low water levels vary. The groups differ from each other in the magnitude and trend of fluctuations. The following grouping is made for convenience in discussion: (1) fluctuations in which the yearly range is small; (2) fluctuations which are affected markedly by local recharge; (3) fluctuations which follow a well-defined seasonal pattern. The types overlap to some extent and various combinations can be found. Figures 27 and 28 illustrate several types of water-level fluctuations.

Yearly Fluctuations in Small Range

The water table in the northern NRTS, in T. 6 N., Rs 31-33 E. ordinarily fluctuates only through a small range, with little or no cyclic yearly pattern (see fig. 28, well 6N-31E-27ba1). A continuous water-level record for the period 1950-55 is available for well 6N-31E-27ba1. A record nearly as long, represented by weekly and monthly measurements, is available for a well in T. 6 N., R. 33 E., showing similar hydrographic characteristics. Water-level records for other wells in that general area average only about three years in length and are too short to be classified readily. During the period May 1950 to June 1952 the range in yearly water-level fluctuations in well 6N-31E-27ba1 was about a foot. The depth of the well was 775 ft and the depth to water was about 214 ft below the land surface. In 1952 the well was deepened to 1,000 ft and the water level

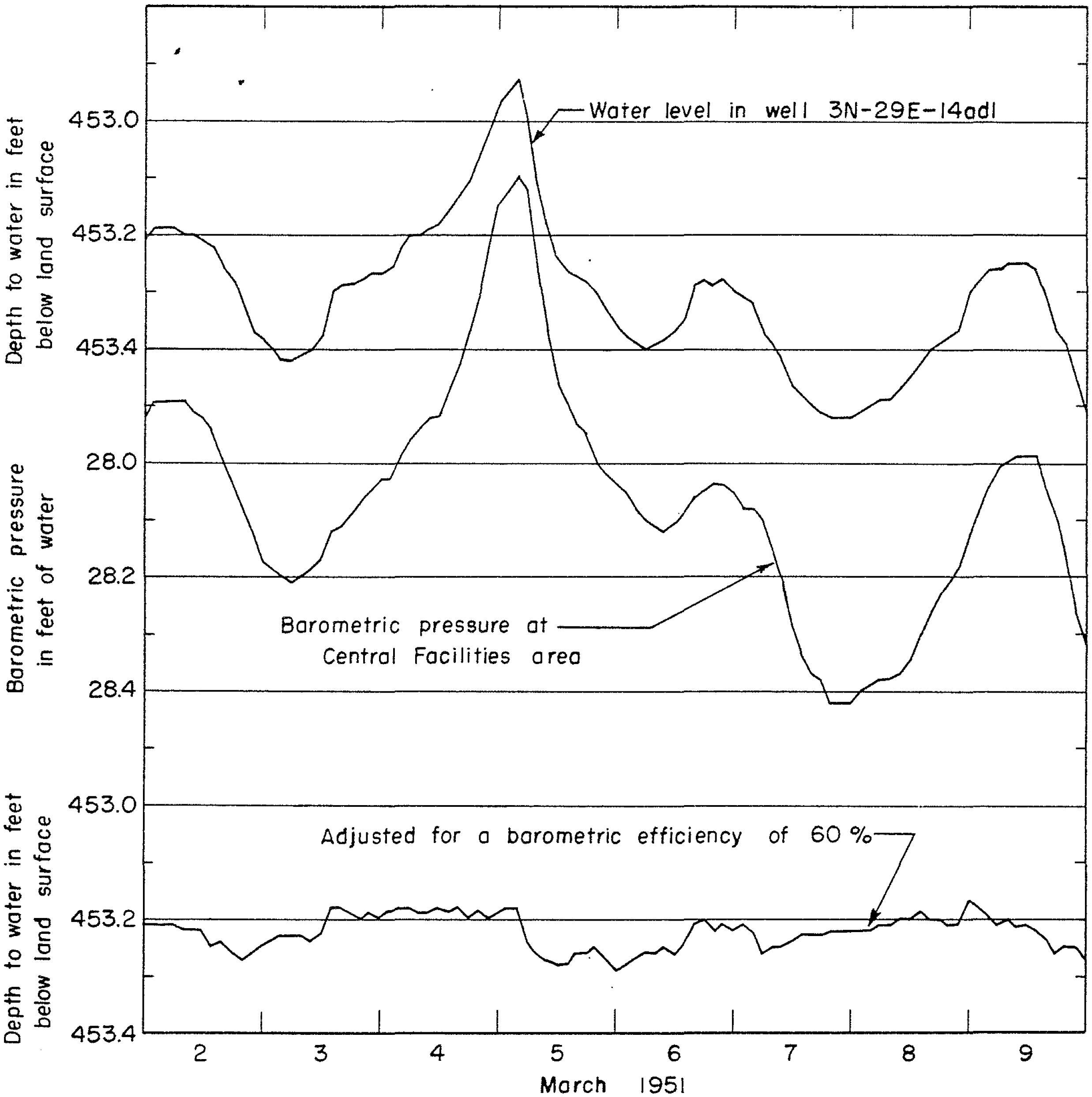


Figure 26.—Hydrograph of well 3N-29E-14adl, unadjusted (above) and adjusted for barometric pressure (below).

GPO 991857

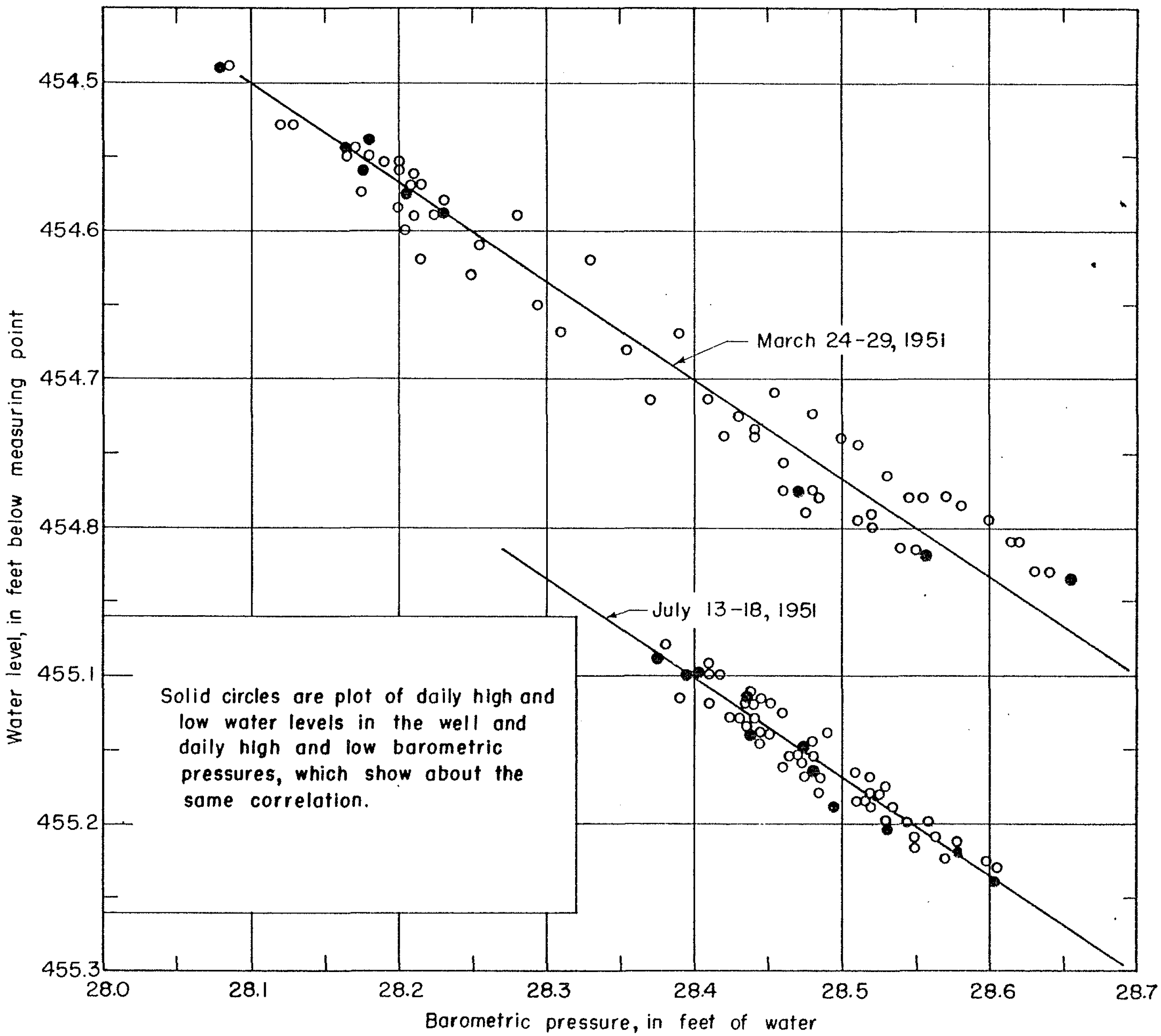


Figure 25.— Correlation of barometric pressure with water levels at two-hour intervals in well 3N-29E-14adl.

figure 26 is based on a computed barometric efficiency of 60 pct. Though the adjustment is quite small, it is an appreciable percentage of the total range of short-term fluctuations and would noticeably affect calculations based on those fluctuations. Nonbarometric factors obviously affected the water-level, because the adjustments leave considerable irregularity in the hydrograph. A correlation of water-level fluctuations in the well with increasing and decreasing barometric pressure shows a slight change in the efficiency of the well during these periods. So part of the irregularity is due to differences in the barometric efficiency of the well during periods of rising and falling pressure.

Significance of barometric fluctuations.--Barometric fluctuations of water levels in wells are caused by fluctuations of atmospheric pressure on the water surface in the well. In ordinary porous geologic material which is permeable throughout and is in free communication with the atmosphere, air moves in and out of the ground freely through the land-surface area. Where the zone of aeration contains nonpermeable layers, these restrict the free exchange of air between the atmosphere and the ground, and pressures may be equalized only by movement of air through local permeable zones or through openings such as wells which are not cased to the water table. Wells in such materials characteristically "blow and suck." In the case of the Snake River basalt, barometric fluctuations of water levels show that tight material occurs between the land surface and the water table and impedes the movement of air, which can escape from or enter the ground only through local fractures, caverns, and wells. A very large volume of air is moved, and circulation of air through wells and caverns often persists in one direction for some hours.

The time scale of the FM 32-day recording gage is so condensed that it is difficult to correlate the trace of water-level fluctuations with hourly barometric-pressure readings. The barometric efficiencies of wells equipped with that type of recorder were computed by correlating daily high and low water levels with daily minimum and maximum barometric pressures. In general, the results were fairly consistent for different seasons of the year. Figure 25 is a scatter diagram of barometric pressure and water levels at two-hour intervals in well 3N-29E-14ad1 during March 24-29, 1951 and July 13-18, 1951. For comparison, the daily high and low water levels in the well and the daily high and low barometric pressures are shown by solid circles. These are the only plottings that could be made from a 32-day chart, but a straight line through those points would not differ appreciably from that drawn through all the points in the figures.

Plottings for longer periods up to one month confirm the short-term correlations. That is, the width of the scatter band for summer periods is narrow and that for winter is somewhat broader, but the slopes of the lines are about the same.

The barometric efficiency of a well is a function of the rate of change of the barometric pressure. Though the water level in a well may follow the diurnal fluctuations rather closely, the longer, slower barometric changes are not recorded with the same clarity and some are not identifiable.

Using a microbarogram of atmospheric pressure, and applying the barometric-efficiency factor to the water-level fluctuations in a well, the hydrograph for the well can be adjusted. The adjusted hydrograph in

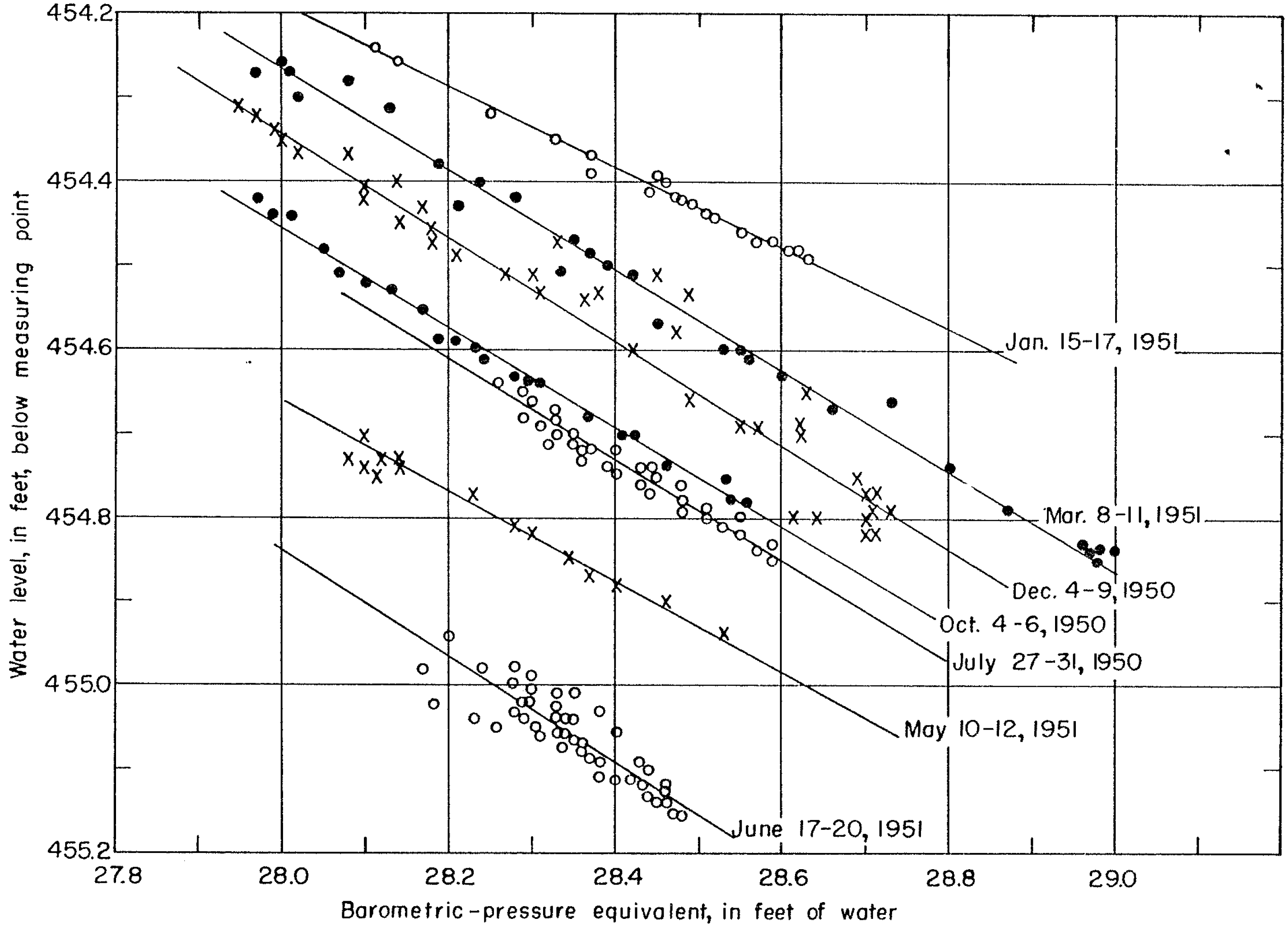


Figure 24.–Correlation of water-level fluctuations in well 3N-29E-14ad1 with barometric pressure.

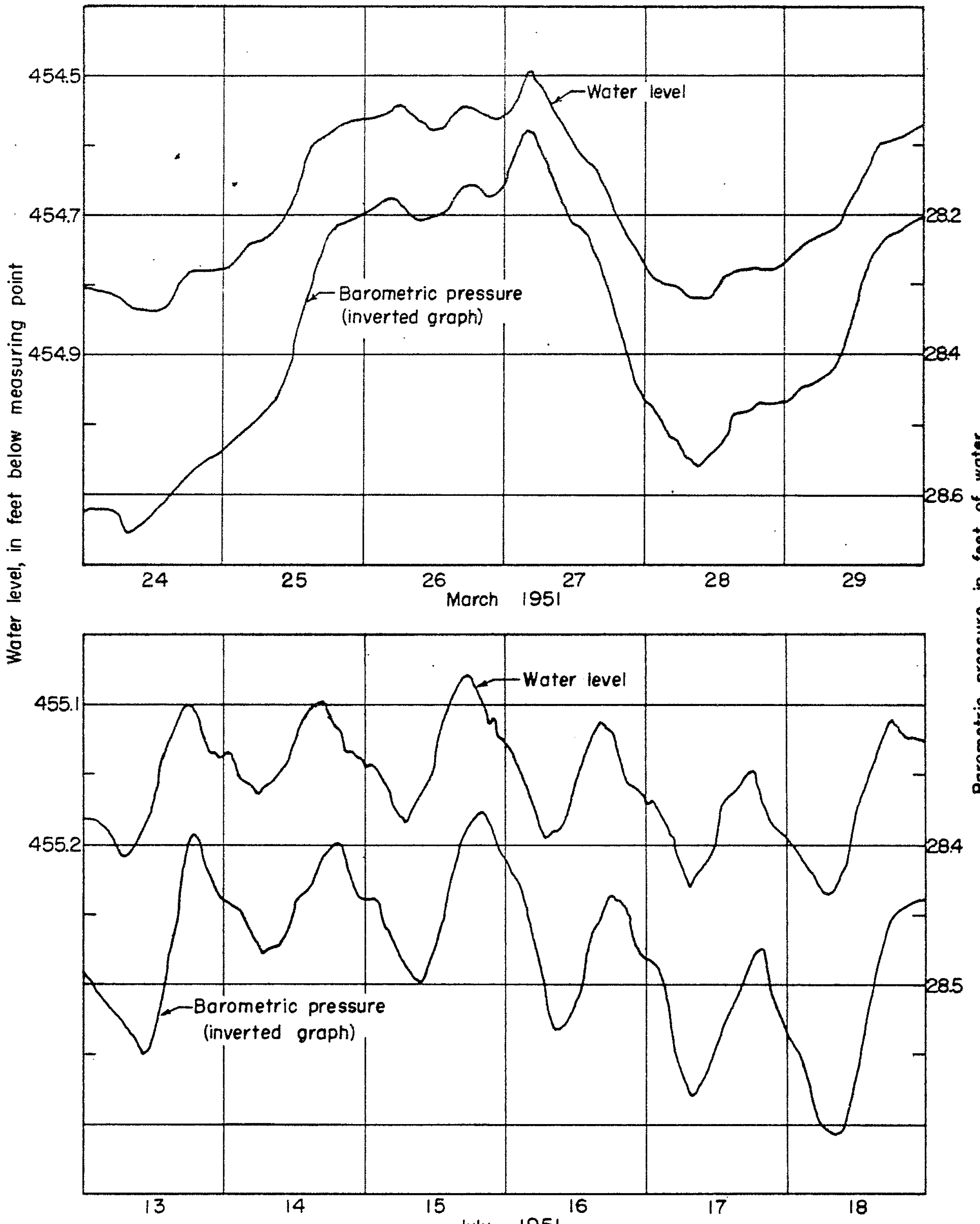


Figure 23.—Barometric water-level fluctuations in well 3N-29E-14adl. Compares typical winter (above) and summer (below) with contemporary barometric-pressure changes. (Microbarographic record from U.S. Weather Bureau.)

GPO 991857

The diurnal pattern of fluctuations is predominant during May through September, with distinct highs and lows during each 24-hour period (fig. 23, lower hydrograph). The low occurs between 6:00 and 8:00 am, and the high between 6:00 and 8:00 pm. The daily fluctuations in summer range from about 0.03 to 0.17 ft and average about 0.1 ft.

Superimposed on the diurnal cycle are longer, irregular fluctuations caused by cyclonic storms moving eastward across the western United States. In winter the storms are irregular and the barometric pressure fluctuates widely. In summer these storms are more regular and fluctuations are small. The cyclonic-pressure changes are most prominent from about mid-September to May and cause relatively large water-level fluctuations (fig. 23, upper hydrograph). The periods of increasing and decreasing pressure usually extend through one or more days and caused water-level fluctuations of about 0.2 to 0.8 ft. The average is about 0.3 ft.

As a result of the seasonal differences in the barometric-pressure patterns, the water-level patterns caused by diurnal fluctuations are dominant in late spring and summer, and those caused by cyclonic storms are dominant in the fall and winter.

Figure 24 is a series of scatter diagrams in which the depth to water in well 3N-29E-14ad1 during several periods is plotted against the contemporary barometric pressure at the Weather Bureau Station in the Central Facilities area. The diagrams indicate exceptionally good correlation between the two variables, and the barometric efficiency of the well averages about 60 pct. The barometric efficiency of other wells on and near the NRTS ranged from about 32 to 95 pct (see table 23).

Table 23.—Barometric efficiency of wells on and near the NRTS

Well number	Barometric efficiency (percent)	Well number	Barometric efficiency (percent)
7N-31E-34ba1	53	4N-30E- 7ad1	45
6N-31E-13db1	86	3N-29E-14ad1	3/60
6N-31E-27ba1	65	3N-30E-31aa1	2/95
6N-32E-11ab1	52	3N-32E-29dd1	45
6N-32E-36ad1	87	3N-33E- 3ab1	88
6N-33E-26dd1	89	3N-29E-25bd1	77
5N-29E-23cd1	1/41	2N-27E- 2dd1	52
5N-31E-14bc1	2/95	2N-28E-35ad1	65
5N-32E-36ad1	91	2N-31E-35dc1	51
5N-34E- 9bd1	64	1S-30E-15bc1	32
4N-29E- 9dc1	28	Average	60

1/ Large nonbarometric changes in water levels make determination of barometric effects difficult.

2/ Apparent efficiency unusually high. Other factors may be involved.

3/ Apparent efficiency ranges between 56 and 67 percent at different times.

In general, barometric fluctuations in wells on the NRTS follow two distinct seasonal patterns. One is a typical diurnal cycle induced by solar heating of the atmosphere; the other is a less systematic fluctuation caused by cyclonic storms (fig. 23). The pressure changes caused by solar heating during the day and by cooling during the night cause regular diurnal low and high pressures throughout the summer. Water levels in wells respond by rising during the day and declining during the night.

In a truly unconfined aquifer the water table communicates freely with the atmosphere through ground air in the zone of aeration. Inasmuch as the barometric pressure affects all parts of the water table about equally and about the same as they affect the water level in the well, the water level does not change with changing pressure.

Barometric fluctuations of water levels commonly occur in wells on the Snake River Plain because the water is partly confined, especially at depth. Impermeable layers in the basalt, both in the zone of saturation and in the zone of aeration, locally confine the ground water to some extent and they impede communication between the atmosphere and ground air (see discussion of ground air, part 2, p. 43-46). In most wells in the Snake River basalt that have been observed systematically, water levels respond to changes in barometric pressure.

The confining bed in an artesian system resists the effect of rising atmospheric pressure. The decline of the water level in the well is a function of the proportion of barometric pressure that does not overcome the resistance of the confining bed. That is, the well is not 100-pct efficient as a water barometer. The barometric efficiency is expressed as a percentage ratio of water-level change in the well to the barometric change, in equivalent units. A barometric efficiency of 60 pct means that the water level in a well rises 0.6 ft in response to a barometric pressure drop equivalent to a head of one foot of water. The computed barometric efficiency of representative wells on the NRTS is shown in table 23.

Barometric Fluctuations

Prominent and regular short-term water-level fluctuations in wells on the NRTS are caused by changes in atmospheric pressure. These are called barometric fluctuations because the wells act as water barometers. Barometric fluctuations do not indicate changes in ground-water storage, but they disclose certain characteristics of the aquifer. At times they mask other water-level fluctuations. Therefore, microbarograms of atmospheric pressure are used in conjunction with records of barometric water-level fluctuations to interpret the hydrographs of wells whenever accurate reading of small fluctuations is necessary. The barometric fluctuations are especially important during pumping tests when small drawdown in a well may be masked completely by barometric fluctuations. Observed water-level fluctuations can be adjusted for barometric effects and this is necessary before making computations based on effects caused by pumping.

Mechanism of barometric fluctuations.--Barometric fluctuations, like seismic ones, ordinarily are more prominent in artesian wells than in wells that tap unconfined water. The water level in an artesian well expresses a balance between the internal pressure in the aquifer and atmospheric pressure on the water surface in the well. The water level changes when the two pressures become unbalanced. Artesian aquifers are elastic (Meinzer, 1928, p. 263-291). When pressure in the aquifer increases, the aquifer may expand and compress the overlying confining bed (the rise and expansion, of course, are extremely small per unit volume). Pressure increases when barometric pressure rises and water is forced from the well into the aquifer. The reverse occurs when atmospheric pressure falls.

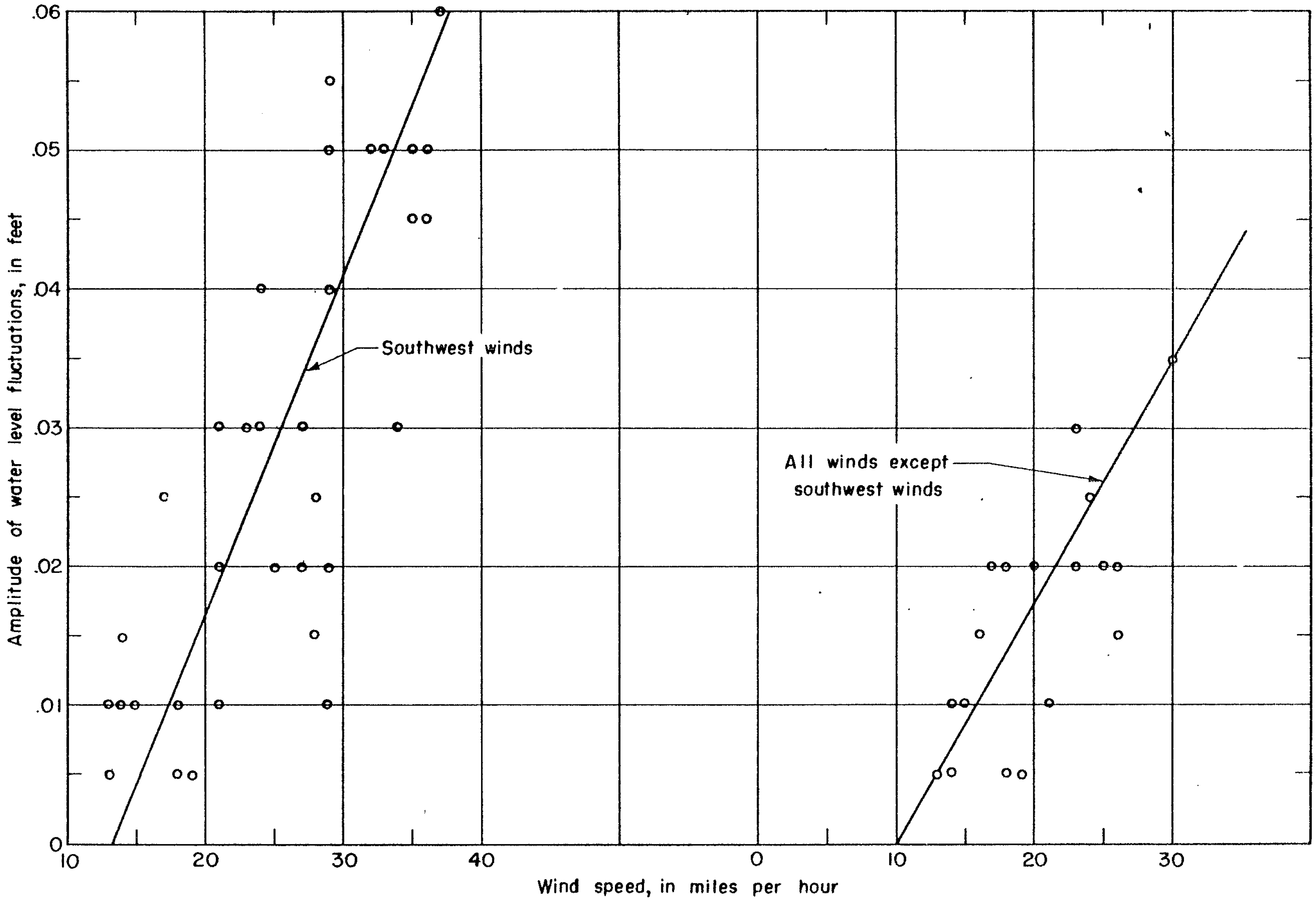


Figure 22.—Effect of wind speed on water-level fluctuations in well 3N-29E-14adl, Sept. 28 to Oct. 2, and Nov. 22-23, 1953.

Figure 22 shows, in addition to the high incidence of southwest winds, the greater range in water-level fluctuations induced by those winds compared to winds from other quadrants. In general, a southwest wind of 12 mph noticeably affects the water level in a well, whereas a wind speed of only about 10 mph from other quadrants is sufficient to cause noticeable fluctuations. The difference probably is due largely to the fact that the southwest winds usually are less gusty than other winds and tend to produce more gradual pressure changes in wells. An increase in speed of 10 mph in a southwest wind causes a water-level fluctuation of about 0.024 ft. whereas a similar increase in wind speed from any other quadrants causes a water-level fluctuation of about 0.017 ft.

Significance of eolian fluctuations.—Although wind-generated water-level fluctuations in wells are small to negligible, the fluctuations often interfere with the interpretation of recorder charts because they mask other fluctuations. This is especially true during pumping tests, when small water-level changes caused by pumping must be identified in order to compute the hydraulic coefficients of the aquifer. Wind-generated fluctuations are less apt to occur in water table wells than in artesian wells. Their occurrence in wells on the NRTS shows that there is poor communication between ground air and the atmosphere, so that air pressure is not applied uniformly to all parts of the water table.

in response to a sharp increase in atmospheric pressure. Figure 21 shows typical periods of wind-generated fluctuations, and intervening periods of relative calm.

Mechanism of wind-generated fluctuations.—Wind-generated fluctuations are caused by changes in air pressure within the recorder shelters and in the well casings. Air-pressure changes inside the shelter, caused by fluctuations in wind speed, are transmitted directly down the well to the water surface. A gust of wind blowing past a recorder shelter increases the velocity and decreases the pressure of the moving air at the ends of the shelter. The net effect is lowered air pressure inside the shelter and the well casing and a rise of the water level in the well. As the gust of wind dies, the pressure in the shelter and well casing returns to normal and the water level declines. Because wind speed and direction never are constant, pressure in the well casing fluctuates continually. Peak gusts, which produce relatively large and very rapid pressure changes within the shelter and well casing, cause the irregular and jagged traces on the charts.

A wind blowing normal to the axis of the ventilator openings in a recorder shelter causes the greatest water-level fluctuations in the wells. For example, the vents in the shelter on well 3N-29E-14ad1 open to the northwest and southeast, and the water level in the well responds more strongly to winds from the northeast and southwest quadrants than to those from other quadrants. On the left side of figure 22 is a plot of hourly southwest wind speeds against the amplitude of the induced water-level fluctuations. On the right side of the figure the fluctuations in response to all winds except those from the southwest are shown.

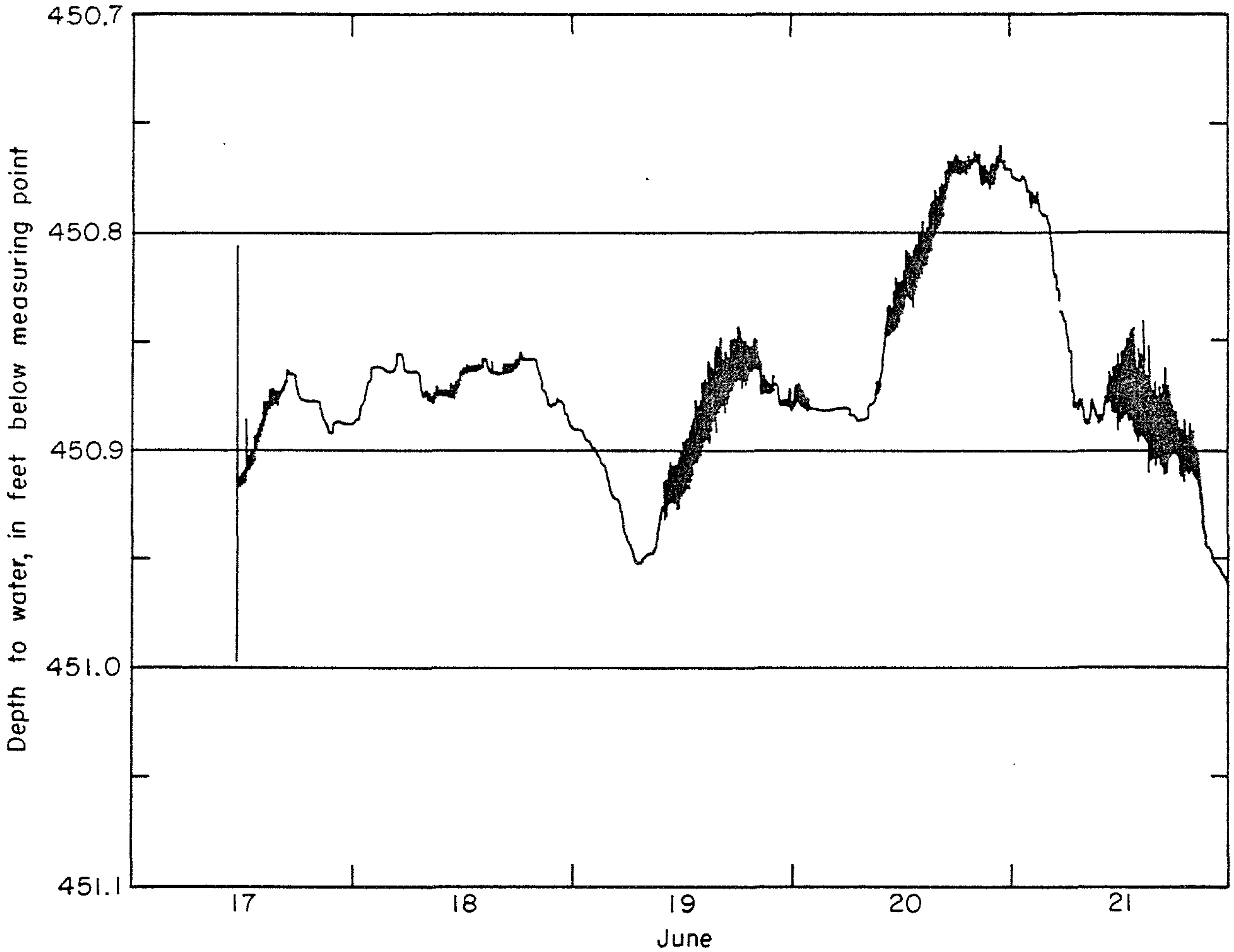


Figure 21.— Wind-generated fluctuations of water level in 1953 in well 3N-29E-14adl, and intervening periods of relative calm with very little fluctuation (reproduction of actual trace on recorder chart).

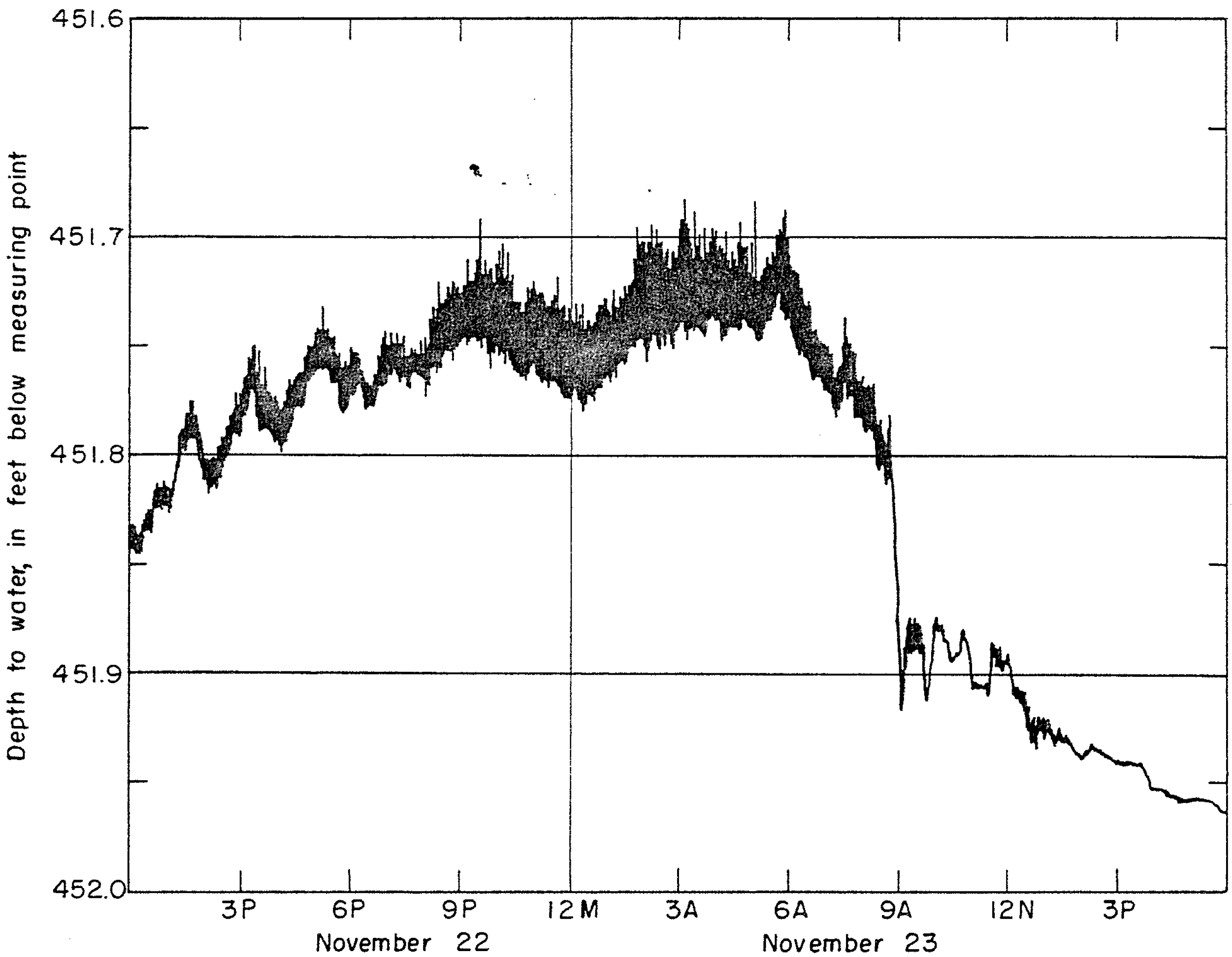


Figure 20.- Wind-generated fluctuations of water level in 1953 in well 3N-29E-14ad1 (reproduction of actual trace on recorder chart).

the well, and the water-level rises. The rise of water level is strictly a well phenomena; nothing but compression and rise of pressure occurs in the aquifer.

Wells in the NRTS respond readily to seismic waves because the aquifer is a complex interlayering of permeable and nonpermeable basalt and interflow sediments. The nonpermeable beds locally restrict communication between successive vertical levels in the aquifer. Water between such layers is subjected to confined compressive force when the seismic waves strike, and responds like artesian water. It is statistically probable that any deep well in the basalt penetrates one or more nonpermeable layers at or beneath the water table and is able to register seismic waves.

Fluctuations Caused by Wind

Wind-generated fluctuations, which are recognizable on most recorder charts from wells on the NRTS, are momentary and do not indicate changes in ground-water storage. Wind effects obscure the record of other water-level changes and lessen the accuracy of water-level readings from a recorder chart. The rapid eolian oscillations also reduce the accuracy of direct water-level measurements. The small fluctuations caused by wind are a minor concern, however, and readings of water level are sufficiently accurate for most purposes.

The amplitude of the observed water-level fluctuations caused by wind averages between 0.01 and 0.02 ft; few exceeded 0.06 ft. Figure 20 is a direct reproduction of the recorder trace of the water level in a well while the wind speed ranged from 6 to 37 mph and averaged about 25 mph. The rapid decline of the water level at about 9:00 am on November 23 was

550 mi northwest of the Humboldt County epicenter, caused a fluctuation of 0.29 ft in the same well. Even successive quakes from the same focus and comparable magnitude do not have uniform effects. The Fallon, Nevada quake of July 6, 1954, magnitude 7, caused a water-level fluctuation of 0.4 foot in well 4N-30E-7ad1, whereas the Fallon quake of August 24, 1954, magnitude 6.8, caused a water-level fluctuation double that of the earlier quake. On the other hand, in some instances there is a direct apparent correlation between the magnitude of the quake, its distance from the well and the amplitude of the water-level fluctuations. For example, an earthquake on Jan. 11, 1953, magnitude 6-1/2, in Yukon, Canada, with an epicenter about 1,650 mi distant, caused a water-level fluctuation of 0.09 ft in the well. The average of the fluctuations caused by the two California quakes is 2.5 times that caused by the Yukon quake, the distance to whose epicenter was 2.5 times that of the California quake.

The surface and transverse waves of earthquakes do not noticeably affect water level in wells. The observed seismic fluctuations are solely in response to the compressional waves, which are analogous to sound waves. Local geologic conditions in the aquifer determine its response to a seismic shock and the efficiency with which the response is transmitted through the aquifer. Water in unconfined aquifers ordinarily does not respond perceptibly to seismic waves because the waves travel very fast, affecting all parts of the water table about simultaneously and about equally, and the compression does not cause water levels to rise in wells. Many artesian aquifers on the other hand, are quite sensitive to seismic shocks because the aquifers are elastic. The compressional wave increases the pressure on the aquifer and forces some water out of it into

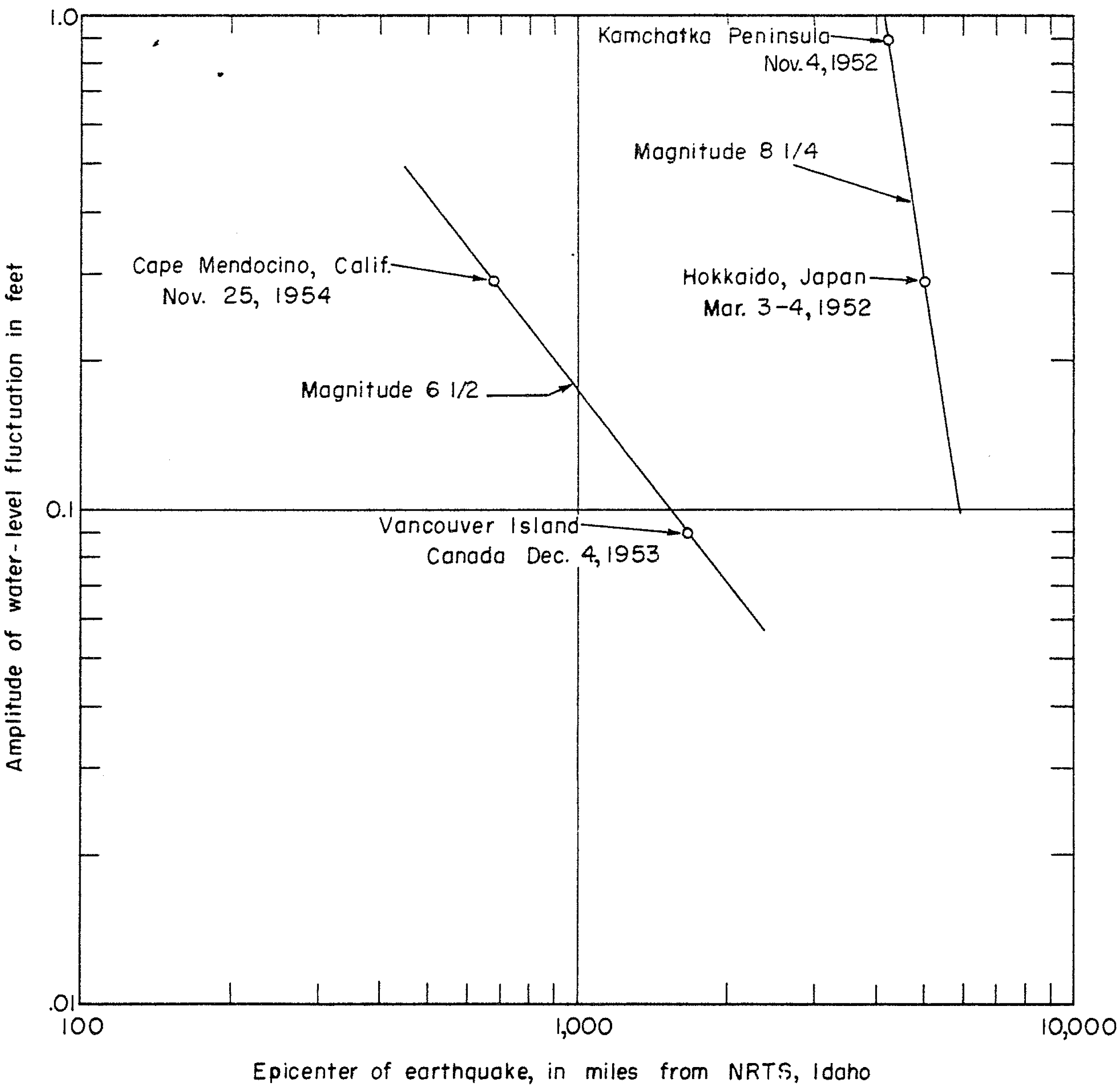


Figure 19.—Plot of amplitude of seismic fluctuations in well 4N-30E-7adl against distance from earthquake focus.

available. The measured depth was 292 feet in 1949, but in 1950 it was deepened to 310 ft. The chief aquifer probably is sand.

Significance of seismic fluctuations.—The seismic fluctuations of water levels at the NRTS show only minor local disturbances. Most of the shocks registered originated at foci hundreds to thousands of miles from the NRTS.

Study discloses a general relation between earthquake magnitude, distance of the epicenter from the wells, and amplitude of induced water-level fluctuations. Figure 19 shows the slopes of the lines obtained by plotting, for earthquakes of equal magnitude, the amplitude of the water-level fluctuations in well 4N-30E-7adl against the arc distance of the quake loci, in miles from the NRTS. Similar plots for other wells in the Station and in the Minidoka North Side Reclamation Project show about the same general degree of divergence between the lines. The steeper slope is obtained by the plot for the Japanese and Siberian earthquakes.

The amplitude of seismic water-level fluctuations in wells probably is not a simple function of the magnitude of a quake and the distance of its epicenter from the wells. The depth of the earthquake focus, the type of earth movement that caused the quake, local and regional geologic conditions, the paths of the seismic waves, and other factors must exert appreciable influence. For example, the earthquake on July 29, 1952, magnitude 6-1/2, had its epicenter in Humboldt County, California about 675 mi southwest of the NRTS. This quake caused a water-level fluctuation in well 4N-30E-7adl having an amplitude of 0.15 ft. An earthquake of similar magnitude on Dec. 25, 1954, whose epicenter was off the coast of Cape Mendocino, California about 675 mi southwest of the NRTS but about

The epicenter of the Tulare Valley, California earthquake, magnitude 7-1/2, was about 600 mi southwest of the NRTS and was recorded by 7 of 10 gages in operation on the Station at the time of the quake. The amplitude of water-level fluctuations in the wells ranged from 0.18 ft to 1 ft. The quake of November 4, 1952 on Kamchatka, Siberia, magnitude 8-1/4, whose epicenter was about 4,100 mi northwest of the NRTS, was recorded by 5 of the 10 gages. The water-level fluctuations in the wells ranged from about 0.04 to 0.09 ft. The greatest observed effects of seismic shocks were those caused by the Fallon, Nevada quakes, amplitude 6.8 to 7, of July 6 and August 24, 1954, whose epicenter was about 400 mi southwest of the NRTS. The first quake was recorded on 9 of 16 gages then in operation and the second on 12 gages. The amplitude of the fluctuations ranged from 0.06 to 0.78 ft in July and 0.02 ft to 1 ft in August.

The quake of August 12, 1953 in Greece, whose epicenter was about 6,700 mi distant, was the most remote quake recorded at the NRTS. The effect of the quake was registered on only one gage, and the amplitude of the water-level fluctuation was 0.03 ft. The gage was equipped with a 10 in. float in a 12-in. casing. The shock probably was dampened out in wells of smaller diameter by the frictional resistance between the floats and well casings.

Seismic-shock traces on several recorder charts do not correlate with any earthquakes reported by seismological stations. The traces probably represent minor seismic disturbances at local foci.

Well 6N-32E-36ad1 (Second Owsley) is the only well on the NRTS on which a recording-gage was operated longer than a year without registering an earthquake shock. The well was drilled many years ago and no log is

Table 22.--Seismic water-level fluctuations in well 4N-30E-7ad1, National Reactor Testing Station, Idaho--Continued

	Water-level fluctuation in well		Epicenter of quake		
	Amplitude (feet)	Time (GCT)	Time (GCT)	Location	Magnitude (Pasadena scale)
Dec. 21, 1954	.55	14-19	19:56	41°N., 124°W., Humboldt County, California	6½-6 3/4
Feb. 27, 1955	.19	21-22	19:21	Off coast of Honshu, Japan	-
Mar. 1, 1955	.14	03-04	08:03	Off coast of Mexico	-
Mar. 17, 1955	.09	21-23	03:22	Near coast of Kamchatka	-
Apr. 4, 1955	.17	17-19	11:11	22°N., 121°E., near south coast of Formosa	6
June 14, 1955	.08	04-05	03:43	6°N., 78½°W., off coast of Colombia	-
Aug. 23, 1955	.05	09-11	09:53	Mindanoa, Philippine Islands	-
Sept. 8, 1955	.05	12-14	10:59	53½°N., 160°E., near coast of Kamchatka	-

a/ Aftershock

b/ Pen trace on recorder chart indistinct.

Table 22.--Seismic water-level fluctuations in well 4N-30E-7ad1, National Reactor Testing Station, Idaho.

(Quake epicenters and magnitudes from published records of the U. S. Coast and Geodetic Survey)

Date	Water-level fluctuation in well: Amplitude (feet)	Water-level fluctuation in well: Time (GCT)	Epicenter of quake: Time (GCT)	Epicenter of quake: Location	Epicenter of quake: Magnitude (Pasadena scale)
Nov. 18, 1951	0.10	09-11	09:35	31°N., 90½°E., eastern Tibet	7½
Mar. 3-4, 1952	.29	24-02	01:23	42½°N., 143½°E., Hokkaido, Japan	8¼
Mar. 5, 1952	.06	15-17	15:46	24½°N., 108½°W., Gulf of California	5 3/4-6
July 29, 1952	.15	05-06	07:03	35°N., 119°W., Southern California	6½
Aug. 20, 1952	.35	15-17	15:25	43°N., 127°W., off coast of Oregon	7-7¼
Nov. 4, 1952	.89	16-18	16:58	52½°N., 159°E., east coast of Kamchatka	8¼
Jan. 11, 1953	.09	22-24	22:53	65°N., 133°W., Yukon, Canada	6½
Aug. 8, 1953	.07	18-21	18:40	52½°N., 159½°E., east coast of Kamchatka	-
Aug. 12, 1953	.03	08-10	09:24	38¼°N., 21°E., west coast of Greece	7¼
Aug. 17, 1953	a/ .04	02-08	02:12	38½°N., 21°E., west coast of Greece	-
Aug. 18, 1953	.06	03-05	04:37	- - , Aleutian Islands	-
Sept. 30, 1953	.18	20-23	23:04	22°N., 107½°W., off coast of Sinaloa, Mexico	6 3/4-7
Nov. 4, 1953	.04	01-04	03:49	12½°S., 166½°E., New Hebrides Islands	7.3
Nov. 17, 1953	.16	12-14	13:30	14°N., 92°W., near coast of Guatemala	7¼-7½
Dec. 4, 1953	.09	14-15	14:55	49½°N., 129°W., off coast of Vancouver Island	6½
Dec. 12, 1953	.08	16-18	17:31	3½°S., 81°W., off coast of Peru	7 3/4
Jan. 7, 1954	.04	04-07	04 & 07	- - , south of Fiji Islands	-
Mar. 19, 1954	.07	09-12	09:54	33.3°N., 116.1°W., Santa Rosa Mountains, Calif.	6-6¼
Apr. 29, 1954	.39	10-12	11:34	29½°N., 112½°W., Gulf of California	7½-7 3/4
July 6, 1954	.40	09-11	11:13	39½°N., 118½°W., near Fallon, Nevada	7
July 6, 1954	a/ .18	19-22	22:08	39½°N., 118½°W., near Fallon, Nevada	6 3/4-7
Aug. 24, 1954	b/ .86	05-07	05:51	39½°N., 118½°W., near Fallon, Nevada	6.8
Oct. 17, 1954	.10	22-24	22:57	31½°N., 116½°W., Lower California	5.8
Nov. 12, 1954	.06	09-12	12:26	31½°N., 116°W., Lower California	6.1
Nov. 25, 1954	.29	10-12	11:17	40½°N., 126°W., off Cape Mendocino, Calif.	6½
Dec. 16, 1954	>1.0	09-11	11:07	39½°N., 119°W., near Fallon, Nevada	7-7¼

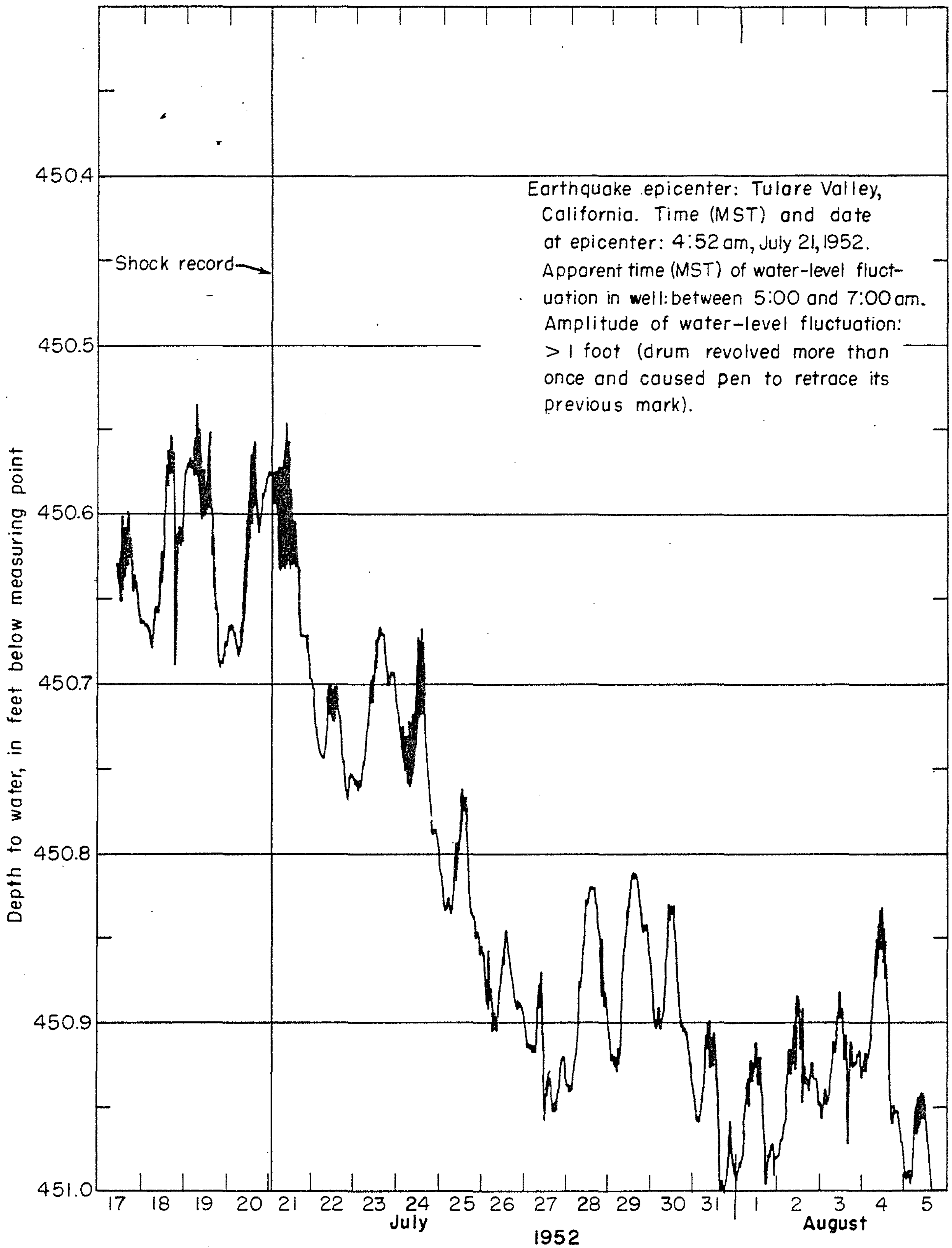
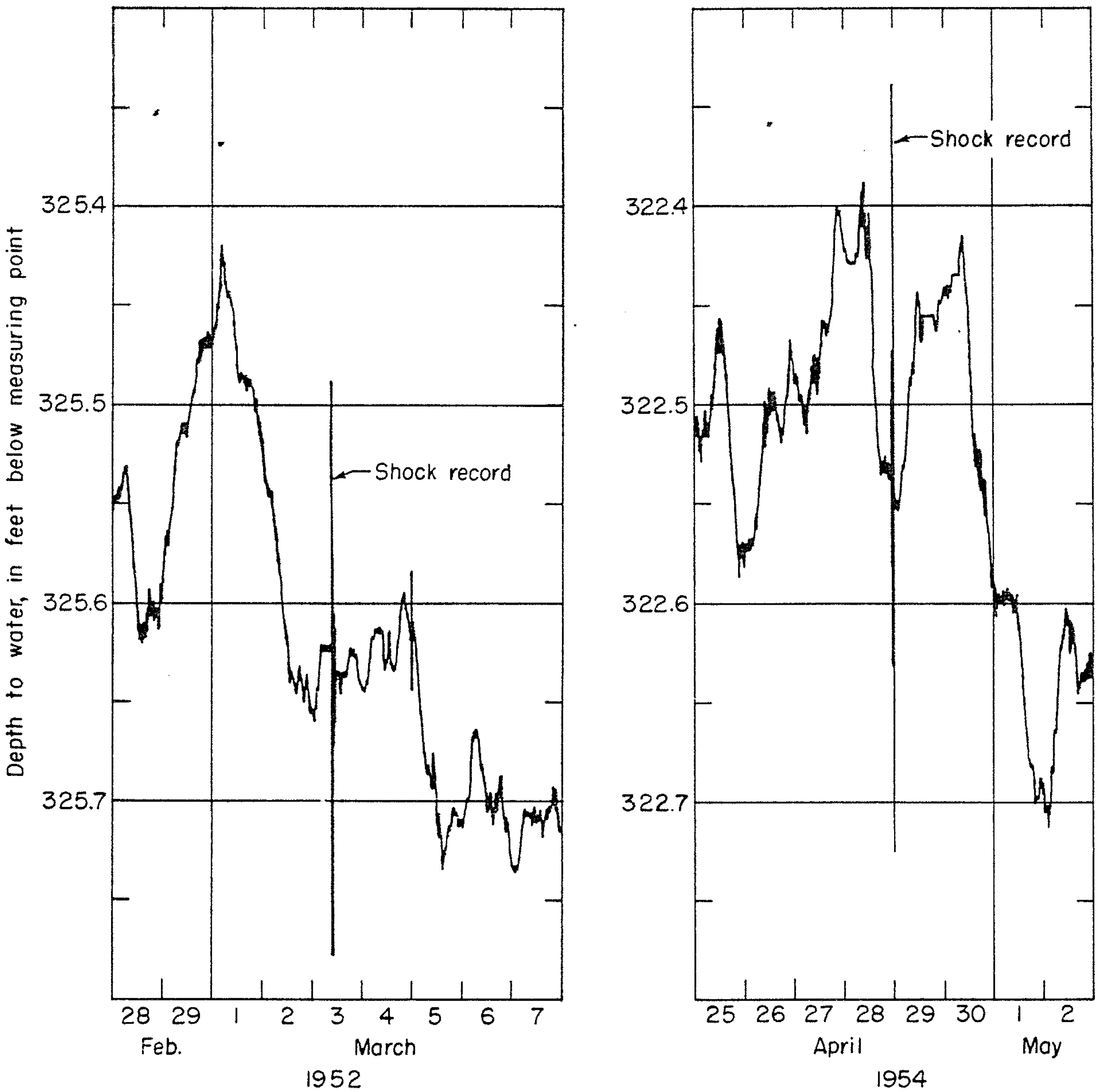


Figure 18.–Hydrograph of well 3N-29E-14ad1 showing water-level fluctuation caused by a seismic shock.

G 91857



A. Earthquake epicenter: Hokkaido, Japan. Total water-level fluctuation: 0.29 foot.

B. Earthquake epicenter: Gulf of California. Total water-level fluctuation: 0.39 foot.

Figure 17.—Records of seismic shocks in hydrograph of well 4N-30E-7adl for parts of 1952 and 1954, showing water-level fluctuations.

apparent arrival time was one to two hours earlier than the shock at the epicenter. Generally, the tremors registered by wells on the plain can be correlated readily with shocks originating beneath known epicenters.

The amplitude of most seismic water-level fluctuations in the Snake River Plain ranged between 0.02 and 0.2 ft. The amplitude of the fluctuations caused by earthquakes originating near Fallon, Nevada, and Tulare Valley and Humboldt County, California exceeded 1 ft in seven wells on the NRTS. The Fallon quake of December 10, 1954 caused a fluctuation exceeding 2 ft in well 3N-29E-14ad1. The chart drum on the gages was so geared that one revolution of the drum (full width of chart) represents either 1 or 2 ft. Fluctuations exceeding 1 ft or 2 ft cause the drum to revolve more than once and the pen retraces its mark (see fig. 18). For that reason, the full amount of the water-level fluctuation could not be read from the charts. The magnitude of the quakes causing the fluctuations ranged from 5-3/4 to 9-1/4 (Pasadena scale). Table 22 lists all seismic shocks recorded by the gage in well 4N-30E-7ad1 from November 1951 to December 1955 and correlates them with earthquakes reported by the U. S. Coast and Geodetic Survey. Records of all identifiable seismic fluctuations in wells in the NRTS are in the files of the Geological Survey.

Very few seismic shocks were recorded in the NRTS in the early period of observations. The gages were activated at that time by floats hung from a single-strand of copper-coated steel wire. More flexible, geometrically wound steel cable was used later and increased the sensitivity of the gages, so that more shocks have been recorded since the change. Typical traces were obtained on recorder charts from well 4N-30E-7adl (fig. 17) for earthquakes having epicenters in the Gulf of California and at Hokkaido, Japan. Because of the short instrumental time scale, the charts do not give a detailed record of the shocks, and the exact time of arrival cannot be determined. It is not known whether the water level in the wells rose or declined first. In a few instances the fluctuation of the water level in a well was so great that the chart drum made one or more complete revolutions and the pen trace on the chart was repeated (fig. 18). The shock from the Tulare Valley, California earthquake of July 21, 1952 caused the chart drum to revolve more than once.

On most of the recording gages a time scale of 0.3 in. per day was used, and the smallest subdivision on the chart was 0.1 in., representing a period of 8 hrs. At times some gages were equipped with gears giving time scales of 1.2 and 2.4 in. per day, and the smallest divisions on these charts represent 2 and 1 hrs, respectively. In general, on charts from most gages the time could not be read closer than within two hours of the actual time. Some errors were introduced also by fast and slow clocks. Hence, it was not possible to determine accurately the arrival time of an earthquake shock, or to compute travel time from the epicenter on the basis of these recorder charts. In a few instances the

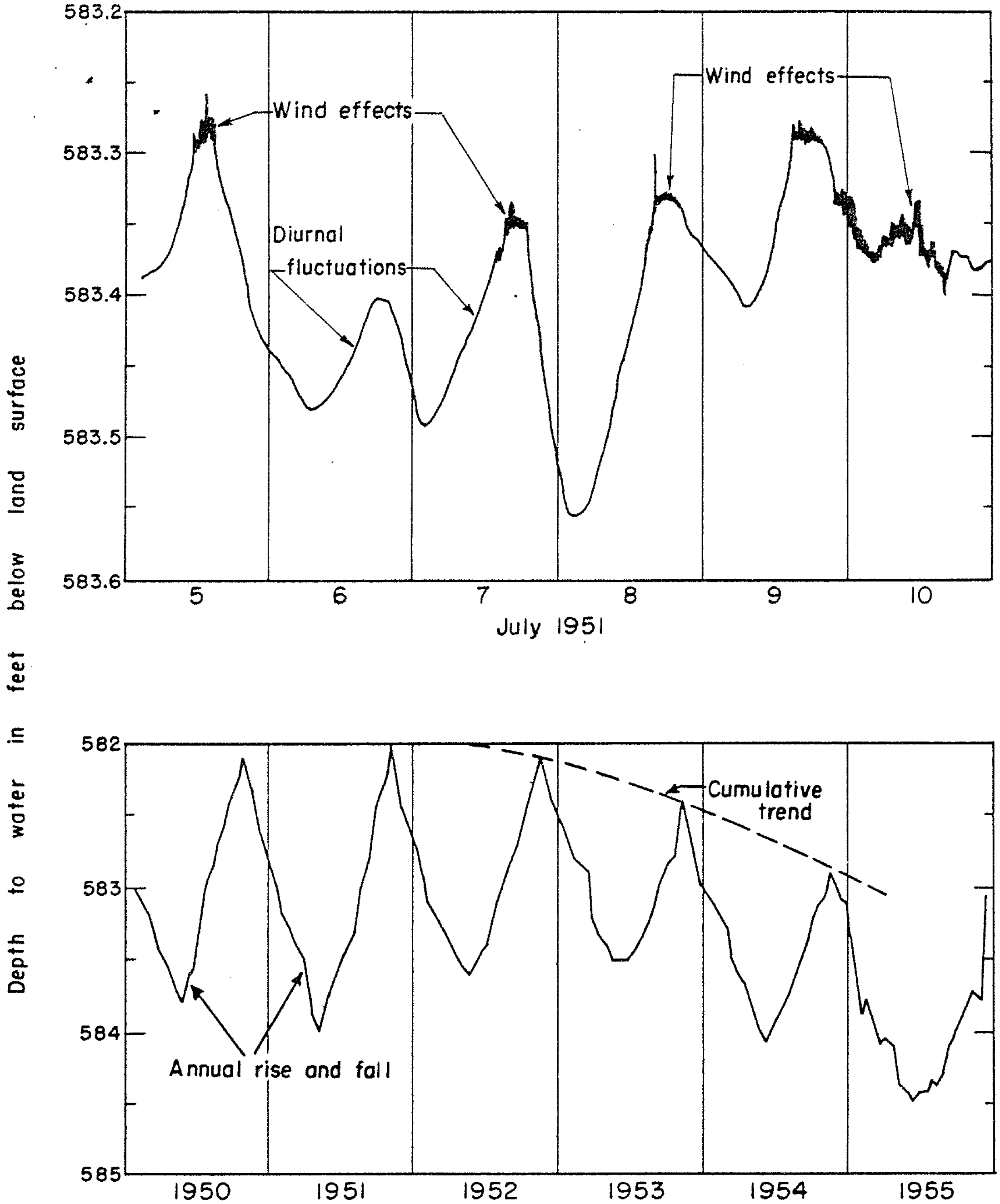


Figure 16.—Hydrograph of well 2N-31E-35dcl (USGS no. 1). Illustrates various types of concurrent fluctuations (above) and seasonal and long-term fluctuations and trends (below).

fluctuations in water levels may be caused by earthquakes, wind gusts, and barometric-pressure change. Each agent causes a characteristic type of fluctuation which is identifiable by the pattern it produces on a recorder chart.

Seismic Fluctuations

The effects of earthquake shocks on water levels in wells have been observed for many years in various parts of the United States (Stearns, 1928; Stearns and others, 1930; Leggette and Taylor, 1937; Taylor and Leggette, 1949; Parker and Stringfield, 1950). The seismic effects on water levels disclose significant facts about the aquifers, which will be discussed later. Of the three main types of earthquake waves, the surface waves seem to be registered most efficiently in wells. These waves exert a compressional force on the aquifer. The Survey maintains a nationwide program of observing earthquake effects in wells, in collaboration with the U. S. Coast and Geodetic Survey, which publishes an annual report of water-level fluctuations caused by earthquake shocks.

Fluctuations caused by seismic tremors were first observed on recorder charts from wells in the NRTS in December 1950. Since then many others have been observed on charts from the NRTS and from southern Minidoka County, Idaho. The float-actuated recording gages used by the Geological Survey in Idaho register an earthquake shock by a vertical line on the hydrograph, extending above and below the point on the hydrograph representing the position of the water level in the well immediately before the seismic shock arrived.

to local and short-term influences, and only seasonal recharge trends ordinarily are identifiable in the fluctuations. In deep wells on the Snake River Plain, short-term to nearly instantaneous fluctuations are caused by wind gusts, barometric-pressure changes, and earthquakes. Except for these, water levels in the deep wells change but little from day to day, and the net changes during longer periods are relatively small at most places.

Fluctuations of the water table are analogous to those of the water level in a surface reservoir. In both cases the fluctuations represent storage changes. Some short-term changes, and all long-term ones result from changes in ground-water storage. The short-term fluctuations also are clues to characteristics of the aquifers which otherwise might be difficult to determine.

Most water-level fluctuations in wells on the Snake River Plain can be readily identified with their causes. The fluctuations are grouped here in four categories, based on their duration: (1) Rapid fluctuations; (2) Diurnal fluctuations; (3) Seasonal fluctuations; (4) Long-term fluctuations and trends. The hydrograph of well 2N-31E-35dcl (USGS no. 1) is typical for wells on the Snake River Plain, and shows some principal types of fluctuations (fig. 16).

Rapid Fluctuations

Rapid changes in water level, other than those caused by pumping from wells, generally are caused by changes in the pressure-load on elastic aquifers or on the water surface in wells. These fluctuations usually do not involve changes in ground-water storage. Rapid discharge of ground-water to the Snake River measurably. On a proportional scale, the reduction caused by depletion at the current rate on the NRTS is negligible.

FLUCTUATIONS OF THE WATER TABLE

The static water level in a well having free communication with an unconfined zone of saturation is essentially continuous with the water table. For the purposes of this report, the two are assumed to coincide exactly. The water levels in artesian and quasi-artesian wells reflect the pressure head in the aquifer. Fluctuations of water levels in wells represent changes in the position of the water table or in the artesian-pressure head. The water-level fluctuations discussed here are those of unconfined water and the water table, including quasi-artesian water.

Periodic measurements of water levels in the NRTS and adjoining area began with a few wells in July 1949. Recording gages have operated continuously in some wells, but most of the gages were rotated among wells in order to obtain recording-gage records for representative periods from as many wells as possible. In addition, records of water levels in various parts of the Snake River Plain have been obtained in connection with other projects of the Survey. Some of those records are used in this report.

A shallow water table commonly fluctuates markedly during short periods, responding to local influences such as recharge from precipitation and irrigation, the diurnal cycle of evapotranspiration, withdrawal of water by pumping, and fluctuations in the stages of streams and lakes which communicate with the aquifer. A deep water table is less responsive

Artificial Discharge

Ground water is pumped from many domestic, irrigation, municipal, and industrial wells in the Snake River Plain. Only a part of the pumped water is consumed by use, and the rest runs off or returns to the ground-water reservoir. Industrial use is chiefly for cooling and cleaning. Consumptive use of domestic water is moderate. The greatest consumptive use is by irrigated crops.

Ground-water withdrawals from parts of the Snake River Plain outside the NRTS far exceed withdrawals on the Station. The approximate volume of withdrawals for irrigation on the plain can be estimated. Shuter, (1953, p. 8), estimated that the yearly withdrawal in 1952 from more than 90 wells in and near the Aberdeen-Springfield irrigation tract, serving 11,200 acres, was about 38,900 ac-ft/year.

Crosthwaite and Scott (1955) estimated that ground-water pumpage for irrigation in the Central Snake River Plain in 1953 was about 374,000 ac-ft. Ground water pumpage on the NRTS during the same year was about 1,150 ac-ft.

During 1955 the average daily pumpage of water on the NRTS was about 2.6 million gallons (see part 1, p. 62). Much of the water returned to ground-water storage through seepage pits and disposal wells, but records of the amount of water returned are incomplete. The average rate of withdrawal on the NRTS in 1955 (somewhat more than 4 cfs) does not noticeably affect the discharge of springs in the valley of the Snake River, because the depletion is far less than the lower limit of detection. If irrigation developments continue as predicted, they will reduce

Natural Discharge

The valley of the Snake River between Milner and King Hill is the principal natural discharge area for the ground water. The aggregate discharge rate averaged about 3,800 cfs in 1902 (Stearns and others, 1938, p. 163), but the rate increased markedly after the advent of irrigation on the Snake River Plain with surface water diverted from the Snake River. The rate now exceeds 5,500 cfs. A general check of the volume of water discharged from the ground water reservoir to the Snake River was obtained by study of the discharge rates at gaging stations on the river at Milner and King Hill. The increment to the Snake River between these stations in 1910 was about 5,250 cfs. By 1917 it had increased to about 7,250 cfs. It was relatively constant from 1917 to 1936 but increased to about 8,000 cfs between 1936 and 1942. In 1953 it was 8,526 cfs. The increment includes some ground-water and surface discharge from the south side of the river, but is largely from springs on the north side. The rate of discharge from these springs was estimated from records of a series of direct measurements of all the principal springs during 1950-55. The increase in the ground-water discharge largely reflects new recharge from irrigation on the Snake River Plain.

Substantial volumes of ground water are discharged in other areas, as by effluent seepage around the borders of American Falls Reservoir, the upstream northern edge of Lake Walcott, and along certain upstream reaches of the Snake River.

Evapotranspiration directly from the water table is proportionately small except in the Mud Lake Basin, in an area in the vicinity of Carey, and in a few other border areas of the plain.

2,900 ac-ft. That amount of pumpage would lower the water table only about 0.2 ft per year, averaged for the whole station.

On the other hand, if the effective porosity of the aquifer is 5 pct, the effective volume of stored water is 21,700,000 ac-ft and the current rate of pumpage would lower the water table only about 0.13 ft per year, averaged for the whole station.

Still assuming a static ground-water body, pumping on the Station at a greatly enlarged continuous rate of 250 cfs would remove about 182,000 acre-feet of water yearly, lowering the water table about 8.3 to 12 ft, and the reservoir would contain a 120- to 80-year supply of water.

Going farther afield, consider the area of the Snake River Plain used in the image-well study — 7,000 sq mi (4.5×10^6 ac) — and assume the same average depth of saturated thickness, — 1,000 ft. At 3.5 and 5 pct effective porosity, the volume of stored water is 1.6×10^8 or 2.2×10^8 ac-ft, respectively. If the total of effective net consumption of pumped ground water on the Snake River Plain were at a rate of 1,200 cfs continuously, annual pumpage would be 874,000 ac-ft. If there were no recharge or artificial discharge, the annual lowering of the water table, averaged for the whole area, would be about 5 ft at 3.5 pct effective porosity or about 3.4 ft at 5 pct.

DISCHARGE OF GROUND WATER

Ground water from the Snake River Plain is discharged naturally from springs and seeps and artificially from wells. Natural discharge is by far the larger, but withdrawal from wells is proceeding at an accelerating rate. Some ground water is consumed directly by evapotranspiration.

The Stationwide aggregate effects from mutual interference among wells, if all were pumped continually for long periods, could be computed. The lowering of water levels throughout the Station at the present rate of consumptive withdrawals, however, is inconsequential. Likewise, the increase of pumping lifts, caused by mutual interference, is economically inconsequential. So it does not seem worthwhile to compute Stationwide aggregate effects at this time. Moreover, further observations and tests will lead to refinements in calculations so that, by the time local and regional effects become significant, useful computations of long-term effects can be made.

Local Storage Capacity of the Aquifer

For comparison with the computed effects of pumping, it is interesting to estimate the effects of pumping that would occur in the NRTS if all water were withdrawn from storage within the Station boundaries. That is, suppose that there were no recharge or replenishment by underflow and that no water were drawn into the Station. Some simple calculations give a hint of the vast amount of water stored in the basalt, and they help to explain why prolonged pumping at current rates on the NRTS have so little observable effect.

The conservatively estimated average effective porosity of the Snake River basalt is 3.5 pct. Assuming that the average thickness of the zone of saturation is 1,000 feet, the total effective volume of stored ground water beneath the 435,000-acre Station is 15,200,000 ac-ft. Nearly a billion gallons of water was pumped in 1955, or about

Noticeable drawdown from pumping was observed at a distance of more than a mile from the pumped well. Theoretically, the cone of influence of well ANP-2 had spread to a distance of about 4 mi at the end of 72 hrs.

The cone of influence would deepen and spread much farther than 4 mi if pumping in well ANP-2 were continued indefinitely. For example, using the theoretic hydraulic systems mentioned earlier in this report and the average hydraulic properties of the Snake River basalt in the NRTS area and throughout the plain, computations show that pumping of well ANP-2 continually at a rate of 1220 gpm eventually would lower water levels in the STR area (about 18 mi southwest of the ANP area) about 0.3 ft. The lowering of pumping levels in wells in the STR area would be negligible. The amount of interference between wells is proportional to the rate of pumping. Thus, pumping well ANP-2 at a rate of 610 gpm would lower water levels in the STR area only about 0.15 ft.

The drawdown in well ANP-2 at a pumping rate of 1220 gpm and a pumping period of 72 hrs was 21.4 ft. If the well were pumped continuously the drawdown would increase until the cone of influence stabilized and equilibrium conditions prevailed. Considering hydrogeologic boundaries and the average hydraulic properties of the Snake River basalt in the NRTS area and throughout the plain, the computed drawdown under equilibrium conditions would be 21.9 ft. Thus, if well ANP-2 were pumped continuously at a rate of 1,220 gpm, the drawdown would be only about 0.5 ft greater than that measured at the end of 72 hrs. The increase in drawdown is proportional to the pumping rate. If the rate of pumping were 2,440 gpm, the increase in drawdown would be about 1 ft. The total drawdown from long-term continual pumping at 2,440 gpm would be about 44 feet.

The lowering of water levels is proportional to the rate of pumping. Consumptive withdrawals of 150 cfs at NRTS, for example, would result in drawdowns at points of observation 10 times those given in table 21.

Table 21.—Estimated effects of potential ground-water development in NRTS on water levels in Snake River Plain

[Assumed consumptive use on NRTS: 15 cfs (11,000 ac-ft per year)]

Point of observation	Drawdown of water levels (feet)
Mud Lake	0.6
Midway between Taber and Blackfoot	0.2
Southern Minidoka County	0.1

Local Effects of Pumping in the NRTS

Withdrawals of ground water from production wells in one part of the NRTS eventually will lower water levels in other parts. The data from an aquifer test during Nov. 12-15, 1953, using a group of wells in the ANP area, illustrate local interference among wells at the end of a relatively short period of pumping. The drawdown of water levels in observation wells, caused by pumping of well ANP-2 at a rate of 1220 gpm for 72 hrs, were as follows:

Well	Distance from pumped well (feet)	Drawdown (feet)
ANP-1	595	0.80
ANP-Disposal	1,960	0.45
IET-Disposal	5,450	0.1

steepens. Therefore, underflow from the NRTS to the plain decreases, and inflow to the NRTS increases. Water levels in the Snake River Plain are affected by changes in underflow from the NRTS to the plain and by changes in inflow to the NRTS from the plain. The magnitude of these effects from pumping at the current low rates, of course, is vanishingly small.

Ground-water pumpage from production wells on the NRTS was about 4 cfs or 2,900 ac-ft in 1955 (see pt. 2, table 10). Much of the pumped water was returned to ground storage through disposal wells and infiltration pits, so the net depletion of ground water was low, probably ranging from 15 to 50 pct of the amount pumped at different facilities. The estimated consumptive use of water on the NRTS in 1955 is 1.3 cfs. Theoretically, withdrawal of this small amount of water will lower water levels in irrigation-well fields on the Snake River Plain only a few hundredths of a foot or less. That is, the lowering from this cause will be indistinguishable from the lowering from other causes.

Suppose that the consumptive use of ground water on the NRTS increases by 1975 to 15 cfs (11,000 ac-ft per year). Computations similar to those described in preceding sections of this report were made to determine the effects of withdrawing 15 cfs on water levels in irrigated areas on the Snake River Plain. The results of the computations (table 21) show that withdrawal at the assumed rate would not lower water levels noticeably in the Mud Lake, Taber or Minidoka areas. No allowance was made in the computations for the partial hydraulic boundary between the NRTS and Mud Lake. The boundary probably reduces greatly the effect of pumping at NRTS on water levels in the Mud Lake area.

computations were made to illustrate the influence on the total drawdown in the NRTS area of possible variances in the average hydraulic properties of the aquifer (table 20). The table shows that the order of magnitude of computed drawdowns is not changed by varying the aquifer properties within reasonable limits. The computed values of drawdown that are based on T and S values of 8×10^6 gpd/ft and 0.05 are believed to be realistic, so the drawdown in the NRTS will be on the order of 5 to 10 ft after a long term of years.

Table 20.--Computed effects of pumping on water levels in the NRTS, at different values of T and S.

Assumed values of T and S	Computed drawdown of water levels in NRTS area (feet)	
	Based on consumptive use in 1955	Based on consumptive use in 1975
$T = 4 \times 10^6$ gpd/ft, $S = 0.025$	7	16
$T = 8 \times 10^6$ gpd/ft, $S = 0.05$	5	11
$T = 1.2 \times 10^7$ gpd/ft, $S = 0.10$	2	6

Effects on Water Levels in the Snake River Plain from Ground-Water Development in the NRTS

The amount of underflow beneath the NRTS is adequate to supply all existing facilities and all additional ones that are definitely foreseen. However, the area is not an isolated cell. Pumping from wells in the NRTS causes a regional lowering of water levels within the NRTS. As a result, the hydraulic gradient out of the southern part of the NRTS is reduced, and the gradient from surrounding areas on the northeast entire saturated thickness of the Snake River basalt is at least twice the value of T computed from test and specific-capacity data. Further, because some basalt drains very slowly, the coefficient of storage computed from test data probably is much smaller than the true coefficient of the aquifer. During the early part of a pumping test, water is readily derived from storage by gravity drainage from relatively large voids; later, this is augmented by the slower drainage from small voids. The long-term average coefficient of storage of the Snake River basalt probably is in the magnitude of 5 pct (0.05). To determine the long-term effects of real and image wells, T and S values of $8x10^6$ gpd/ft and 0.05 were substituted in the nonequilibrium formula.

Computed drawdowns in the NRTS as the result of present and future developments on the Snake River Plain were shown in table 19. Geologic and hydrologic studies made in the area between the NRTS and Mud Lake indicate the presence of partial hydraulic boundaries in that area, which would tend to reduce the drawdown that would be caused at the NRTS by withdrawals in the Mud Lake area. The influence of the partial hydraulic boundaries can not be appraised with the information now available. The anticipated total drawdowns at the NRTS after 100 years, which would be caused by the estimated pumpage in 1955 and 1975, are 4 and 10 ft, respectively. This estimate makes some allowance for the partial boundaries near Mud Lake.

The theoretical hydraulic system was modified somewhat and computations of drawdowns were made, using T values from $4x10^6$ to $1.2x10^7$ gpd/ft, and S values from 0.025 and to 0.10. These conditions are believed to be minimum and maximum possibilities of the aquifer situation. The

Table 19.—Estimated ground-water pumpage and its effects on water levels in the NRTS at the end of 100 years

Area of heavy withdrawal	Withdrawal rate (Acre-feet per year)		Consumptive use (Acre-feet per year)		Computed lowering of water levels in NRTS area (feet)	
	1955	1975	1955	1975	Based on 1955 consumptive use	Based on 1975 consumptive use
Mud Lake Basin	270,000	400,000	160,000	240,000	2.3	3.9
Minidoka Co.	160,000	500,000	100,000	300,000	1.2	3.7
Taber-W. Bonneville	80,000	240,000	60,000	180,000	1.1	2.5
Aberdeen-Springfield	60,000	90,000	33,000	50,000	0.6	0.9
E. Jerome Co.	28,000	40,000	14,000	20,000	0.2	0.3
Gooding & W. Jerome Co.	9,500	18,000	5,000	9,000	-	0.1
Total	607,500	1,288,000	372,000	799,000	5.4	11.4

Effects on Water Levels in the NRTS from Regional Ground-Water Development

Estimated ground-water withdrawals from irrigation wells on the Snake River Plain in 1955 and 1975 are given in table 19. Little more than half of the pumped water is consumptively used. The unconsumed portions run off as surface waste or return to the ground-water reservoir. Unconsumed irrigation water which filters into the aquifer offsets part of the effects of pumping.

The hydraulic system outlined earlier and estimated amounts of consumed water given in table 19 were used to determine the drawdown of water levels in the NRTS area as a result of ground-water development on the plain. For computation purposes, withdrawals from areas of heavy development were assumed to be concentrated at points. The effects of the real pumping wells and the image wells associated with the hydrogeologic boundaries of the Snake River basalt were computed by means of the nonequilibrium formula. A time interval of 100 years was used in computations because it takes that long for the full effects of pumping at remote places to reach the NRTS.

The results of aquifer tests and specific-capacity data indicate that the average coefficients of transmissibility and storage of the Snake River basalt are $4x10^6$ gpd/ft and 0.025, respectively. These values were computed from data on wells which penetrate only part of the aquifer and which were pumped during relatively short periods. A study of the gradient of the water table and underflow beneath the Snake River Plain indicates that the average coefficient of transmissibility of the

Thus, the basalt aquifer is not infinite in areal extent; it is finite, and the Snake River is a finite boundary. This condition, and the fact that the river is not a straight-line demarcation, were provided for in our hypothetical hydraulic system by increasing the distances from heavily pumped areas to the line of recharge and by assuming an aquifer with rectangular boundaries.

The influence of the hydrogeologic boundaries on the regional effects of pumping were determined by means of the image-well theory (see Ferris, 1949). The image-well theory as applied to ground-water hydrology may be stated as follows: The drawdown in a pumped well is increased by pumping from another well. A barrier boundary also increases the drawdown, and the effect on the pumped well is the same as though the aquifer were infinite and another well, discharging at the same rate, were located across the real boundary on a perpendicular thereto and at the same distance from the boundary as the real pumped well. For a recharging boundary the principle is the same except that the image well is assumed to be recharging the aquifer instead of pumping from it. The hydraulic system discussed earlier consists of parallel boundaries. Analysis of parallel boundaries by the image-well theory requires a multiple image-well system (Knowles, 1955). The image-well theory was applied in deriving the estimates in the following section of this report.

system consisting of a rectangular aquifer, 190 and 65 miles in length and width, and enclosed by two intersecting barriers and two intersecting recharging boundaries. The barrier boundaries are the borders of the Snake River basalt, about 10 miles northwest and 80 miles northeast of the CPP area in the NRTS. These boundaries coincide approximately with the bases of mountains and hills that border the entire Snake River Plain on the northwest, north and east. The recharging boundaries are two segments of the Snake River.

Along the Snake River above Milner Dam, influent reaches alternate with effluent ones. Springs and seeps discharge ground water to the river in some reaches, and water percolates from the river into the aquifer in others where the water table is below stream level. From Milner Dam to Bliss, the water table is higher than the river and ground water is discharged continually.

Lowering of water levels along river reaches where influent or effluent conditions exist will lessen the discharge of ground water to the river or increase percolation of water from the river into the aquifer. Thus, the effluent reaches of the river where the water table is above the bed of the river are recharge boundaries which limit the spread of the cone of influence. Pumpage on the plain ultimately will reduce the amount of water discharged from springs in the valley of the Snake River and increase stream losses along influent reaches of the river.

In certain reaches of the river, as from Roberts to Firth and from Lake Walcott to the vicinity of Milner the regional water table is below the bed of the river. Further lowering of water levels would not increase river losses, so these reaches of the river are not recharge boundaries.

and to an observation point were 20 mi, about 50 years would elapse before the cone stabilized at the observation point. By then the cone would have spread more than 200 miles from the pumped well, if the aquifer were that extensive, so the boundaries of the aquifer would have been reached long since. Thus, ground-water withdrawals remote from NRTS ultimately will lower water levels in the Station, and pumping at the Station will lower water levels elsewhere on the Snake River Plain.

Influence of Hydrogeologic Boundaries

The hydrogeologic boundaries of the Snake River basalt are irregular in shape and extent, and the hydraulic properties of the aquifer vary from place to place on the plain. The regional lowering of water levels by pumping, however, is not affected much by irregularities in hydrogeologic boundaries nor by variability of hydraulic properties. The regional lowering is controlled largely by the average hydraulic properties of the Snake River basalt within the large areas of influence of pumping and by the general shape and extent of the boundaries.

The order of magnitude of the regional effects of pumping can be determined by treating the hydrogeologic boundaries as straight-line demarcations and making studies of a hypothetical hydraulic system that satisfies the hydrogeologic conditions of the aquifer. The results of aquifer tests and specific-capacity data can be used to estimate the average hydraulic properties of the aquifer.

Studies in the Snake River Plain show that the effects of hydrogeologic boundaries on the response of the aquifer to pumping from wells can be simulated by mathematical analysis of a hypothetical hydraulic

Regional Effects of Pumping

The important factors controlling the response of the Snake River basalt to the ground-water development just described are (1) the hydraulic properties of the aquifer (coefficients of transmissibility and storage), and (2) the hydrogeologic boundaries of the aquifer (border of the Snake River basalt and areas of recharge and natural discharge).

Under natural conditions, before development of wells, the water in the Snake River basalt was in a state of approximate dynamic equilibrium in which discharge from the aquifer balanced recharge. Discharge by wells disturbs the natural state of equilibrium. Pumping develops a cone of influence around the pumped well. The cone, with its center at the pumped well, spreads radially outward as water is taken from storage in the aquifer and water levels are lowered in the area of influence. The cone continues to grow until (1) the lowered water levels cause increased recharge to or decreased natural discharge from the aquifer, or a combination of these, and (2) the hydraulic gradient between the pumped well and the recharge and natural discharge areas is sufficient to bring to the well the amount of water pumped. The dimensions of the cone of influence depend upon the hydraulic properties of the aquifer, the location and character of hydrogeologic boundaries, and the rate of pumping.

Considerable time elapses before the cone stabilizes, but thereafter no water is taken from storage, and a new state of approximate equilibrium is established. For example, computations show that if the distance from a pumped well in the Snake River basalt to a recharge area year by 1975; in the Mud Lake basin, about 400,000 acre-feet per year by 1975; and in the Taber-West Bonneville County area, about 240,000 acre-feet per year by 1975.

Table 18.—Estimated pumpage of ground water for irrigation on Snake River Plain, 1955

Area	Water pumped (acre-feet per year)
Mud Lake Basin	270,000
Southern Minidoka County	160,000
Taber-W. Bonneville County area	80,000
Aberdeen-Springfield tract	60,000
Eastern Jerome County	28,000
Gooding and Western Jerome Counties	9,500
Big Lost River Valley	14,000
Little Lost River Valley	11,000
National Reactor Testing Station	2,900
Total	635,400

In the Mud Lake Basin ground-water withdrawals are chiefly from a perched water-bearing zone at shallow depth and from artesian wells. In the Big Lost and Little Lost River valleys, pumping is from an unconfined water body where the depth to the water table ranges from 3 to 100 ft below the land surface. Pumping in these areas is from ground-water bodies that are tributary to the plain, and increased pumpage of the tributary water will reduce underflow to the plain.

levels elsewhere. Currently, there is no shortage of ground water in the Snake River Plain, but downstream users of Snake River water have a stake in continuous copious discharge by the springs. Users of ground water on the plain ultimately will compete for water with downstream users. Also, heavy pumpage in small areas will lower water levels locally. That is, local competition will develop long before regional competition.

Most of the accessible water supply of the Snake River and its tributary streams has been appropriated. Expansion of the irrigated area is contingent upon more efficient use of surface water and on the availability of ground water. Irrigation with ground water increased markedly on the Snake River Plain during the 10 years from 1946 to 1955, and about 75 pct of the increase occurred after 1950. Most pumping of ground water has been done along the fringe of the Snake River Plain and in tributary valleys. Since 1950 developments have spread from the fringes toward the interior of the plain, as in the Taber-West Bonneville County area and north of the Minidoka Irrigation District in Minidoka County.

Withdrawals of ground water for irrigation on the Snake River Plain are estimated in table 18. Pumpage was about 635,000 acre-feet per year in 1955 and is expected to double by 1975 (see table 19). The expected increases will be small in the Little Lost River Valley and in eastern Jerome County, but large in other areas. The centers of areas of heavy withdrawals are located near Mud Lake, about 10 miles northwest of Blackfoot (Taber-West Bonneville County area), west of American Falls Reservoir (Aberdeen-Springfield tract), in southern Minidoka County and in eastern Jerome County. The largest future ground-water developments will be in southern Minidoka County, approaching 500,000 acre-feet per unified, forming a single regional system. The NRTS taps water in that system upstream from many sources of recharge, but downstream from other sources. The Station, therefore, is situated favorably for a large and permanent water supply. The available supply, however, is not limitless, owing to geologic factors that control the movement and accessibility of the water, and to physical, economic, and competitive factors that determine the feasibility of water recovery. The economic factors are not discussed in this report.

The perennial ground-water yield of the Snake River Plain, computed from water records for the natural-discharge area between Milner and Bliss, is at least 4 million acre-feet a year. The average continuous discharge of the springs is more than 5,000 cfs and may be as much as 5,500 cfs. This represents all or most of the ground-water yield from the plain to the east. All ground-water developments on the plain will tap the supply, and each development will have local and regional effects on water levels. Therefore, the following factors will now be considered: (1) competitive demands for water; (2) regional effects of pumping; (3) local and regional effects of ground-water development on the NRTS.

Competitive Demand for Water

Several large developments of ground water for irrigation are in progress on the Snake River Plain, and other large developments are expected. The water discharged from springs in the Snake River valley is derived from the same regional body of ground water that supplies these developments. The ground-water withdrawals ultimately will lower water levels at the NRTS, and withdrawals on the Station will affect water

Contributions to total underflow beneath the NRTS at the latitude of the STR area

	Cfs	Ac-ft/yr
Underflow from Little Lost River Valley	85	61,000
Underflow from Birch Creek Valley	140	100,000
Underflow from Mud Lake Basin	100	71,000
Seepage loss from Big Lost River	25	18,000
	350	250,000

The computed contributions from the Little Lost River Valley and Birch Creek Valley are much larger than those estimated by Stearns and others (1938). They estimated that underflow is 37,000 ac-ft/yr from the Little Lost River Valley and 57,000 ac-ft/yr from the Birch Creek Valley. Their estimates were based largely on streamflow losses and they did not include underflow from the headwater areas of the streams, which does not appear at the land surface.

POTENTIAL GROUND-WATER DEVELOPMENT AND ITS EFFECTS

Although the Snake River Plain is large and diverse, certain features are characteristic of the plain in general. From northeastern Idaho southwestward and westward to the valley of the Snake River between Milner and Bliss, the net direction of ground-water underflow is southwestward and westward. Throughout the plain only a small part of local precipitation becomes recharge water. Most of the perennial supply of ground water is replenished by underflow from tributary peripheral areas. Throughout the plain, permeable basalt aquifers yield water copiously to wells. The systems of aquifers and tributary aquifers in the plain are compensate for the decreased hydraulic gradient in that section. Based on aquifer-test and specific-capacity data, and allowing for the fact that the wells tested penetrate only part of the Snake River basalt, the estimated average coefficient of transmissibility is $2x10^6$ gpd/ft in the part of the section east of the STR area and $4x10^6$ gpd/ft in the western part. Values of I and L for the two parts of the section were scaled from plate 2. The computed underflow through the eastern and western sections is $6.1x10^7$ and $1.6x10^8$ gpd respectively. The estimated total underflow beneath the NRTS at the latitude of the STR area, therefore, is $2.2x10^8$ gpd or about 350 cfs.

The sources of the water moving through the section of the aquifer under discussion are (1) underflow from the Little Lost River Valley, (2) underflow from the Birch Creek Valley, (3) underflow from the Mud Lake Basin, (4) and seepage loss from the Big Lost River. The average rate of seepage loss from the Big Lost River north of the 4480 water-table contour, estimated from the streamflow data in appendix 2, is about 25 cfs (18,000 ac-ft/yr). Underflow contributions from the Little Lost River Valley, Birch Creek Valley, and the Mud Lake Basin were computed by the modified Darcy equation, using estimated values of T and flow lines drawn to delimit the sections of the aquifer through which the increments of underflow are moving. Following are the results of the computations:

VOLUME OF UNDERFLOW THROUGH THE NRTS

The volume of underflow through an area can be computed by the following modified form of the Darcy equation (see Knowles, 1955):

$$Q = TIL$$

in which Q is the discharge in gallons per day, T is the coefficient of transmissibility in gallons per day per foot, I is the hydraulic gradient in feet per mile, and L is the width in miles of the cross section through which discharge occurs.

The rate of underflow through the Snake River basalt within the NRTS at the latitude of the STR area, between the 4480- and 4490-foot water-table contours (see plate 2), has been computed. The transmissibility of the aquifer is variable within this section. The water-table contour map (see plate 2) indicates that the average transmissibility in the section of the aquifer east of the STR area is much less than that to the west. Based on a study of limiting flow lines, the water moving through the section of the aquifer west of the STR area is for the most part underflow from the Little Lost River Valley. A part of the underflow from the Little Lost River Valley, underflow from Birch Creek Valley, and underflow from the Mud Lake basin is moving through the section of the aquifer east of the STR area. Underflow from Birch Creek Valley and the Mud Lake basin is much greater than that from the Little Lost River Valley. Therefore, the volume of underflow east of the STR area is much greater than that west of the STR area. However, the hydraulic gradient of the water table west of the STR area is much greater than that east of the STR area. The greater width of the cross section east of the STR area does not fully however, that the lower water table in the west is caused by some other factor, such as an area of high permeability in the southwest, which would tend to bleed water off and create a gradient in that direction.

Ground water that flows southeastward into the NRTS from the Little Lost River Valley seems to have only local effects on the direction of movement through the Station.

In summary, water enters the NRTS from the north, northeast and northwest and moves southward to southwestward through the Station. The direction of movement beneath a given point or specified area, however, probably differs appreciably from the directions indicated by the map.

Most planning maps for construction areas on the NRTS include a directional "rose" showing true and magnetic north, prevailing wind direction, direction of surface drainage, and direction of ground-water underflow. The direction of underflow should be prominently labelled as a rough approximation.

Regional Movement

Ground-water in the Snake River Plain moves generally southwestward toward the famous springs along the valley of the Snake River between Milner and Bliss. The hydraulic gradients are variable but generally they are on the order of 5 to 10 ft/mi. Practically nothing is known about the hydrology of the vast, uninhabited central lava plain between the NRTS and the belt of irrigated land on the west, adjacent to the spring-discharge area. Most of the water from the northeast moves through the uninhabited plain on its way to the discharge area.

The maps accompanying this report are not precise. However, they are accurate because they are based on a reasonable amount of data, they are consistent with the data, and they show the true position of the water table within a small margin of error. Therefore, they correctly generalize the form and position of the water table.

Movement Beneath the Station

Accepting the water-table map (pl. 2) as an approximately correct regional picture, it shows that water enters the northern NRTS from the north, west of Circular and Antelope Buttes. Underflow in this area is chiefly from the Birch Creek Valley and the western part of the Mud Lake Basin. A few miles to the south in the NRTS this water veers southwestward.

West of Circular Butte, underflow from the Mud Lake basin is southward into the plain west of the NRTS. This water also veers southwestward and some of it enters the southeastern part of the NRTS. At least three factors, separately or together, may cause this behavior. (1) The influx of water from the Mud Lake basin may be greater than that from the north. A heavier influx from Mud Lake would tend to build up the water table in the east and create a gradient toward the Station. This cause is at least partly responsible for the observed phenomena. (2) A non-permeable barrier south of the Mud Lake basin would tend to impound ground water, which would establish a westward gradient and move in that direction. No direct evidence of a barrier is available. (3) A preponderance of preferred directions of permeability may trend westward or southwestward, shunting water in that direction. It seems more likely, Cooling cracks tend to be normal to the cooling surfaces of a lava flow, regardless of the direction of movement. Open joints caused by differential movement and settling of hardened lava crusts seem to be quite random in orientation. The longitudinal crevasses in border rolls are parallel to the front of the flow, and the transverse fractures on the border roll are about normal to the front.

In all cases, as successive flows accumulated, the basalt would consist of successive layers having differing directional permeabilities. In projection onto a plane these directions would form a random pattern of interwoven lines.

Many of the late lava flows erupted from a line of volcanic vents along the axis of the Snake River Plain, and some flows spread northward and southward (see part 2, p. 77-78). If this was true also of the deeper lava flows in the zone of saturation, and if the direction of greatest permeability has some systematic general relation to the direction of flow, then the average direction of greatest permeability beneath the NRTS may be westward or southwestward. Vectorial variation may account for some pecularities of the configuration of the water table beneath the NRTS.

In summary: for mathematical treatment and descriptions of conditions, idealized assumptions are necessary because, without them, no description of the behavior of water or an aquifer could be derived at all. The results must be weighed against observed facts and modified on the basis of experience and judgment. Similarly, observed data must be interpreted and used with understanding of their nature and limitations, and the distinction between precision and accuracy must be recognized.

therefore, are not shown by water-table contour maps. These maps, based on water-table altitudes in widely scattered wells, are constructed by the method of logical contouring, modified somewhat by geologic judgment. A map, therefore, is a rather crude generalization. It correctly represents the general configuration of the water table, and from it the general regional directions of underflow may be inferred. Thus, the map is accurate without being precise about local details.

For these reasons, the writer prefers to think of underflow as being along the path of least resistance — a path which seldom can be determined or shown on a map. The path of least resistance is not necessarily the direction of maximum permeability either, because permeability is only one of several factors that control the water movement. At most places, movement probably is at an angle to the direction of maximum permeability and deviates from the apparent direction of maximum gradient shown on a map.

Through wide areas and large depth intervals, the orientation of directional variations in the permeability of basalt layers is random. For example, lava tubes, whether large or very small, tend to extend roughly parallel to the direction of flow of the lava. Highly fluid thin flows of lava develop a lineation of small tubes and pores in the direction of flow, and in some flows these may constitute most of the effective porosity. A succession of flows that spread radially outward from a volcanic throat or fissure would have a radial pattern of lines of maximum permeability. In overlapping flows from a neighboring eruptive center, however, the lines would have different directions.

of movement cannot be calculated accurately for any specific locality because of the heterogeneity of the aquifer and other quantitative unknowns. A general or average velocity might be determined by identification and direct measurement through test holes of a thread of water at successive points. The water follows tortuous paths through the basalt, however, and its speed along those paths can not be measured. The apparent speed would be net velocity of change of position in a straight-line direction between the points.

Direction of Movement

Ordinarily it is assumed that the movement (underflow) of ground water is in the direction of the maximum grade of the water table, at right angles to the water-table contours. This assumption is correct for the idealized conditions which must be assumed in mathematical treatments — a frictionless fluid in an isotropic aquifer. But natural aquifers are not perfectly homogeneous, and the Snake River basalt is conspicuously heterogeneous in its physical properties. For example, great directional differences in permeability and porosity are obvious from inspection. For practical purposes, a compromise must be reached between idealized and actual conditions.

Where ground-water moves through successive rock zones which differ in their permeability, the flow lines are refracted in a way analogous to the refraction of light rays as they pass through media having differing optical densities. Precise water-table contours would bend wherever the flow lines bend, but the real pattern is so complex and localized that it eludes detection in a coarse grid of observation points. The details, The velocity of movement through massive basalt obviously is small to nil. For the basalt as a whole, crude estimates indicate average velocities in the range of about 5 to 30 ft per day, depending on the values assumed for porosity, saturated thickness, and transmissibility. Assuming a saturated thickness of 1,000 feet, a gradient of 5 ft per mile, and uniform distribution of porosity, the following range of velocities is indicated:

Transmissibility (gpd/ft)	Porosity (pct)	Velocity (ft/day)
2,000,000	1	26
do	2	13
do	3.5	7.4
do	5	5
5,000,000	2	32
do	3.5	18
do	5	13

On the other hand, we know that the porosity of the aquifer is not uniform, some zones being extremely permeable and others having little or no permeability. Most of the movement is through the highly permeable zones, where the velocities greatly exceed those in other zones. So trial computations have been made, using average transmissibility values, and estimates of the thickness and porosity of highly permeable zones. These computations indicate velocities ranging up to 300 ft/day.

From place to place and from layer to layer, the velocities evidently vary from almost nil to several hundred feet per day. The speed

In the case of water flowing through heterogeneous interflow sediments, the critical value of R probably is small, but the velocity is so low that laminar flow can be assumed safely. In the basalt, on the other hand, the paths of flow are complex and devious. The range in absolute velocities through cracks, crevices, joints, and other voids must be very wide. Possibly, turbulence occurs where the openings are small and the velocity is high, but the velocities even there may be too low to cause turbulence.

These generalizations lead to no definitive answers to practical problems. Natural ground water mixes with introduced solutions to some extent as a result of turbulence in the close vicinity of intake wells. Dispersion also causes some dilution. But so long as we lack quantitative information about the extent of dilution by mixing and dispersion, we must assume that ground-water flow is laminar and that no measurable mixing occurs except by chemical diffusion.

Rate of Movement

No underflow velocities of ground water have been determined on the Snake River Plain, but experiments and observations are in progress. Determinations of velocity at specific locations would be very valuable to establish a range of magnitude, but the values would not be generally applicable to the entire NRTS. Theoretic calculations have been made, based on hydrologic observations and assumptions, and if these were checked empirically at a few locations, procedures might be developed for deriving reasonably accurate estimates for other locations.

The nature of ground-water movement is important in many ways, and it is a critical factor in the waste disposal problem of the atomic-energy industry. Where flow is laminar, liquid waste that reaches the zone of saturation may be diluted but little except by the process of diffusion. A heavy liquid waste might gravitate into "dead" areas, low in the zone of saturation, where the flow velocity is small to nil, and remain there for a long time. It might not be a hazard even though undiluted. If flow is turbulent, however, considerably more mixing and dilution of wastes will occur.

An index of the tendency of flow to be turbulent is the Reynolds number, which is defined by the equation, $R = \frac{LV}{v}$, in which R is the Reynolds number, L is a typical dimension (a tube diameter, a depth of channel, etc), V is the velocity of flow, and v is the kinematic viscosity of the fluid.

The threshold of turbulent flow in closed full pipes generally is assumed to be about at $R = 2,000$; for free surface flow it is about $R = 500$; for ground-water flow the experimentally determined threshold is between 1 and 10 (Fancher and others, 1933). The critical value of R varies considerably with boundary geometry, owing partly to the arbitrary definition of the dimension, L. For pipes, L is the diameter; for free surface flow it is the depth of the water; and for ground-water flow it is the average diameter of grains in the water-bearing medium. The concept of "grains" is not applicable to basalt, but many of the water-bearing openings are more like rough, irregular pipes. Assuming various diameters of "pipes" and using water velocities in ranges up to several hundred feet per day, the computed Reynolds numbers are far below the threshold of turbulence.

from gravel aquifers, and the nature and rate of movement also may differ radically. There are bound to be wide local variations in the speed of movement. Therefore, knowledge of the velocity past a point, if it could be determined, might have little practical value. No empirical data are available about water movement in the basalt and, for the present, study of the problem must be based largely on theoretic principles.

Nature of Movement

We have no direct information about the nature of ground-water movement because it cannot be observed directly. The movement usually is assumed to be laminar (streamlined). Darcy's law, the Thiem formula, the Theis nonequilibrium formula and others assume laminar flow. Under laminar-flow conditions velocity varies directly as the loss in head.

The applicability of Darcy's law and its linear relationship of velocity to head under conditions of laminar flow has been disputed from time to time. Absolute proof of its applicability is lacking, but circumstantial evidence supports it. The observed responses of aquifers and hydraulic systems to pumping and recharge are quite close to the responses calculated from the classic formulas. Therefore, the assumptions on which the formulas are based must be substantially correct for ordinary aquifers under ordinary conditions. Although the Snake River basalt is not an ordinary aquifer, its observed areal responses to recharge and discharge also conform reasonably closely to calculated responses. The weight of evidence now available favors belief that turbulent motion is rare, if it occurs at all, except in the close vicinity of discharging and recharging wells, heavily flowing springs, and perhaps in ground-water cascades.

"underground rivers," vast "underground lakes," mysterious sources of water deep in the earth's interior, or great underground tunnels that somehow tap the Great Lakes or the ice fields of the far north, as has been postulated by imaginative laymen. The pore spaces of ordinary rocks and sediments are quite adequate to store and transmit very large quantities of ground water, and precipitation on the basin of the Snake River easily accounts for all water perennially discharged from the plain.

GROUND-WATER MOVEMENT

Ground water moves constantly from recharge areas toward discharge areas. The rate of movement generally is slow compared to that of surface water. Early experiments (Slichter, 1905) on Long Island, New York and in the Mohave, San Gabriel, and Rio Hondo River valleys in California indicated underflow velocities up to 96 ft per day in alluvial aquifers. Studies in the South Platte Valley near Ogallala, Nebr. (Slichter and Wolff, 1906), showed velocities of about 14 ft per day. The conditions of the tests were not fully described so they cannot be duplicated by check experiments. Later computations, based on permeability and hydraulic gradients, have uniformly indicated much slower average underflow velocities in similar aquifers. Fifty to 60 ft per day is believed to be in the high range of velocities of underflow in alluvial aquifers under ordinary conditions.

The mode and rate of movement of ground water are important factors in several operational problems on the NRTS, especially liquid-waste disposal. The basalt aquifers of the Snake River Plain differ radically

3.5 pct in the basalt and 25 pct in the lesser volume of water-bearing sediments -- probably is at least 5 pct. If so, the effective volume of stored water is about $5x10^8$ ac-ft, or about 135 times the storage capacity of the surface reservoirs.

All the water stored in effective pore space is "holdover" storage and is between 200 and 300 times the average amount of holdover storage in the surface reservoirs. If the plainswide average of precipitation is 10 in., its average yearly volume on the plain is about $8.3x10^6$ ac-ft. The water stored in effective pore space is equivalent to all the precipitation on the plain in 60 years. These estimates accent the size of the ground-water resource.

Although the volume of stored water is enormous, the effective volume which it would be practical to use is much smaller. Most of the water is in dead storage, and to pump a large percentage of it would entail heavy drawdown, high pumping lift, and excessive cost. That is, the long-time average rate of withdrawal cannot exceed the long-term average rate of replenishment without "mining" holdover storage that would not be renewed. The unconsumed perennial yield of this aquifer is about $4.2x10^6$ ac-ft. Spread out in the zone of saturation east of King Hill, this would occupy a thickness of only 8.4 ft of the aquifer.

These very crude estimates surely are substantially in error, but they are useful, as will be shown later. Also, they belie certain popular notions. For example, they illustrate that the tremendous ground-water yield of the Snake River Plain does not require the presence of

AMOUNT OF GROUND WATER IN STORAGE

The ground-water reservoir of the Snake River Plain east of King Hill underlies nearly 11 million acres of land. Drilling records show that water-bearing rocks are present to depths of at least 1,500 ft at the NRTS and 1,600 ft at Idaho Falls. The maximum and average depths of the saturated zone are not known because neither the full depth nor the whole area has been explored. Nevertheless, we can estimate the amount of water in storage, using general knowledge about the aquifer and its properties.

Assume, for the moment, that the average thickness of water-saturated material is 1,000 ft. The median small-pore porosity of the basalt is about 13 pct (see p. 58). The estimated effective formational porosity of the basalt from open fractures and other megascopic voids is about 3.5 pct (see p. 61). Interflow sediments and sediments intertongued with the basalt around the plain probably average about 25 pct in porosity. Their volume is less than that of the basalt, but they raise the overall average porosity somewhat. Therefore, it seems likely that the plains-wide average porosity of water-bearing material is about 20 pct.

On that basis, the 10-million acre area east of King Hill contains 10^{10} ac-ft of saturated material in which the total water content is $2x10^{9}$ ac-ft. This is roughly a thousand times the combined storage capacity of all the surface reservoirs on the Snake River above King Hill.

Only a part of the total volume of water is in effective pore space. Most of it is in capillary pores and closed voids from which it would not drain by gravity. The plainswide average effective pore space — assuming

The estimate actually is conservatively low despite increased consumptive use of water in the Mud Lake basin since 1927. The estimates for contributions from the Centennial and Beaverhead drainage were based on surface discharge of streams and loss of water to the ground in the streams. Ground-water underflow from the mountains and valleys that never appears at the surface was not included. The total unconsumed underflow through the Mud Lake basin may be two or more times the amount estimated. None of the water is directly available to the NRTS. Nevertheless, the basin is a major factor in the regional hydrology because through it moves a large increment of water to the plain in and near the NRTS. This fact would bear materially on the water-supply problem at the Station if a very large supply were developed in the future. The area of influence of large fields of heavily pumped wells might extend far beyond the boundaries of the Station, and the wells then would draw in water from the adjacent area.

Mud Lake and associated small lakes, ponds, and sloughs are fed by effluent ground water. The hydrology of the basin is extremely complex, for it contains extensive bodies of perched ground water, copiously productive artesian aquifers in some areas, and a main zone of unconfined ground water. All or most of the ground water that is not consumed in the basin is discharged southward and southwestward, becoming a part of the main body of water in the basalt beneath the Snake River Plain.

Plain during 1921-1927 were 36,700 ac-ft from the Little Lost River valley and 57,300 ac-ft from Birch Creek valley. Later estimates are given in the section of this report titled "Volume of underflow through the NRTS".

Underflow from the Mud Lake Basin

The Mud Lake basin, which lies east and northeast of the northern part of the NRTS, receives large contributions of ground water by underflow from the valleys and mountains to the north, chiefly the Centennial and Beaverhead Mountains. Surface drainage into the Mud Lake basin is very small, and no water leaves the basin as surface runoff. Stearns and others (1938, p. 231) estimated that the underflow contributions to the plain from Beaver and Camas Creeks above Dubois during 1921-24 averaged about 109,000 acre-feet annually, and that from Medicine Lodge Creek was about 49,500 ac-ft. Much of the ground water is used for irrigation in the Mud Lake basin, but the unconsumed residual ground-water recharge probably is on the order of 75,000 ac-ft. In addition, a large increment to the basin is believed to be contributed by underflow from Egin Bench, an irrigated area east of the basin. Stearns and others estimated that that contribution in 1920-27 averaged not less than 140,000 ac-ft yearly.

All the estimates were crude and are outdated because of subsequent developmental use of additional ground water, but they show that the Mud Lake basin is an important hydrologic component of the Snake River Plain. The unconsumed annual increment of ground water to the basin is on the order of about 215,000 ac-ft/yr.

from adjacent intermontane valleys and highlands and from adjacent parts of the plain. Areas that contribute tributary underflow include the eastern slope of the Big Lost River Range, the Lemhi Range, the Beaverhead Mountains, the Centennial Range south of the Salmon River and Continental Divide, intervening valleys, and the Mud Lake basin.

Underflow from Nearby Valleys

The Big Lost River Valley above Arco is one source of recharge to the Snake River Plain. Water discharged by underflow along the channel of the river, as well as deep percolation losses from irrigation, joins a local body of ground water in the valley and moves by underflow into the plain. The water passes south of the NRTS without contributing directly to wells on the Station, but it helps to maintain storage in the regional ground-water reservoir. Stearns and others (1938, p. 245) estimated that the average contribution of ground-water underflow from the Big Lost River Valley to the plain during the period 1920-27 was 226,000 ac-ft yearly. The estimate was based largely on streamflow depletion and estimates of evapotranspiration. Very likely the actual contribution is nearer 300,000 ac-ft.

Underflow from the valleys of the Little Lost River and Birch Creek directly replenishes the ground water beneath the NRTS. Ground water from the valley of Little Lost River enters the Station beneath the central part of its northwest boundary, and underflow from Birch Creek enters from the north. Stearns and others (1938, p. 232, 233) estimated that the average annual contributions of ground water to the Snake River

The above estimates were derived by over-simplified methods and they are indicative only. They may be revised when longer records of streamflow are available and when other pertinent data are obtained. Most of the river water reaches the zone of saturation in parts of the NRTS where it cannot be recovered directly by existing wells. Nevertheless, the water helps to maintain regional ground-water storage and is, in effect, a contribution to the NRTS supply.

Ephemeral Drainageways and Playas

Small playas and undrained depressions are scattered widely over the NRTS and most of these contain ponded water after occasional periods of flash runoff. Records of measurements in three playas along the Big Lost River channel show the general magnitude of rates of infiltration in the playas (see p. 30-35). The data are important mainly because they show that infiltration does occur in the playas despite the low permeability of the sediments. The observed infiltration occurred during transient conditions that will recur only rarely, so no special computation was made of recharge in playa areas. The small average amount of yearly recharge from the playas was included automatically in the previous estimates of average direct infiltration of precipitation.

Recharge by Underflow

Recharge by underflow is by far the larger source of ground-water replenishment in the vicinity of the NRTS and perhaps in the whole Snake River Plain. Ground water beneath the NRTS is replenished by underflow or was dissipated by evapotranspiration between station 1 and Big Lost River playa 3. Some transpiration occurred in the playa areas and in a few places along the river channel where soil and gravel retain sufficient soil moisture to support sparse growths of willow, wildrose, and cottonwood. But most of the evapotranspiration loss was by evaporation. Evaporation from the river and transpiration along its borders in the 40-mile reach in the NRTS is assumed to have averaged 0.1 ft/day. The area of the evapotranspiration belt is about 200 ac. The estimated evapotranspiration loss was about 20 ac-ft/day, or slightly more than 8,500 ac-ft during the 426-day period. The residual discharge of the river during that period — about 128,000 ac-ft (39 billion gallons) — presumably became ground-water recharge.

The mean discharge of the river past station 1 during the 9-year period of record (see table 1) was about 58 cfs (see p. 14). None of the water escaped from the NRTS by surface runoff. The channel at station 1 is in basalt and ungaged underflow past the Station probably is negligible. Ordinarily, the river does not extend much beyond the old diversion dam in sec. 6, T. 2 N., R. 9 E. Evapotranspiration loss in the reach of a few miles below the Station is probably about 3,500 ac-ft/yr. The estimated average yearly contribution of the Big Lost River to ground-water recharge below station 1 during the period 1947-55 was about 42,000 ac-ft, and the average daily contribution was about 115 ac-ft — substantially more than the total amount of water for which the Commission had filed application for water rights.

Where basalt is exposed (no soil cover), water from heavy rain and snowmelt enters the ground rapidly, and the proportion reaching the zone of saturation is higher than in soil-covered areas. Depending on the seasonal distribution and storm-intensity of precipitation, and on the snow cover and its melting rate, ground-water recharge on the Station may range from nil in some years to an inch or more in other years. Assuming an average of 0.25 in. yearly, the volume of direct recharge on the entire station (about 44,000 ac) is about 900 ac ft, equivalent to a continuous flow of 1.25 cfs. This is about equal to the estimated consumptive use by NRTS facilities in 1955 and about 30 pct of estimated gross pumpage in that year.

Recharge from Surface Drainageways

Big Lost River

Upstream runoff in the Big Lost River is stored in Mackay Reservoir and is diverted above Arco for irrigation. Excess runoff which spills over the dam is augmented by return flow from irrigated areas below the reservoir and is discharged onto the plain. Some of this water evaporates, but a large percentage sinks into the river bed and becomes ground-water. The river loses water in its entire course across the plain below the irrigated area and eventually disappears.

The gage record at station 1 (table 1, appendix 2) shows that from August 1, 1951 to September 30, 1952, a period of exceptionally high runoff (see figure 2), the river discharged about 136,960 ac-ft (44.2 billion gallons) of water past the Station. The river channel is permeable throughout the NRTS, and all the water either sank into the ground during spring thaw periods when melting snow supplies water to the ground rapidly and in excess of the soil-moisture requirement, and during occasional torrential summer rains that produce excess water.

The gravel deposits in the NRTS contain fine material in their matrix, but the field moisture capacity is relatively small and may be exceeded at times by the supply of water available. In gravel areas, as on the alluvial fan of Birch Creek and the floodplain of the Big Lost River in the southern NRTS, deep percolation occurs when a relatively large or sudden supply of water from rain and melting snow is available. The ability of the gravel to absorb and transmit water is well shown by studies of the loss of water by infiltration in the channel of the Big Lost River.

The moisture capacity of windblown soil, lake and playa beds, and dune sand is high, because the porosity of these materials ranges from 30 to 60 pct. A thick layer of fine-grained sediments like that in the Birch Creek playa and ANP area (up to 80 ft or more in thickness) could absorb as soil moisture all precipitation that is available in that area. Evapotranspiration probably is about equal to precipitation and very little water is apt to percolate through the sediments and become ground water. Tests of samples from borings in the Birch Creek playa show that the moisture content is very low at depths of 15 to 20 ft. The evidence of desiccation of the playa sediments (see part 2) is another indication that the moisture supply is not enough to satisfy the field capacity, much less to provide ground-water recharge.

Recharge from Precipitation

The first call on water applied to the land, whether by precipitation or otherwise, is for restoring soil moisture, which evaporates or is transpired by plants. Water percolates downward to the zone of saturation only where or when there is an excess over the soil-moisture demand. Precipitation on the plain is low and is rather well distributed through the year. Where there is soil cover on the basalt, most precipitation becomes soil moisture, and very little is left over for recharge.

The normal precipitation, averaged for the Snake River Plain as a whole, is about 10 in. a year. Potential evapotranspiration is about 25 in., but the actual amount never equals the potential because not enough water is available. On the other hand, not all precipitation is consumed, because temporary excesses of moisture occur. From studies made elsewhere, the writer believes that about 0.75 in. of the precipitation on the plain becomes ground-water recharge yearly. If so, in the 17,000 sq mi (10,900,000 ac) of the Snake River Plain the total volume of recharge directly from precipitation would be about 681,000 ac-ft annually.

The assumed average amount of recharge cannot, of course, be applied to specific areas on a unit basis. On the NRTS, for example, precipitation is 7.5 in. per year — barely enough to keep the native vegetation alive. During warm months potential evapotranspiration far exceeds precipitation, and rain in warm months probably does not cause recharge in areas where there is soil cover. During winter months the ground is frozen. Moisture transfer in and through frozen ground probably is almost entirely in the vapor phase, so the principal recharging of ground water occurs

GROUND-WATER HYDROLOGY

The Snake River Plain is the gathering ground for water contributed by a drainage area more than four times the size of the plain. The contributing area includes most of the 35,800 sq mi in the drainage basin of the Snake River and its tributaries east of King Hill. The plain is about half that area, the rest being mountains and tributary valleys.

Precipitation is the ultimate source of all water in the basin (ignoring the small amount of juvenile water that may rise from depth.) Low rates of precipitation prevail throughout the basalt plain, but the rates in the surrounding area, especially the mountains, are much higher (see p. 8). Considerably more than half the water that ultimately reaches the plain originates as precipitation on adjoining highlands.

RECHARGE

Some ground water beneath the plain is derived from direct infiltration and deep percolation of precipitation. Some is from infiltration of surface water along the channels and in flooded areas of surface streams. Some is contributed by underflow from valleys adjoining the plain. And some is acquired by mass percolation from mountain areas adjacent to the plain. On the NRTS, recharge occurs directly from precipitation; by infiltration from the Big Lost River and from ephemeral and intermittent drainageways and playa basins; by underflow from the valleys of the Big Lost and Little Lost Rivers and Birch Creek; by underflow from the north and northeast; and by percolation from adjoining foothills. The contributions from each source differ widely in amount, but only a general quantitative evaluation can be made at this time.

Differences in texture among the various layers of a sedimentary sequence markedly affect their permeability, especially in a direction perpendicular to the bedding planes (vertical permeability). This is illustrated by the two samples from T. 2 N., R. 29 E., SE¼NE¼, S. 1 (see table 17, p. 122). The permeability in a direction parallel to the bedding (horizontal permeability) usually exceeds the vertical permeability. Unfortunately, these samples were disturbed and the illustration is not spectacular. At infiltration test site number 4 (T. 5 N., R. 31 E., SW¼, S. 9) a thin silt lens was observed at the depth of sampling. A sample (sample 55IDA20) of the upper sand had a permeability three times as great as a sample of the same sand with contained silt lens (sample 55IDA20a). The variations in vertical permeability are illustrated by tests of samples from successive depths at several locations (figure 15). These variations in the permeability of various layers in stratified sediments are a principal obstacle to accurate calculation of infiltration rates in the zone of aeration.

GPO 991857

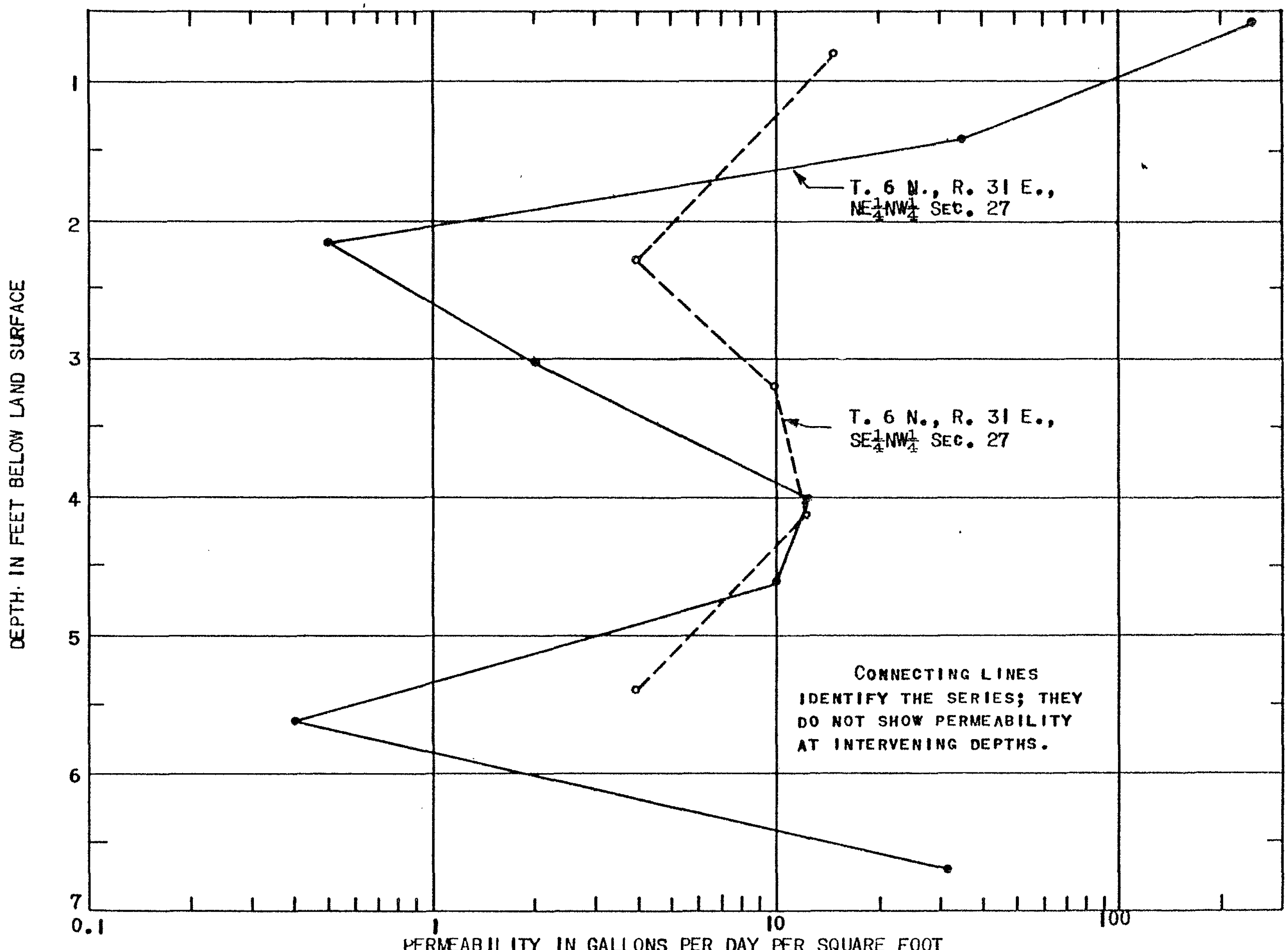
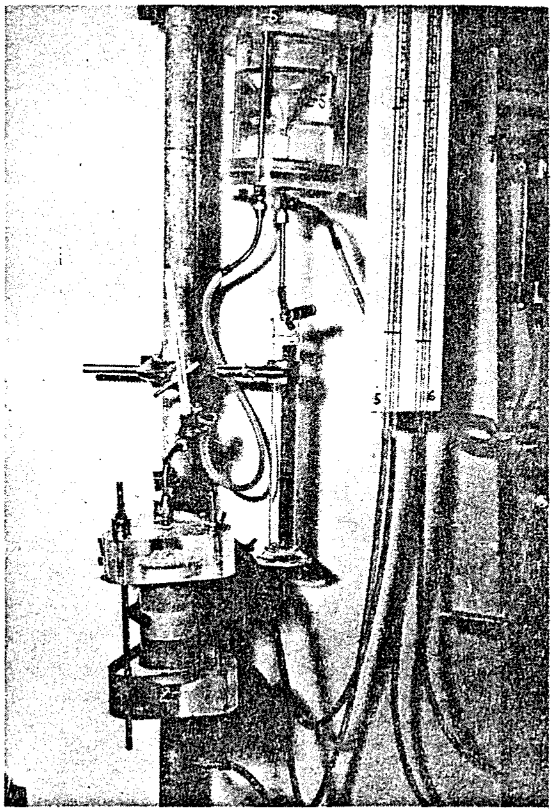


Figure 15.—Variations in permeability with depth in two series of samples.



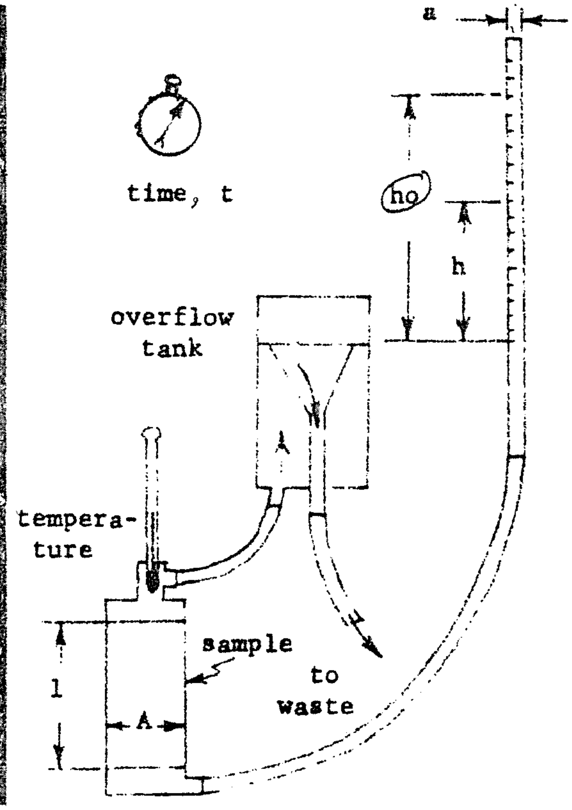


Figure 14. — Variable - head permeameter.

The greatest permeability recorded exceeded 400 gpd per sq ft and the least was 0.005 gpd per sq ft. In general, the playa sediments have very low permeabilities -- 10 gpd per sq ft or less, and the loess is only slightly more permeable. The Terreton lake bed sediments have medium-range permeabilities. Dune sands probably are the most permeable of all the fine-grained sediments.

The apparent permeability of fine-grained samples that were disturbed sampling are abnormally high because water moved mainly through fractures rather than through the natural pores. Some of the clay samples that were moist when collected dried and shrank before they were tested. The annular space between the shrunken sample and the enclosing cylinders was an unnatural avenue for ready movement of water.

The coefficients of permeability of the fine-grained sediments were determined by means of variable-head permeameters. The variable-head permeability method involves measurements of the rate of fall of water level in a manometer connected to a sample through which water is percolating. The basic formula for the use of the variable-head permeameter is obtained by integrating Darcy's Law,

$$k = 2.3 \frac{aL}{At} \log \frac{h_o}{h} C_T$$

in which k is the coefficient of permeability, h_o is the head in the manometer at zero time, h is the head at the end of any given elapsed time, t is the elapsed time, A is the area of the sample cylinder in square centimeters, a is the area of the manometer, L is the length of the sample in centimeters and C_T is the dimensionless ratio of the viscosity of water at the observed temperature to the viscosity at 60°. Using meinzer units, the above formula becomes,

$$k = 48,815 \frac{aL}{At} \log \frac{h_o}{h} C_T$$

in which k is in gallons per day per square foot under a hydraulic gradient of one foot per foot at 60°, A and a are in square centimeters, L is in centimeters, t is in seconds, h_o and h are in centimeters, and C_T is dimensionless.

As with coarse-grained sediments, de-aired water was used as the percolation liquid. The coefficient reported here is the maximum value observed after several runs and represents saturated permeability. The variable-head permeability apparatus used in the Hydrologic Laboratory is illustrated in figure 14.

Table 17.--Laboratory permeability of core samples of fine-grained sediments--Continued

Location of source (township, range, and section)	Depth (ft below land surface)	Field sample number	Laboratory sample number 1/	Description of material	Coefficient of permeability (meinzers)	Remarks
T. 3 N., R. 32 E.						
SE¼SE¼, S. 29	0.4	7V	53IDA4	Soil mantle	120	Slightly disturbed
T. 2 N., R. 29 E.						
SE¼NE¼, S. 1	13.0	1H	53IDA1	Silt, clayey	0.6	Horizontal sample; greatly disturbed
SE¼NE¼, S. 1	13.0	2V	53IDA2	Silt, clayey	0.4	Vertical sample; greatly disturbed
NW¼SW¼, S. 3	1.0		55IDA22	Loess	17	
T. 2 N., R. 30 E.						
NW¼NW¼, S. 15	0.4	14V	53IDA8		30	
T. 2 N., R. 31 E.						
SE¼NE¼, S. 16	0.4	13V	53IDA7		45	Slightly disturbed
SW¼SE¼, S. 35	0.6	11V	-		6	
SW¼SE¼, S. 35	0.5	12V	53IDA6		34	

1/ Laboratory numbers were assigned to samples tested in the Hydrologic Laboratory, Lincoln, Nebraska. No laboratory numbers were assigned to samples tested at the temporary field laboratory.

Table 17.--Laboratory permeability of core samples of fine-grained sediments--Continued

Location of source (township, range, and section)	Depth (ft below land surface)	Field sample number	Laboratory sample number 1/	Description of material	Coefficient of permeability (meinzers)	Remarks
T. 6 N., R. 31 E.						
SE¼NW¼, S. 27	0.8	22V	53IDA11		15	
SE¼NW¼, S. 27	2.3	23V	53IDA15		4	
SE¼NW¼, S. 27	3.2	24V	53IDA12		10	
SE¼NW¼, S. 27	4.1	25V	52IDA13		12	
SE¼NW¼, S. 27	5.4	26V	53IDA13		4	
SW¼, S. 34	2.0		55IDA19	Clay, silty	1	
SE¼SW¼, S. 9	6.0		55IDA20	Silt, clayey	138	
SE¼SW¼, S. 9	6.0		55IDA20a	Sand, silty	54	
NW¼, S. 20	Surface		55IDA23	Dune sand	190	
T. 5 N., R. 30 E.						
SE¼NE¼, S. 2	0.4	35V	52IDA9		0.03	
SE¼NE¼, S. 2	0.3	36V	52IDA4		0.005	
SE¼NE¼, S. 2	1.0	37V	52IDA5	Sand, med.	420	
NE¼NE¼, S. 3	1.0	3V	52IDA1	Silty, clayey	22	
NE¼NE¼, S. 3	1.6	4V	-	Sand, fine	17	
NE¼NE¼, S. 3	2.5	5V	53IDA3	Silt	41	
NE¼NE¼, S. 3	3.5	6V	-	Sand	120	
NE¼NE¼, S. 3	4.4	20V	-		1	Disturbed
NE¼NE¼, S. 3	4.5	21V		Sand, w/silt layer	9	Disturbed
NE¼NE¼, S. 3	0.4	8V			0.3	
NE¼NE¼, S. 3	0.4	9V	53IDA5		52	Slightly disturbed

Table 17.--Laboratory permeability of core samples of fine-grained sediments

Location of source (township, range, and section)	Depth (ft below land surface)	Field sample number	Laboratory sample number 1/	Description of material	Coefficient of permeability (meinzers)	Remarks
T. 6 N., R. 31 E.						
SW¼NE¼, S. 13	0.6	38V	52IDA6	Clay, silty	80	Greatly disturbed
SW¼NE¼, S. 13	1.2	39V	52IDA7	Clay, silty	22	Greatly disturbed
SW¼NE¼, S. 13	1.8	40V	52IDA8	Clay, silty	4	Greatly disturbed
SW¼NE¼, S. 13	3.0		55IDA18	Clay, silty	10	
SW¼NW¼, S. 24	0.6	34V	53IDA16		38	
SE¼NW¼, S. 24	0.6	19V	53IDA10		3	Slightly disturbed
SW¼NE¼, S. 27	1.0		55IDA21	Clay, silty	12	
NE¼NW¼, S. 27	0.6	18V	-		220	
NE¼NW¼, S. 27	0.6	15V	52IDA12	Sand, silty	160	
NE¼NW¼, S. 27	1.0	16V		Dune sand	180	
NE¼NW¼, S. 27	1.4	27V	52IDA3	Sand, silty	38	
NE¼NW¼, S. 27	2.2	28V	53IDA14		0.5	Slightly disturbed
NE¼NW¼, S. 27	3.1	29V	-		2	
NE¼NW¼, S. 27	4.0	30V	52IDA10	Clay, silty	12	
NE¼NW¼, S. 27	4.6	31V	52IDA2	Clay, silty	10	
NE¼NW¼, S. 27	5.6	32V	52IDA11		0.4	
NE¼NW¼, S. 27	6.7	33V	-		32	
SE¼NW¼, S. 27	0.8	17V	53IDA9		30	

1/ See footnote at end of table

seepage losses from the Big Lost River in similar sediments. Auger sampling showed that the water from the infiltrometer spread laterally through a wide zone. The volume of the crudely hemispherical body of sediment which was wetted beneath infiltrometer site 4, for example, was approximately 1,000 cubic feet. At the calculated porosity, it would require most of the water that entered the ground just to moisten the earth, leaving little or none to percolate downward by gravity. Much of the water was pulled by capillarity away from the column of earth immediately under the cylinder. Not until the sediments were saturated would the observed infiltration rate approach the permeability. Most of the tests were operated for about a week, whereas a test of moderately fine material probably would have to be run for several weeks to get a true infiltration rate. In very fine-grained material an infiltrometer test might require months.

Permeability

Relatively undisturbed samples of a wide range of the fine-grained sediments from the NRTS were tested in the laboratory. The samples were taken with specially constructed coring equipment in 2-inch diameter cylinders. The cylinders were capped, sealed and transported to the laboratory where the samples were tested intact in the cylinders. The results of the laboratory tests are summarized in table 17. Some of the samples were unavoidably disturbed during coring and transporting, and these are suitably noted in the table.

Table 16.—Infiltration rates through double-ring infiltrometer

Location	Test number	Nature of material	Infiltration rate ft/day	Infiltration rate gpd/ft^2	Remarks
T. 6 N., R. 31 E., NW¼ Sec. 13	1	Silty clay, calcareous; in Birch Creek playa near ANP site	1.0	7.5	Inside ring
			3.2	24	Outside ring
			2.8	21	Both rings
SW¼ Sec. 34	3	Silty clay, calcareous; in Big Lost River playa 3	1.8	13	Both rings (short time)
			9.5	71	Both rings (long time)
T. 5 N., R. 31 E., SW¼ Sec. 9	4	Fine sand, silty, and silt, clayey; Terreton Lake beds	5.7	43	Both rings
T. 4 N., R. 30 E., SW¼ Sec. 29	5	Fine to coarse gravel, sandy, with some calcareous cement; Big Lost River alluvium in borrow pit E. of STR	41	306	Both rings

Below the cylinders the bulb of wetted earth during each test was much wider than the diameter of the infiltrometer. Test no. 4 was made in the bottom of a borrow pit 4 feet deep in the SE$\frac{1}{4}$SW$\frac{1}{4}$ Sec. 9, T. 5 N., R. 31 E. The earth materials probably are Terreton lake beds, consisting of very fine-grained silty sand which is densely compacted and dry. These beds overlie basalt at a depth of about 7.6 feet below the natural land surface. The natural moisture content of the soil at infiltration test site 4 was about 12 percent, depending on depth. The moisture content at the outside of the lower end of the cylinder rose to 18 percent, and the 18-percent belt extended downward to a point 3.5 feet below the cylinder and laterally 4 feet from the center line of the cylinder.

The results of several infiltration tests are summarized in table 16. Test number 5 is included to compare infiltration rates in gravel with those in finer sediments. The data indicate that the infiltration rates would be exceedingly low in the fine-grained materials that cover much of the northern one-third of the Station, but they would be relatively high in the Big Lost River alluvial gravels in the central and southern part of the Station. The dune sands that mantle much of the northern one-third of the Station would have moderate infiltration rates. No infiltrometer tests were made in the medium-grained sediments of the central part of the Station, but laboratory permeability tests indicate that the infiltration rates would be on the order of 25 to 50 ft per day.

These results are not positive indications of infiltration rates that may be expected in seepage ponds for waste disposal. In all cases the rates are higher than those computed from measurement of natural below the land surface. The rings received water from a two-tank trailer; float valves maintained the water in both cylinders at constant levels. The flow of water was calculated from the record charts of water-level recorders in each tank. The head of water in the cylinders was slightly less than 5 ft. Covers over the tank trailer and the cylinders reduced evaporation.

Vertical flow through inhomogeneous sediments is divergent, but the basic theory applicable to infiltrometer tests requires that downward flow into and through the sediments be nondivergent. Where a subsurface horizon is less permeable than the horizon in or directly beneath the infiltration cylinders, flow will be divergent. Many investigators have believed that divergent-flow effects can be minimized if the infiltration cylinder is surrounded by a larger guard cylinder, thus the term double-ring infiltrometer. The infiltration rate from the inner cylinder supposedly is a truer rate than that from the outer cylinder. Other investigators have found no significant difference between infiltration rates obtained with the single-ring infiltrometer and double-ring infiltrometers. However, these investigators usually carried on their tests in uniform soils where there was small chance for divergent flow. As the question is controversial, the double-ring infiltrometer was used in the belief that the guard ring at least does not disturb a test. The results show that the guard ring does have an effect, but which infiltration rate to accept (inner or outer) remains controversial.

We tested a moderate range of sediment types at wide intervals on a north-south line through the northern NRTS. Determinations of the moisture content were made of soil samples from each infiltrometer site to show the soil-moisture distribution before and at the end of the test.

All the fine-grained sediments are potentially useful, in varying degrees, as media for the disposal of liquid waste and for the burial of solid waste. Their chief natural hydrologic function, however, is to store soil moisture, most of which is transpired and evaporated.

Whereas the coarse-grained sediments were favorable for ready infiltration and percolation of water, the fine-grained sediments are much less favorable. Where only fine-grained materials are present, special problems arise in liquid-waste disposal, drainage of road subgrades, mud control, stability of materials under structural loads, and other factors.

The rates of infiltration in the playas were summarized on pages 30-35. They indicate only the general order of magnitude of rates that might be expected in artificial infiltration ponds in those areas if the disposed liquid did not chemically alter the fine-grained clayey sediments and reduce their permeability.

Hydraulic Properties

Infiltration rates for fine-grained sediments were determined in the field by means of infiltrometers. The permeability of these sediments was tested in a temporary field laboratory and in the hydrologic laboratory. The tests will be described briefly with a tabulation of the results.

Infiltrometer Tests

Several field infiltration tests were made in 1955 with double-ring infiltrometers. The apparatus was a 6-in. steel cylinder centered in an 18-inch cylinder, both 5-ft long. It was installed with its bottom 4 ft

The maximum permeability observed during a test (saturation permeability) probably is near the field permeabilities of material in the zone of saturation, so we report that value as the permeability of a sample. However, the permeability of the repacked sample may differ appreciably from that of the undisturbed material in nature.

The surficial sediments in the NRTS seldom are saturated except in the upper few inches or for short times. Therefore, the saturated permeability of a sample in the laboratory represents no common natural field condition in the zone of aeration, and the laboratory value can not be used directly to compute the rate at which infiltration by unsaturated flow would occur. The permeability varies directly with the degree of saturation, and values decrease at saturations of less than 100 percent. Nevertheless, it is desirable to have a uniform basis for comparing the permeability of various materials. As it is difficult to determine permeability at a predetermined degree of saturation, the saturated permeability of each sample is a value that can be defined and reproduced.

FINE-GRAINED SEDIMENTS

The playa deposits and Terreton lake beds that overlie the Snake River basalt in the northern part of the NRTS are fine grained and their permeability is low. Nevertheless, the material is porous and the sediments can slowly absorb substantial amounts of water. The loess that mantles the basalt and alluvium in much of the NRTS is slightly more permeable than the playa and lake beds. The windblown fine sand, which is prevalent in the northern part of the Station is moderately permeable.

Factors that affect permeability measurements.--If readings are taken at least every hour during permeability determinations, the following events usually occur. During a short period, the apparent permeability increases to a maximum value. Thereafter, the permeability decreases, finally "leveling-off" at a minimum value. This general trend persists during several to many days. Also, the permeability usually fluctuates during each day.

The permeability increases while air is being absorbed from the sample, reaching a maximum when all gas is absorbed and saturation of the sample pores is complete. The air or other gases trapped in a sample plugs pore spaces and reduces the apparent permeability. A vacuum system is used to de-air the tap water used as the percolation fluid. The de-aired water absorbs air and carries it out of the sample. The pore spaces in coarse-grained sediments are generally large enough that much of the air is flushed out of the sample mechanically rather than being absorbed in the fluid. Therefore, the maximum permeability for coarse-grained sediments usually is reached very early in the test.

Maximum permeability is reached when all gases have been removed and the sample is completely saturated. However, by that time other factors tend to reduce the permeability. For example, the water may contain enough solid matter of microscopic size to plug some pores in the sample or in the filter disks at each end of the sample. Some of the finer particles of the sample migrate and plug pores. Small amounts of dissolved substances in the water may react chemically with the sample, especially if it contains an appreciable amount of colloidal clay particles. The development and growth of microorganisms also would decrease the permeability.

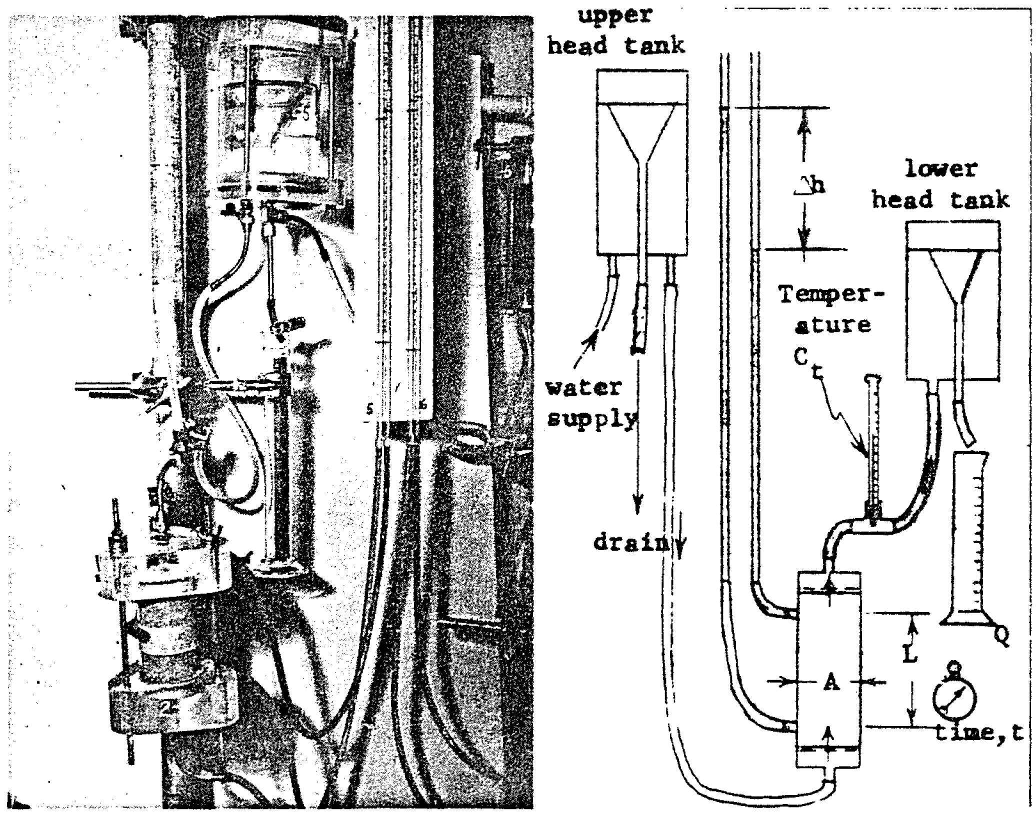


Figure 13. — Constant -head permeameter.

The disturbed samples were packed into the permeameter cylinders with a jolting-type packing machine. (Denver Hydrologic Laboratory.)

Laboratory method of analysis.--Permeability is determined in the laboratory by observing the rate of percolation of water through a sample of known length and cross-sectional area. The coefficients for samples of the coarse-grained sediments were determined by means of constant-head permeameters in which the flow of water was vertically upward through the material (fig. 13).

The constant-head method requires that the rate of discharge through a sample be measured while the difference in head of water at the top and bottom of the sample is held constant. From Darcy's Law, the basic formula for the constant-head permeameter is

$$k = \frac{QL}{At(\Delta h)} C_T \qquad (6)$$

where k is the coefficient of permeability, Q is the volume of percolation, L is the length of the sample, A is the cross-sectional area of the sample cylinder, t is the length of the period of flow, Δh is the difference in head at the top and bottom of the sample and C_T is a temperature correction to convert the permeability at any given temperature to that at 60 degrees Fahrenheit. Expressing permeability in meinzers, the previous formula becomes

$$k = \frac{21,200\ QL}{At(\Delta h)} C_T \qquad (7)$$

in which k is in gallons per day per square foot under a hydraulic gradient of one foot per foot at 60 degrees Fahrenheit, Q is in cubic centimeters, L and Δh are in centimeters, A is in square centimeters, t is in seconds, and C_T is the ratio of the viscosity of water at the observed temperature to the viscosity at 60 degrees Fahrenheit.

Table 15.--Laboratory permeability of repacked samples of coarse-grained sediments

Location of source (township, range and section)	Description of location	Depth (ft below land surface)	Laboratory sample number	Coefficient of permeability, (gal/day/sq ft)
T. 3 N., R. 29 E., SE$\frac{1}{4}$, Sec. 14	Foundation excavation, MTR	23	55IDA25	460
SW$\frac{1}{4}$SW$\frac{1}{4}$, Sec. 34	Gravel pit, north of EBR	6	55IDA27	560
T. 4 N., R. 30 E., SW$\frac{1}{4}$, Sec. 15	Gravel pit, Big Lost River	8	55IDA26	700
SW$\frac{1}{4}$, Sec. 29	Gravel pit, east of STR	8	55IDA24	780

gradient of one foot per foot at a temperature of 60 degrees Fahrenheit (the unit is sometimes called a meinzer). Because water is specified, viscosity and density effects are neglected.

Coefficients of permeability, ranging from 0.00001 to 30,000 gallons per day per square foot, have been measured in the laboratory. The value of the coefficient depends in general upon the degree of sorting and upon the arrangement and sizes of particles. Usually it is low for clay and other fine-grained or tightly-cemented materials, and high for coarse, clean gravel. Few productive natural water-bearing materials have coefficients less than 10, and for most it is above 100.

Four disturbed channel samples of coarse-grained sediments were shoveled from exposed vertical surfaces in several gravel pits and in a foundation excavation in the Big Lost River alluvium. The results of laboratory tests of repacked samples are reported in table 15. The data show that the permeability is good but not exceptional, ranging from 460 to 780 gallons per day per square foot. The permeability and other properties of these sediments favor ready infiltration and percolation of water or other fluids.

Infiltration Rate in Channel of Big Lost River

High runoff in the Big Lost River through the NRTS during many months in 1951-53 afforded opportunities to observe and measure natural infiltration rates in material ranging in texture from coarse gravel to fine silt. The observed rates of infiltration probably are within the range that may be expected in infiltration basins in sediments like those in the channel. Computations from field measurements were summarized on page 24. The rate in the infiltrometer described above was several times greater than any rate observed in the natural river channel.

Permeability

Hydraulic permeability is the capacity of a porous substance to transmit water under pressure. The basic law governing the flow of fluids through porous media was established by Darcy, who demonstrated experimentally that the rate of flow is proportional to the hydraulic gradient. Darcy's Law may be expressed in useful form as

$$q = kiA \qquad (4)$$

in which A is the total cross-sectional area through which a rate of flow, q, occurs; i is the hydraulic gradient (the head, h, divided by the length of flow, l); and k is the coefficient of permeability, which will be defined later. It follows that

$$k = \frac{q}{iA} \text{ or } \frac{ql}{hA} \qquad (5)$$

For ground-water hydraulics, the Geological Survey defines the coefficient of permeability as the rate of flow of water, in gallons per day, through a cross-sectional area of one square foot under a hydraulic

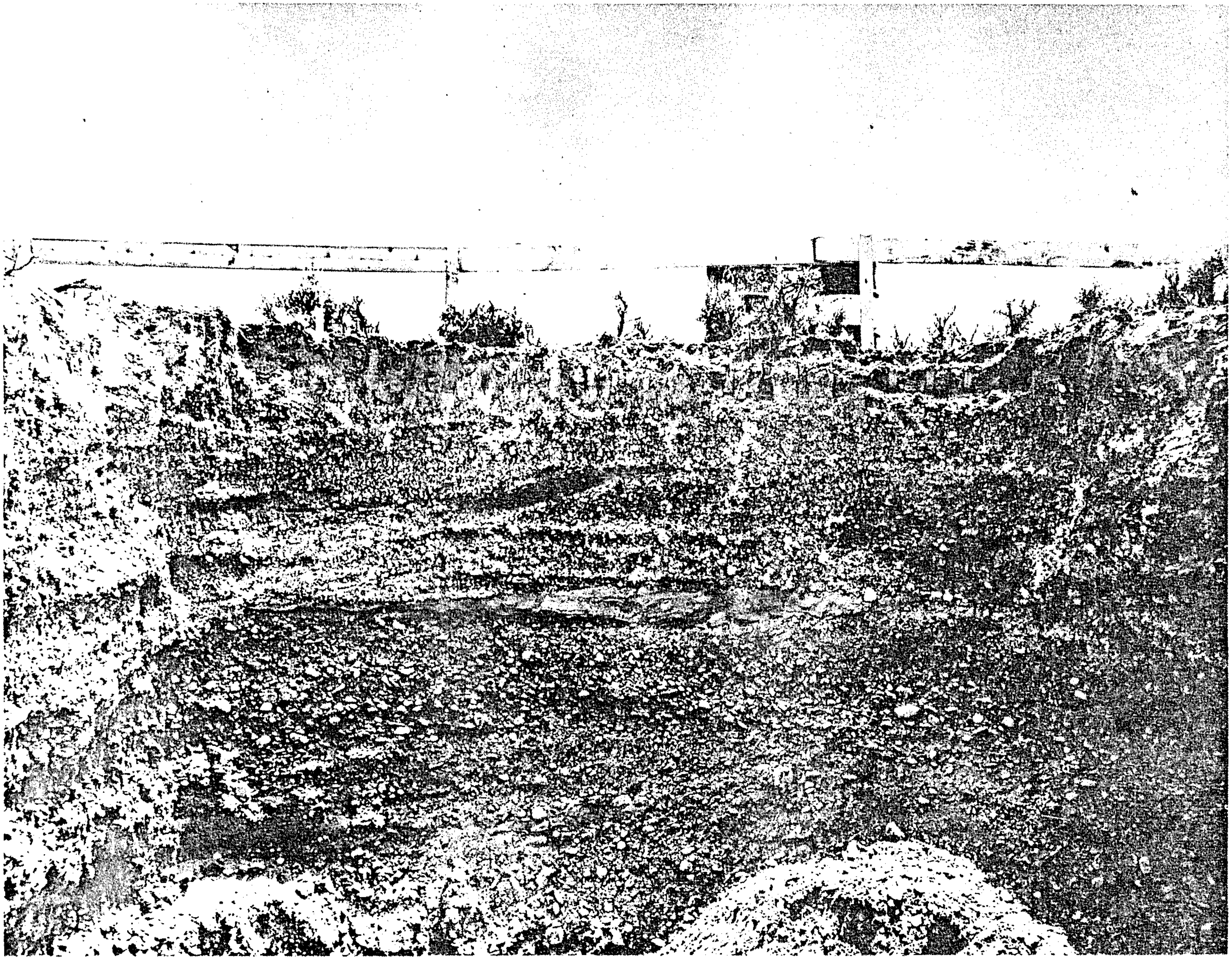


Figure 12.—Gravel exposed in pit where infiltration test was made near MTR plant

saturation to develop in the gravel, and when that occurred the rate of infiltration might be reduced. The rate in the infiltrometer was much higher than any noted during study of infiltration in the channel of the Big Lost River.

The infiltrometer test was improvised because information was needed immediately at the MTR site. It also helped to define conditions and factors that must be contended with in further tests. Infiltration rates depend on the physical nature and condition of the earth material and the character of the water, and these may change as infiltration and percolation proceed. Among the more important factors are the internal physical structure and composition of the sediment and the condition of its surface; the chemical and mineralogic composition of the sediment, including the organic contents; the amount of entrapped air retained during infiltration; the chemical quality and turbidity of the water; the temperature of the water and the sediments; the distribution and amount of soil moisture, or moisture tension; the barometric pressure; the kind of equipment used; the hydraulic head applied; and the duration of the test.

Waste water discharged to infiltration pits on the NRTS is a chemical solution which is markedly different from the natural water, and the sediments on the NRTS, especially the fine-grained ones, are susceptible to mineralogic and chemical change from reaction with waste water. Some infiltration tests should be made with water that is chemically comparable to the wastes that are discharged to the ground. Some tests should be made also in small pits or ponds that simulate operating conditions.

the dug hole 8 ft south of the cylinder, showing that there was lateral movement and development of a saturated zone of earth beneath and around the cylinder. At the end of 26 hours the water level in the dug hole reached its highest level, 0.25 ft lower than the bottom of the cylinder. Thereafter , the water level in the hole declined and after 66 hours the hole was dry.

The total amount of water that passed through the core was 115,970 gallons. Excluding the early part of the test, before midnight on May 21, the amount of water was 67,020 gal in 8.5 days, and the average rate of percolation in that period was 5.5 gpm (7,900 gpd, or 630 gpd/ft^2).

"Falling head" observations of the water level in the cylinder after the test was shut down confirmed the computation of the average, giving an infiltration rate of 5.6 gpm at the end of the test.

A plot of the small fluctuations of hydraulic head over the gravel core disclosed no obvious direct correlation with fluctuations in the rate of flow. Temperature appreciably affects water viscosity, and the principal variations in the infiltration rate after May 21 very likely were caused by temperature fluctuations in the water. The higher rates occurred during the midpart of each day, when the temperature of the exposed storage tank was highest.

The average flow rate of 630 gpd/ft^2 would be a poor guide for designing an infiltration pond in gravel like that at the MTR site. The saturated permeability of the material with all ground air excluded might be somewhat higher than that indicated. On the other hand, layers of nonpermeable fine-grained material occur in the gravel (fig. 12). Such a layer beneath a large basin might cause a perched zone of

GPO 991857

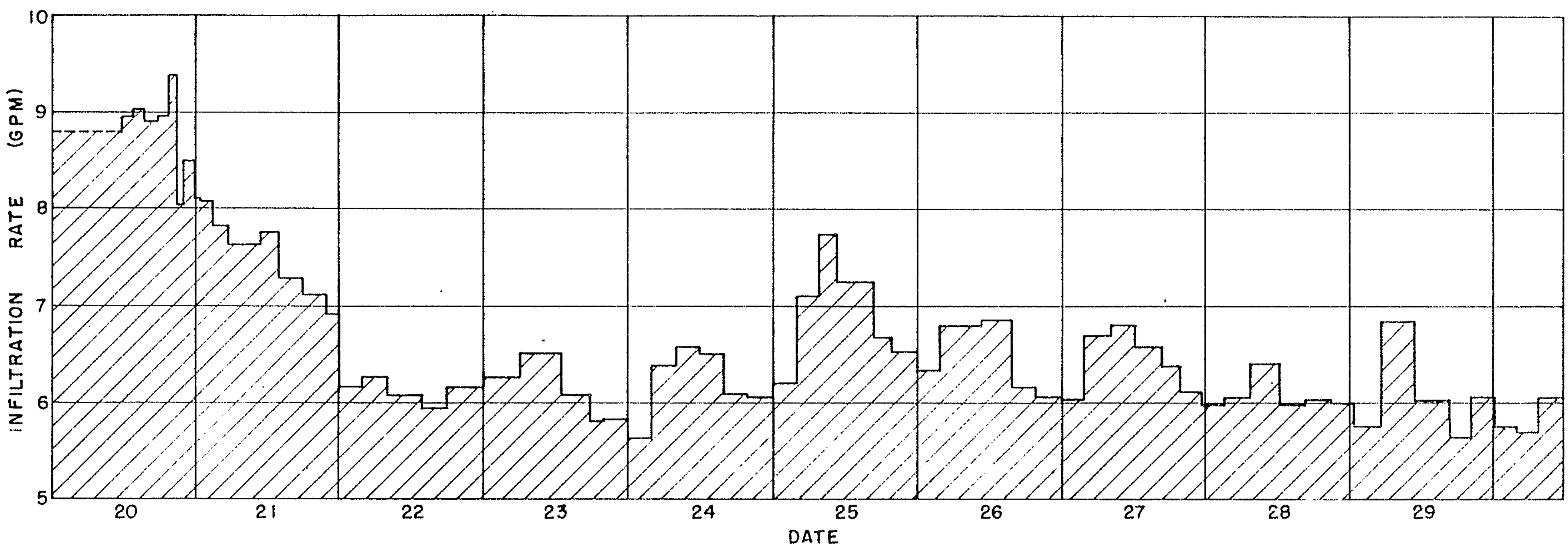


Figure 11.—Infiltration rate through undisturbed gravel core in concrete ring infiltrometer, May 19–31, 1950.
(Not adjusted for temperature changes.)

to lubricate the cylinder and to seal it to the core. When the cylinder reached the desired depth, a seal of asphalt emulsion was placed around the outer side of its bottom edge to prevent piping by water, and the space around the cylinder was backfilled with tamped soil and gravel. The cross-sectional area of the core was 12.6 sq ft.

A layer of graded, clean pebble gravel was spread over the core in the cylinder and covered with a perforated metal plate to reduce turbulence in the incoming water. Water was provided from a 1,000-gallon storage tank and provision was made for maintaining a reasonably constant head of water over the core. A canvas cover over the cylinder reduced evaporation, which was estimated to be only 38.5 gallons during the test, and was disregarded. Meter readings and head measurements from a staff gage in the cylinder were made about every hour during the test, which lasted 289 hours. No water-temperature data were obtained.

The bottom of a dug hole in the ground about eight feet south of the cylinder, was below the level of the cylinder bottom. The development of a zone of saturation around the infiltrometer was shown by the appearance of ponded water in this hole during the test.

The rate of flow through the core was relatively rapid during the first 48 hours (fig. 11), but thereafter it fluctuated about a norm of somewhat more than 6 gpm. The average rate of infiltration declined slightly during the last 5 days of the test. The first 25 minutes of the test produced enough saturation in the core to cause ponding of water and development of head in the free part of the cylinder. After about 35 hours the head was 13 in. Thereafter, the head fluctuated between 13 and 14.5 inches. At the end of 7.5 hours water appeared in reduce the permeability much. A discontinuous thin bed of nonpermeable clay and silt rests between the gravel and underlying basalt at some places. Perched zones of saturation may develop on the bed wherever excess water percolates through the gravel. Small permanent or ephemeral natural bodies of perched water may rest on the silt at some places. Water is perched on the silt and perhaps on tight basalt beneath part of the MTR-ETR site. The distribution of the silt bed beneath the Big Lost River flood plain has been delineated only locally, as at the MTR site, where it is 3 to 3.5 feet thick.

Hydraulic Properties

Preliminary experiments have been made on infiltration rates in the coarse-grained sediments. The permeability of coarse-grained sediments was not determined by well-performance tests ("pumping tests") but several samples of these sediments were tested for permeability in the Hydrologic Laboratory at Denver, Colorado.

Infiltration Through Infiltrometers

In May 1950 a crude infiltration test was made in gravel near the present site of the MTR disposal pond, using an improvised ring infiltrometer in the bottom of a dug pit. A concrete cylinder, 48 in. in diameter and 4 ft long, was lowered around a core of undisturbed material by setting the cylinder on the ground and digging around it, allowing the cylinder to settle and enclose the core. Cobbles that projected out of the core beneath the edge of the cylinder were removed and replaced with clay grout. Also, the surface of the core was plastered with clay

WATER-BEARING PROPERTIES OF THE SEDIMENTS

The surficial sediments that overlie the Snake River basalt in the NRTS lie far above the main water table. They contain ground water only in local perched zones of saturation, as was described on page 46. The yields of the perched aquifers to wells were not tested because these aquifers are not a prospective source of water supply for the NRTS.

The sedimentary interflow beds in the main zone of saturation yield water, but it was not practical to test their capacities separately from that of the basalt. Most of the sediments are poor water-bearers, are troublesome in well construction, and are best excluded from finished wells.

The surficial sediments can absorb a considerable amount of water, and coarse material transmits water readily. Thus, the sediments are an important factor in local ground-water recharge and in the disposal of liquid waste. Some of their hydraulic properties were tested and further study is planned.

COARSE-GRAINED SEDIMENTS

The alluvial sand and gravel in the flood plain of the Big Lost River, in the alluvial fan of Birch Creek, and in foothills alluvial fans absorb and transmit water readily. These properties are important factors in the regional regimen of ground-water recharge and in local provisions for the disposal of liquid waste. Much of the gravel contains secondary deposits of calcium carbonate at shallow depth, but at most places the amount of secondary minerals is small and does not

$$r = \frac{24.2\ Q^2}{Tk} \qquad (3)$$

where r is the optimum spacing, in feet, Q is the pumping rate in each well (wells assumed to be pumping at equal rates), in gpm, T is the coefficient of transmissibility, in gpm/ft, and k is the capitalized cost of pipeline and transmission lines between wells, in dollars per year per foot of intervening distance.

The optimum spacing of wells in the ANP area was computed using this formula. The coefficient of transmissibility determined from the results of aquifer tests on wells in the ANP area is about $8x10^5$ gpd/ft. Assume that k is $12.00 and is capitalized at 10 pct per year, or $1.20 per ft per year for capital charges, overhead, depreciation and maintenance. Assume further that the pumping rate of each of two wells is 1,000 gpm. Substituting these and other values given above in the formula, the economic-optimum spacing would be about 25 feet. The problem here has been over-simplified to illustrate the principles involved. In construction planning, other factors would have to be considered and undoubtedly would affect the "economic-optimum."

The chief significance of this computation is its demonstration that, under ordinary conditions on the NRTS, wells may be spaced as closely as is desired without creating excessive pumping cost. The result thus directly contradicts our earlier recommendations that wells be spaced at wide intervals of about 500 feet. When those recommendations were made they seemed prudent because little was known about the aquifer, and the optimum spacing could not be computed.

ft/ft, and a T value of 3.3×10^6 gpd/ft computed from aquifer-test data for well CPP-3 (see table 14), the permissible spacing is 580 feet. Although this indicates an apparent safety factor of about 3 in the existing spacing, it must be emphasized that this would be true only of a homogeneous aquifer. The real margin of safety, if any, in the existing setup is indeterminate.

The minimum permissible spacing varies directly as the rate of discharge. Further increases in the pumping and disposal rates should not be made unless a recalculation, based on the latest available hydraulic data, shows that the existing spacing would be adequate. The final test of "adequate spacing," of course, is whether or not any water is recirculated.

Economic-Optimum Spacing

Owing to the cost of constructing and maintaining power lines, pipelines, service roads, and to other factors, wells at a given site should be spaced as close together as is possible without excessive pumping costs.1/ The less the distance between wells, the greater their interference and the greater the drawdown and pumping lift. The greater the distance between wells, the less their interference but the greater the cost of connecting pipelines and electrical installation.

The cost of installing two wells, insofar as it is affected by their spacing, is computed by a formula derived by Theis (1957)

1/ For operational reasons, of course, it may be desirable to space wells more widely than the economic-optimum.

Permissible Spacing

Jacob (1950, p. 348, eq. 34) derived an equation for computing the permissible distance between production and disposal wells. Assuming that the disposal well is directly down the maximum water-table gradient from the pumped well, the permissible spacing may be computed from the following form of Jacob's equation:

$$A = \frac{2Q}{\pi T I_o} \qquad (2)$$

where A is the permissible distance, in feet, between production and disposal wells, Q is the pumping rate or disposal rate (assumed to be equal), in gpd, T is the coefficient of transmissibility, in gpd/ft, and I_o is the natural hydraulic gradient of the water table, in ft/ft. Like other ground-water equations, the one above necessarily assumes idealized aquifer conditions which differ widely from real conditions.

The distance between the pumped wells and disposal wells in the CPP area is 1,800 feet. The spacing of the wells was decided upon before the wells were drilled, at a time when values of T were not available. The results of aquifer tests, pumping-rate data, and the equation given above were used to determine the adequacy of the spacing.

The hydraulic gradient of the water table in the CPP area before pumping started was about 5 feet per mile or 0.001 ft/ft. The installed capacity of wells CPP-1 and -2 in 1955 was 2,500 gpm (3,600,000 gpd). Ordinarily, only one well is pumped at a time and pumping is not continuous; the probable average total pumping rate of the wells is 3,000,000 gpd. Using that value for Q, a hydraulic gradient of 0.001

well-construction data, and values of specific capacity were substituted in the nonequilibrium formula to obtain rough estimates of the coefficient of transmissibility of the Snake River basalt. Judging from the common range and average of the specific capacities given in table 11, the coefficient of transmissibility of the Snake River basalt ranges between $1x10^5$ and $7x10^6$ and averages about $4x10^6$ gpd per foot for the Snake River Plain as a whole. Based on data given in table 10 the average coefficient of transmissibility of the Snake River basalt for the NRTS area is about $2x10^6$ gpd per foot.

Well Spacing

The desirable distance between wells varies with the physical conditions and the uses of wells and, like well construction, is an individual problem at each site. The problem of determining the desirable spacing of wells is chiefly economic. Widely spaced wells interfere with each other very little, but the cost of interconnecting pipelines and electrical transmission lines is greater than for closely spaced wells. Where one well is pumped and the other is an intake (disposal) well, the desirable spacing (here called "permissible spacing") is that which assures no recirculation of waste liquid through the pumped well. Where all wells are pumped, the desirable spacing (here called "economic-optimum spacing") is that which assures maximum water production at the least cost for construction and for long-term operation and maintenance.

The wells for which aquifer-test data are available penetrate only part of the saturated zone in the Snake River basalt. Layers of relatively impermeable basalt and fine-grained interflow sediments separate permeable zones at many places, and they also impede hydraulic communication between the zones penetrated by wells and deeper parts of the basalt. Therefore, the computed values of T probably apply chiefly to the part of the basalt penetrated by wells or in free communication with them. The coefficient of transmissibility of the entire thickness of the Snake River basalt probably greatly exceeds the values computed from data on partially penetrating wells.

The computed coefficients of storage vary from 0.01 to 0.06 and are in the range characteristic of water-table conditions. These coefficients apply largely to the part of the aquifer through which the water table fluctuated during the tests, and they may not represent the whole basalt aquifer. Furthermore, because water drains from some of the basalt very slowly, longer pumping tests might give considerably larger coefficients of storage. The coefficients of S computed from the data for several aquifer tests in the ANP area and given in table 14 range from 0.01 to 0.02. Considering the complex characteristics of the basalt the agreement between the values of S is as close as can be expected.

Derivation from Specific-Capacity Data

Specific-capacity data can be used to estimate the coefficient of transmissibility (Theis and others, 1954). The average coefficient of storage computed from the results of aquifer tests (see table 14),

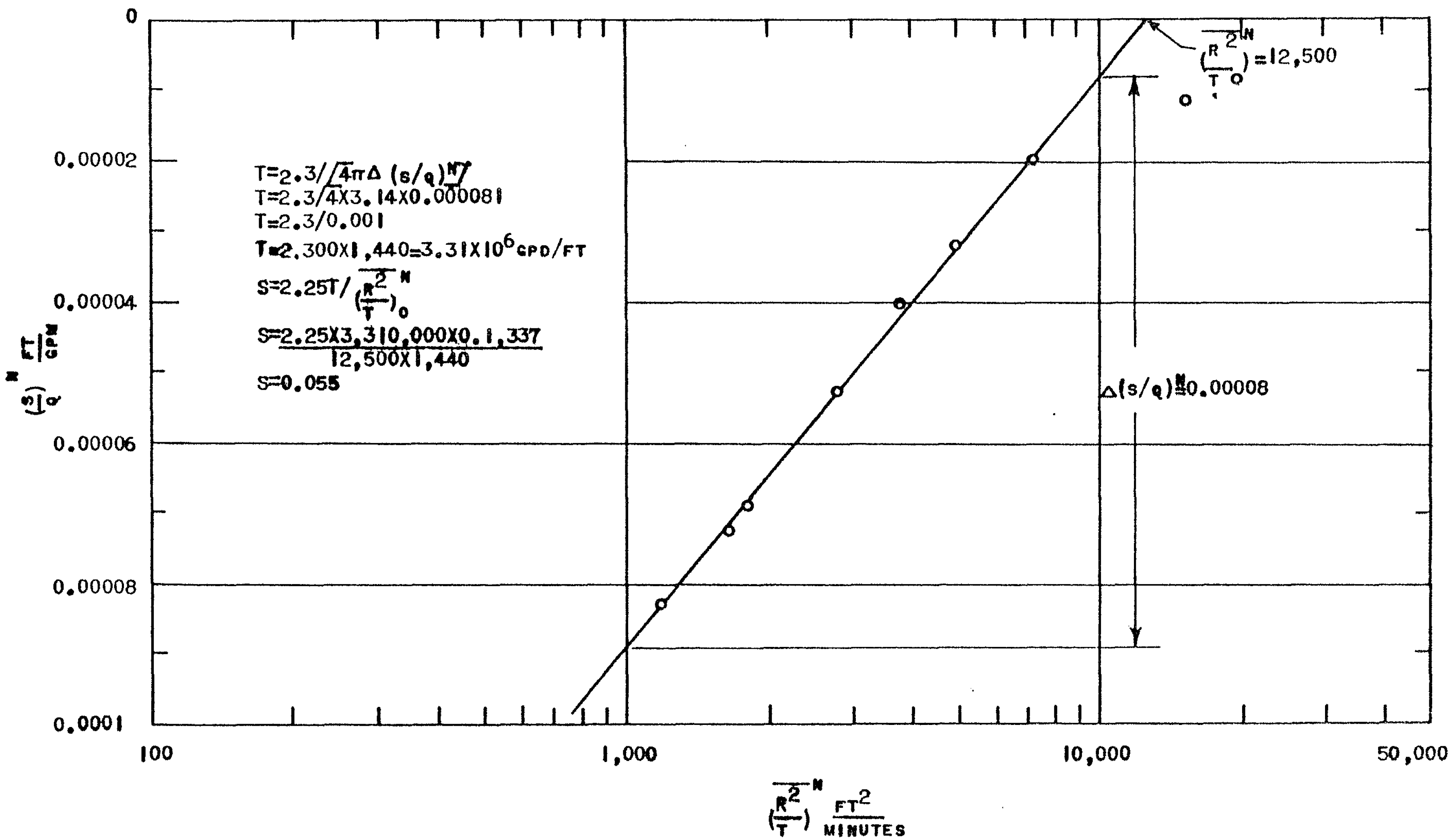


FIGURE 10.--GENERALIZED COMPOSITE DRAWDOWN GRAPH FOR WELL 3N-30E-19CB1 (CPP-3).

Table 14.--Coefficients of transmissibility and storage of the aquifer at the NRTS and vicinity

Pumped well	Date of test	Observation well	Observation of transmissibility (gpd/ft)	Coefficient of storage	Penetration of pumped well below water table (feet)
6N-31E-13ac1 (ANP-1)	4/16-17/53	None	700,000	-	157
do	4/30/53	6N-31E-13db1 (USGS 24)	950,000	0.01	157
do	7/20-23/53	6N-31E-13ca1 (ANP-Disposal)	640,000	0.01	157
6N-31E-13ac2 (ANP-2)	11/11-16/53	6N-31E-13ac1, -13ca1, -12ac1 (ANP-1, ANP-Disposal, IET-Disposal)	800,000	0.02	133
5N-31E-10cd1	7/10-11/53	None	570,000	-	177
4N-30E- 7ad1	8/29-30/50	do	1,700,000	-	367
4N-30E-30ad1 (NRF-2)	8/3-5/51	do	3,700,000	-	163
3N-30E-19bc1, and -24ad1 (CPP 1 and 2)	11/11-13/51	3N-30E-19cb1 (CPP-3)	3,300,000	0.06	153 & 136
2N-29E- 1db1 (CF-2)	2/27/51	None	160,000	-	209

logarithmic mean for well CPP-3—Continued

Log_{10} (r^2_k/t^k) (ft²/min)	ΔQ_k (gpm)	$\Delta Q_k \times \mathrm{Log}_{10}$ (r^2_k/t^k)	Log 10 $(r^2/t)_n$	$\overline{(r^2/t)}_n$ (ft²/min)	s_n (feet)	$(s/Q)_n$ (ft/gpm)
3.223	965	3,110				
3.248	535	1,740				
3.346	1,940	6,500				
3.826	- 400	-1,520				
	3,040	9,830	3.230	1,670	.22	.0000724
3.146	965	3,040				
3.164	535	1,700				
3.243	1,940	6,300				
3.571	- 400	-1,420				
	3,040	9,620	3.160	1,450	.245	.0000806
3.079	965	2,970				
3.097	535	1,640				
3.161	1,940	6,150				
3.412	- 400	-1,360				
	3,040	9,400	3,090	1,230	.25	.0000822
3.076	965	2,960				
3.090	535	1,650				
3.155	1,940	6,130				
3.403	- 400	-1,360				
	3,040	9,380	3.080	1,200	.25	.0000822
3.045	965	2,940				
3.064	535	1,640				
3.124	1,940	6,060				
3.348	- 400	-1,330				
4.281	-1,940	-8,200				
	1,100	1,110	1.010	10	.21	.00019

Table 13.—Specific drawdown and weighted

Date and time	k	n	Discharge well	r_k (feet)	t^k (minutes)	(r^2_k/t^k) (ft^2/min)
November 13						
6:30 A	1		CPP-2	1,830	2,000	1,670
	2		CPP-2	1,830	1,890	1,770
	3		CPP-1	1,830	1,510	2,220
	4		CPP-2	1,830	500	6,700
		4				
1:10 A	1		CPP-2	1,830	2,400	1,400
	2		CPP-2	1,830	2,290	1,460
	3		CPP-1	1,830	1,910	1,750
	4		CPP-2	1,830	900	3,720
		4				
7:50 A	1		CPP-2	1,830	2,800	1,200
	2		CPP-2	1,830	2,690	1,250
	3		CPP-1	1,830	2,310	1,450
	4		CPP-2	1,830	1,300	2,580
		4				
8:15 A	1		CPP-2	1,830	2,825	1,190
	2		CPP-2	1,830	2,715	1,230
	3		CPP-1	1,830	2,335	1,430
	4		CPP-2	1,830	1,325	2,530
		4				
11:10 A	1		CPP-2	1,830	3,000	1,110
	2		CPP-2	1,830	2,890	1,160
	3		CPP-1	1,830	2,510	1,330
	4		CPP-2	1,830	1,500	2,230
	5		CPP-1	1,830	175	19,110
		5				

logarithmic mean for well CPP-3

$\log_{10}$ (r^2_k/t^k) (ft²/min)	ΔQ_k (gpm)	$\Delta Q_k \times \log_{10}$ (r^2_k/t^k)	Log 10 $(r^2/t)^n$	$\overline{(r^2/t)}^n$ (ft²/min)	s^n (feet)	$(s/Q)^n$ (ft/gpm)
4.483	965		4.483		.025	
3.835	965	3,700				
3.945	535	2,110				
	1,500	5,810	3.870		.05	
3.747	965	3,620				
3.835	535	2,050				
4.483	1,940	8,700				
	3,440	14,370	4.180	15,140	.04	.0000116
3.621	965	3,490				
3.687	535	1,970				
4.033	1,940	7,830				
	3,440	13,290	3.860	7,240	.07	.0000203
3.525	965	3,400				
3.575	535	1,910				
3.818	1,940	7,400				
	3,440	12,710	3.700	5,000	.11	.0000320
3.446	965	3,320				
3.487	535	1,860				
3.674	1,940	7,120				
	3,440	12,300	3.580	3,800	.14	.0000406
3.348	965	3,230				
3.382	535	1,810				
3.521	1,940	6,830				
	3,440	11,870	3.450	2,820	.18	.0000523
3.270	965	3,160				
3.297	535	1,760				
3.408	1,940	6,620				
4.047	- 400	-1,600				
	3,040	9,940	3.270	1.860	.21	.0000691

Table 13.—Specific drawdown and weighted

Date and time (1951)	k	n	Discharge well	r_k (feet)	t^k (minutes)	(r^2_k/t^k) (ft^2/min)
November 11						
11:00 A	1	1	CPP-2	1,830	110	30,420
5:20 P	1		CPP-2	1,830	490	6,840
	2		CPP-2	1,830	380	8,820
		2				
7:10 P	1		CPP-2	1,830	600	5,580
	2		CPP-2	1,830	490	6,840
	3		CPP-1	1,830	110	30,420
		3				
10:30 P	1		CPP-2	1,830	800	4,180
	2		CPP-2	1,830	690	4,860
	3		CPP-1	1,830	310	10,800
		3				
November 12						
1:50 A	1		CPP-2	1,830	1,000	3,350
	2		CPP-2	1,830	890	3,760
	3		CPP-1	1,830	510	6,570
		3				
5:10 A	1		CPP-2	1,830	1,200	2,790
	2		CPP-2	1,830	1,090	3,070
	3		CPP-1	1,830	710	4,720
		3				
10:10 A	1		CPP-2	1,830	1,500	2,230
	2		CPP-2	1,830	1,390	2,410
	3		CPP-1	1,830	1,010	3,329
		3				
11:50 A	1		CPP-2	1,830	1,800	1,860
	2		CPP-2	1,830	1,690	1,980
	3		CPP-1	1,830	1,310	2,560
	4		CPP-2	1,830	300	11,150
		4				

Summary of Results of Tests, 1949-55

The results of aquifer tests during 1949-1955 are summarized in table 14. Values of the coefficient of transmissibility of the Snake River basalt range between $1.6x10^5$ gpd/ft and $3.7x10^6$ gpd/ft. The higher values of T (more than $3x10^6$ gpd/ft) are those computed from data for the CPP and NRF areas. The specific capacity of well MTR-1 indicates that the coefficient of transmissibility exceeds $3x10^6$ gpd/ft in the MTR area also. The data in table 14 indicate that the average coefficient of transmissibility of the basalt in the ANP area is about $8x10^5$ gpd/ft. The lowest value of T, $1.6x10^5$ gpd/ft, was computed from test data for well CF-2, in the southern part of the NRTS.

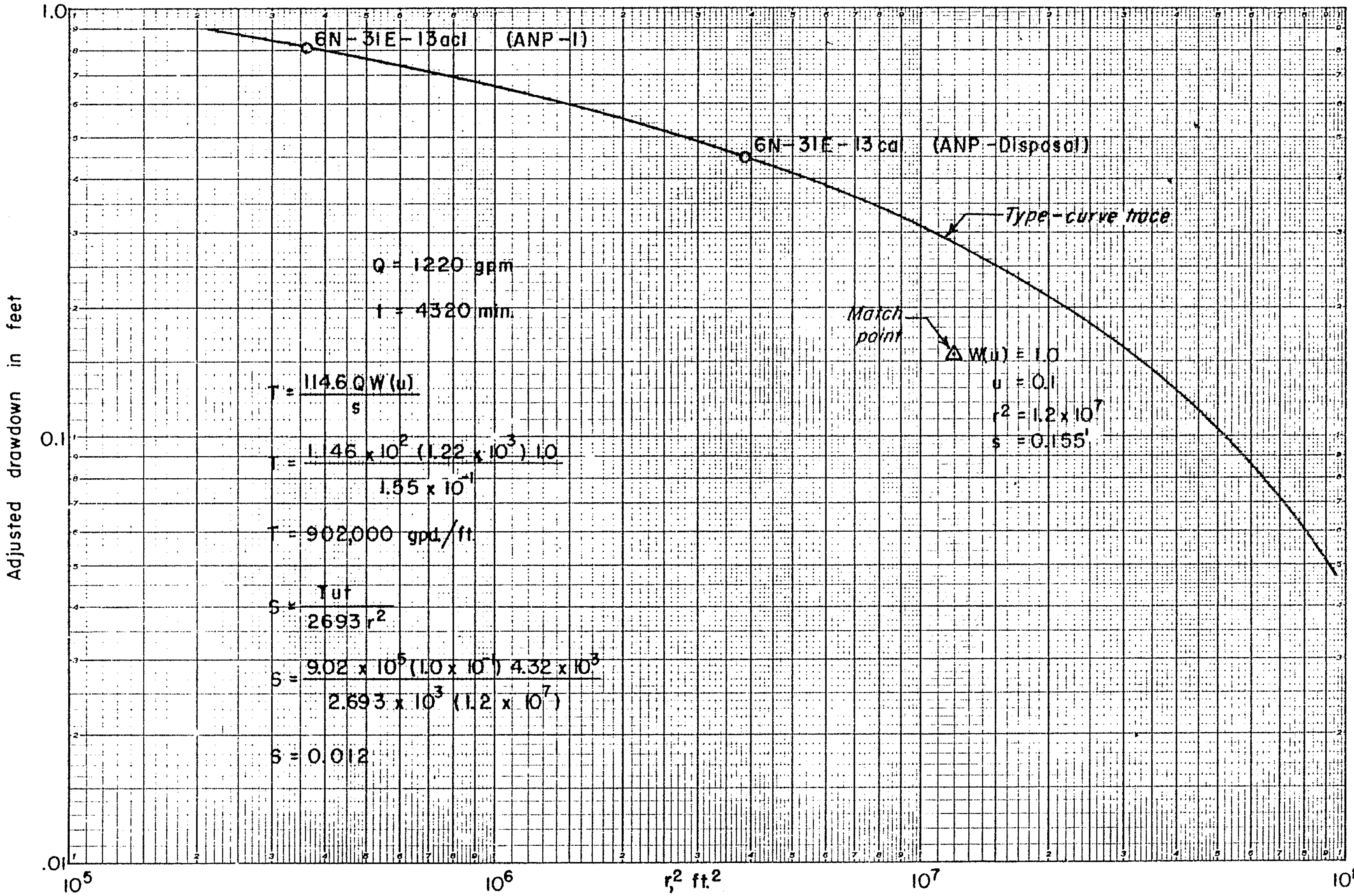


Figure 9.—Distance drawdown curve for well 6N-31E-13ac2 (ANP-2)

Interpretive studies of the time-drawdown data for well ANP-1 show that the hydraulic properties of the basalt vary within the area of influence around well ANP-2. None of the changes, however, is sufficiently great even to approximate the effect of an impermeable boundary. Such a boundary would distort the cone of influence, and the values of T and S computed from distance-drawdown data would disagree with those computed from unaffected time-drawdown data. The agreement between the two sets of T values supports belief that no important impermeable boundaries exist within the area of influence.

CPP Area

Wells CPP-1 and -2 were pumped simultaneously for 39 hours on November 11-13, 1951. The pumping rates for wells CPP-1 and -2 were 1,500 and 1,940 gpm, respectively. The drawdown from concurrent pumping of the two wells was measured in well CPP-3. The conditions of the test are described and water-level data are tabulated in appendix 2.

The test data were analyzed by the generalized composite-drawdown graph method (Cooper and Jacob 1946). Computations to determine values of the weighted logarithmic mean $\overline{(r^2/t)}^n$ and the corresponding values of the specific drawdown $(s/Q)^n$ are shown in table 13.

The calculated values in table 13 were plotted on semilogarithmic paper to obtain the generalized composite-drawdown graph in figure 10. A straight line was drawn through the plotted points. The slope of the line and the intercept of the line with the line of zero specific drawdown were substituted in the equations shown in figure 10 to compute the hydraulic coefficients of the aquifer. The computed values of T and S are 3.3×10^6 gpd/ft and 0.06, respectively.

Table 12.—Hydraulic coefficients of the aquifer in the ANP area

Well number	Type of data	Coefficient of transmissibility (gpd/ft)	Coefficient of storage
6N-31E-13ac2 (ANP-2)	Time-drawdown	800,000	-
6N-31E-13ac1 (ANP-1)	do	820,000	0.02
6N-31E-13ca1 (ANP-Disposal)	do	700,000	.02
6N-31E-13ac1, -13ca1, (ANP-1 and ANP-Disposal)	Distance-drawdown	900,000	.01
	Average (rounded)	800,000	0.02

Extension of the distance-drawdown curve in figure 9 indicates that the 72-hour test sampled an area of basalt having a diameter of about 8 miles. The computed coefficients represent the average hydraulic properties of the basalt within that large area of influence. Considering the complex characteristics of the basalt, the agreement between the several values of T and S that were computed from data obtained from the three wells is as close as can be expected. However, as is obvious from the preceding section of this report, the performance of a new well at a specific location in this area cannot be predicted accurately by using the computed values of T and S. Because of local irregularities in the hydraulic properties of the basalt, the performance of a new well might differ substantially from that of the existing ones.

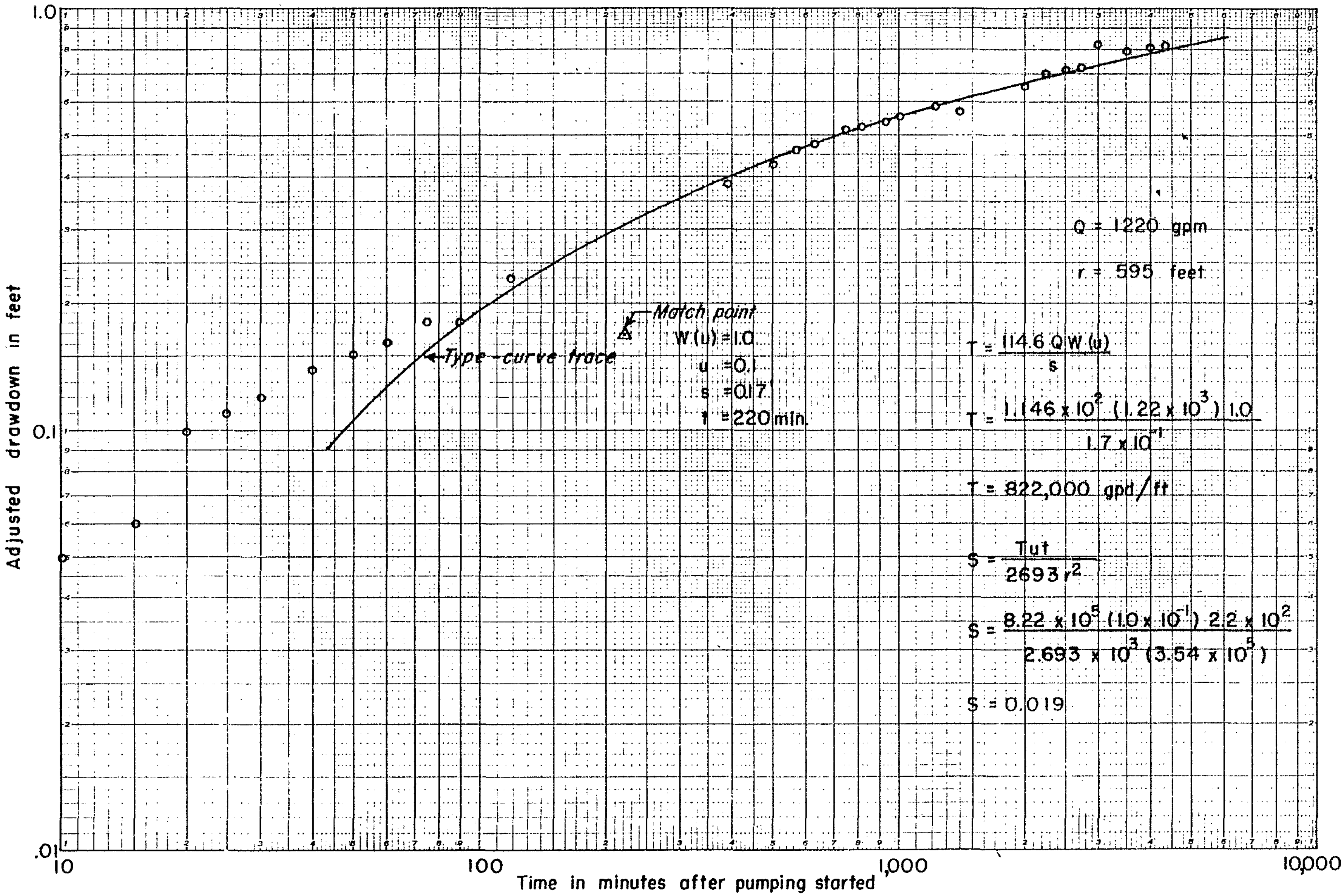


Figure 8.—Time-drawdown curve for well 6N-31E-13ac1 (ANP-1).

Adjusted drawdowns in observation wells ANP-1 and ANP-Disposal at the end of the test were plotted on logarithmic paper against the squares of the distances from the respective observation wells to the pumped well, to obtain a distance-drawdown curve. This curve is part of the profile of the cone of influence around the pumped well. The type curve was matched to the distance-drawdown curve, match-point coordinates were substituted in the formula, and the coefficients of transmissibility and storage were computed as shown in figure 9.

The drawdown in well IET-Disposal, which is 5,450 ft from the pumped well, cannot be determined accurately because barometric water-level fluctuations almost completely masked the fluctuations caused by pumping. However, water-level data show that the drawdown in this well was in the magnitude of 0.1 ft at the end of the test.

The values of T and S, computed from time-drawdown and distance-drawdown data, are listed in table 12.

Drawdown in the pumped and observation wells was determined by comparing extrapolated graphs of water levels measured before pumping started with the graphs of water levels measured during pumping. The drawdown data were then adjusted for atmospheric-pressure changes during the test. Adjusted drawdowns were plotted against time on logarithmic or semilogarithmic paper.

As an example, figure 8 shows the time-drawdown data for well ANP-1, adjusted for a barometric efficiency of 88 pct. The test data were analyzed by the graphical method of superposition (Wenzel, 1942, p. 88-89) to solve the nonequilibrium formula. The type curve was matched to the latter part of the time-drawdown data, and match-point coordinates were substituted in the formula to compute the coefficients of transmissibility and storage. Computations for well ANP-1 are shown in the figure.

The drawdown during the first 90 minutes of pumping departed from the type curve because the basalt aquifer does not respond instantaneously to pumping, as would a homogeneous, isotropic, and perfectly elastic artesian aquifer. There is a time-lag in the discharge of water released from storage in the aquifer and the different hydraulic properties of two successive permeable zones can not be integrated instantaneously by the cone of influence. As pumping continued, however, the flow between zones having differing permeability adjusted itself, the effects of gravity drainage became small, and the effects of differences between the water-bearing characteristics of the basalt and assumed ideal characteristics became negligible.

outlined by Ferris (1949) and Cooper and Jacob (1946) were used to calculate coefficients of transmissibility and storage.

Water levels in wells at the NRTS fluctuate in response to fluctuations in atmospheric pressure. These effects are discussed fully in the section of this report entitled "barometric fluctuations". The fluctuations are sufficiently large to mark the influence of pumping, so the drawdown data collected during the aquifer tests were adjusted for changes in atmospheric pressure before they were used to compute the hydraulic coefficients. Adjustments for each well were based on its barometric efficiency, which is the ratio of the change in the water level in the well to the associated change in atmospheric pressure. The barometric efficiency of a well is determined by comparing prominent changes in atmospheric pressure with corresponding fluctuations in water level during a time interval when water levels were not affected by pumping.

The data for two aquifer tests will now be presented in detail to illustrate methods of analysis. The results of these and 8 other tests are summarized on later pages.

ANP Area

During pumping of well ANP-2 on November 12-15, 1953, the drawdown was measured in wells ANP-1, ANP-Disposal, and IET-Disposal. Pumping was started at 12:30 pm on November 12 and continued for 72 hours at a constant rate of 1,220 gpm. The water-level data are tabulated and the conditions of the test are described in Appendix 2.

Much of the water pumped during an aquifer test is withdrawn from storage by gravity drainage from part of a permeable zone. That is, some of the aquifer material is dewatered. The flow of water toward a well is three-dimensional because gravity drainage of stored water involves vertical flow components during slow dewatering of the aquifer. However, the nonequilibrium formula describes only two-dimensional radial flow with instantaneous release from storage. Therefore, the formula accurately defines drawdown in the Snake River basalt only after the flow of water between permeable interflow zones having different hydraulic properties reaches approximate stability. At that stage, the vertical-flow components of gravity drainage are small, dewatering is essentially complete, and the flow is largely radial and essentially two-dimensional. Before that stage the actual drawdown deviates from the theoretical. The initial transitional period may be several minutes or hours in duration, depending on aquifer conditions. During prolonged pumping the cone of influence adjusts to the average and overall hydraulic properties of the aquifer, and the formula then closely describes the drawdown in wells. Thus, meaningful values for hydraulic properties of the Snake River basalt can be derived if aquifer tests are made properly and if judgment, based on knowledge of geologic conditions, is used in interpreting data.

Aquifer Tests

The data for 10 aquifer tests at the NRTS during 1949-1954 were analyzed to determine the hydraulic properties of the Snake River basalt. No tests were made in 1955. The nonequilibrium formula and methods aquifer is infinite in areal extent and uniform in thickness; that it is homogeneous and isotropic; that it is confined between impermeable beds; that the coefficient of storage is the same for all parts of the aquifer; that water is released from storage instantaneously with a decline in head; and that the well has an infinitesimal diameter and penetrates the entire thickness of the formation. The nonequilibrium formula is a particular solution of a partial differential equation describing two-dimensional radial flow. Theoretically the formula is not valid for water-table conditions, because the flow towards a well in water-table aquifer is three-dimensional and release of water from storage is not instantaneous with the decline in head.

The ideal conditions assumed in the derivation of the formula do not, of course, prevail in any aquifer. In pahoehoe flows of the Snake River basalt, for example, permeable zones occur largely along the contacts between separate flows. Massive layers of relatively impermeable basalt and fine-grained interbeds separate many of the permeable zones, and the hydraulic connection between these zones is poor. The hydraulic permeability of the basalt, like the effective porosity, is a characteristic of the formation and not of the rock itself, because the rock is essentially impermeable. The vertical formational permeability of the basalt is much smaller than the horizontal.

Most wells in the Snake River basalt extend through two or more permeable interflow zones. The permeability is not likely to be the same in any two successive zones. Water in upper zones commonly occurs under water-table conditions, whereas, water in lower zones ordinarily is confined to some extent and occurs under quasi-artesian or artesian conditions.

HYDRAULIC PROPERTIES

The coefficients of transmissibility, T, and storage, S, are significant hydraulic properties of an aquifer. The coefficient of transmissibility is the rate of flow of water, in gallons per day, through a vertical strip of the aquifer 1 foot wide and extending the full saturated thickness under a hydraulic gradient of 100 percent (1 foot per foot) at the prevailing temperature of the water. The coefficient of storage is the volume of water released from or taken into storage per unit surface area of the aquifer per unit change in the component of head normal to that surface. Values for the coefficients are derived mathematically from field data.

Derivation and Significance of Coefficients

The hydraulic properties of the Snake River basalt were determined by means of aquifer tests in which the effects of pumping were observed. The drawdown caused by pumping a well at a constant known rate is measured in the pumped well and at observation wells which tap the same aquifer. Graphs of drawdown versus time after pumping started, and of drawdown versus distance from the pumped well, are used to solve equations which express the relation between the coefficients of transmissibility and storage of an aquifer and the lowering of water levels in the vicinity of a pumped well.

The formula most widely used to analyze aquifer-test data is the nonequilibrium formula (Theis, 1935). The formula, like other similar ones, was developed on the basis of the following assumptions: that the

Table 11.—Specific capacities of wells in the Snake River Plain

Location	Number of wells	Average depth (feet)	Average penetration below water table (feet)	Average diameter of casing (inches)	Average length of test (hours)	Range of specific capacity (gpm/ft)	Average specific capacity (gpm/ft)	Average pumping rate (gpm)
Jefferson County	5	200	72	16	16	33–14,000	3,200	2,400
Minidoka, Jerome and Lincoln Counties 1/ Group 1	15	266	83	21	–	110– 9,000	2,300	2,900
2	21	264	94	24	2½	110–20,000	4,500	2,500
3	29	282	91	21	3	200– 6,000	1,700	2,200
4	31	276	88	21	3	59–14,000	1,800	1,800
5	22	308	83	21	4½	23–17,000	1,800	1,700
6	18	402	112	21	2½	63– 3,000	800	1,500
7	30	271	71	21	3½	150–22,000	3,900	1,800
Bingham County	15	252	71	18	8	16– 7,700	1,800	1,900
NRTS, Butte County	14	590	200	16	24	12– 2,900	900	900
Gooding County	9	120	60	12	2	300– 2,700	900	1,000
Jerome County	6	362	77	16	3	6– 1,600	900	1,700
Bonneville County	6	207	88	18	4	33– 1,000	500	1,800
Blaine County	2	226	179	16	2	37– 110	70	1,400
Camas County	3	210	77	16	6	30– 110	60	1,200

1/ U. S. Bureau of Reclamation Minidoka Project, North Side Pumping Division.

small, well losses at pumping rates of several cubic feet per second are appreciable parts of the total drawdowns in the wells. For example, the well loss of 3.1 ft in well CPP-1 at a pumping rate of 2,475 gpm is nearly 53 pct of the total drawdown, 5.9 ft, at that rate. Large head losses occur as water enters the wells at high velocities through the gravel envelope and perforated casing. Seventy-five feet of the 16-inch casing in well CPP-1 are perforated with 3/8 by 3-in. longitudinal slots, providing a total of 11 sq ft of entrance area. Gravel is packed around in the slotted pipe in the well. Assuming that the effective open area is 5.5 sq ft, the entrance velocity would be 60 ft per min at a pumping rate of 2,475 gpm.

Table 11 summarizes the data now available on the specific capacities of production wells in the Snake River basalt throughout the Snake River Plain. The average yield of a 16-inch well during an 8-hr test was about 2,000 gpm per foot of drawdown. The average depth of the wells is 290 ft, and the average penetration below the regional water table is 100 ft. Although the specific capacities range from 6 to 22,000 gpm per foot, the common range is from 60 to 3,200 gpm per foot.

Specific-capacity data can be used to estimate the coefficient of transmissibility of aquifers, and this will be done later in this report.

The specific capacity of well NRF-2 during the test on June 11-12, 1951 was unusually low, either because the well was not properly developed before the test, or because entrained air blocked part of the aquifer, or both. The test on Aug. 3-5, 1951 showed that subsequent development by pumping greatly increased the specific capacity.

The specific capacity of well CPP-3 is abnormally low because at the time of the test only the lower part of the casing was perforated. The casing was not perforated opposite permeable zones in the upper part of the basalt.

The yields of some wells in the basalt have been increased greatly by relatively small increase in their depth. For example, when well NRF-1 was 485 feet deep, its specific capacity was 16 gpm/ft. After the well was deepened to 535 feet, its specific capacity was 2,860 gpm/ft. This and many similar cases illustrate that in many wells the yield is obtained principally from a few extremely permeable zones. Water is not contributed uniformly by the entire penetrated thickness of the zone of saturation.

During the period 1949-1955, two or more tests at different pumping rates were made on several wells. The specific capacities of wells MTR-1, CPP-1 and -2, and ANP-1 decrease as the pumping rate increases (table 10), which indicates that well loss is appreciable.

Accurate well-loss constants can not be computed from the available data. Their order of magnitude, however, was determined by a method outlined by Jacob (1947). The value of C is less than 0.05 for well MTR-1; it seems to be about 0.1 for wells CPP-1 and -2, but the data for these wells is not consistent. Although the well-loss constants are relatively

Table 10.—Specific capacities of wells in the NRTS—Continued

Well Number: Field	Well Number: Location	Date of test	Duration of test (hours)	Pumping rate (gpm)	Drawdown or buildup (+) (feet) 1/	Specific capacity (gpm/ft)	Diameter of casing (inches)	Depth of well (feet)	Depth to static water level (feet)	Well construction
USGS 30	5N-33E-13bd1	7/27-28/53	24	250	0.1	2,500	8	405	265	Casing perforated, 276-290 ft, 300-317 ft, 8-in. open hole 360-405 ft.
USGS 31	5N-33E-10cd1	7/10-11/53	24	280	1.6	175	8	428	251	Casing perforated, 285-305 ft; open hole 306-428 ft.
ANP-1	6N-31E-13ac1	4/16-19/53	24	1,070	7.7	139	16	365	208	Casing perforated, 200-335 ft.
		8/31-9/2/53	do	1,130	10	113	do	do	do	
		7/20-23/53	do	1,240	11.5	108	do	do	do	
ANP-2	6N-31E-13ac2	11/12-16/53	24	1,220	21.3	57	16	345	212	Casing perforated 235-335 ft.
		8/31-9/2/53	do	1,130	+28.7	39	do	do	211	

1/ Plus values represent buildup of water level during recharge test.

2/ Well CPP-2, 500 ft distant, was pumped intermittently during the test at 2,300 gpm, which may account for the anomalous apparent specific yield.

Table 10.--Specific capacities of wells in the NRTS--Continued

Well Number		Date of test	Duration of test (hours)	Pumping rate (gpm)	Drawdown or buildup (+) (feet)1/	Specific capacity (gpm/ft)	Diameter of casing (inches)	Depth of well (feet)	Depth to static water level (feet)	Well construction
Field	Location									
CPP-2	3N-29E-24ad1	4/21-22/51	24	1,030	+1	1,030	16	605	452	Casing perforated, 458-483 and 551-600 ft.
		11/5-7/54	3	1,455	1.3	1,120	16	605	451	
		11/11-13/51	24	1,500	2.4	625	16	605	452	
		6/29-30/51	24	1,850	2.7	685	16	605	451	
		11/22-24/54	24	2,500	3.2	781	16	605	451	
CPP-3 Disposal	3N-30E-19cb1	9/11-12/51	24	800	+38.1	21	16	598	451	Casing perforated, 490-593 ft. Casing perforated, 412-452 ft 10/51.
NRF-1 (STR-1)		8/3-4/50	24	1,010	62	16	18	485	363	Casing perforated, 394-478 ft.
		11/17-18/50	12	1,400	0.5	2,800	18-16	535	do	18-in. casing perforated, 394-478 ft, 16-in. casing perforated 485-530 ft.
NRF-2 (STR-2)	4N-30E-30ad1	8/3-5/51	48	1,430	0.5	2,860	16	528	365	Casing perforated, 373-397 ft, 422-448 ft, 497-523 ft.
		6/11-12/51	24	1,245	+4.8	259	do	528	364	
USGS 12	4N-30E- 7ad1	8/29-30/50	24	420	5	84	12-10	692	325	12-in. open hole 587-692 ft.

Table 10.--Specific capacities of wells in the NRTS

Well Number: Field	Well Number: Location	Date of test	Duration of test (hours)	Pumping rate (gpm)	Drawdown or buildup (+) (feet)1/	Specific capacity (gpm/ft)	Diameter of casing (inches)	Depth of well (feet)	Depth to static water level (feet)	Well construction
CF-2	2N-29E- 1db1	2/27/51	3	235	15.6	15	20-16	681	472	16-in. casing perforated, 521-651 and 661-681 ft.
EBR-1	2N-29E- 9ca1	8/12-14/49	24	800	17	47	18	1,075	596	Casing perforated, 600-750 ft, open 10½-in. hole 750-1,075 ft.
MTR-1	3N-29E-14aa1	5/26-29/50	24	810	0.3	2,700	18	600	456	Casing perforated, 481-581 ft.
		3/27/50	19	1,125	0.5	2,250	18	600	456	
MTR-2	3N-29E-14aa2	1/22-24/51	24	750	61	12	18-12	772	457	18-in. casing perforated, 496-571 ft, 12-in. casing perforated, 558-567 ft, open 10-in. hole 572-772 ft.
		10/28-11/ 2/52	24	590	47	13	18-8	772	455	8-in. casing perforated, 645-744 ft.
CPP-1	3N-30E-19bc1	12/15-17/50	24	1,130	1.9	594	16	585	449	Casing perforated, 460-485 ft, 527-577 ft.
		3/31-4/1/51	24	1,140	1.9	600	16	585	449	
		11/11-13/51	24	1,940	4.8	404	16	585	450	
		11/5 - 7/54	45	2,475	5.9	2/419	16	585	448	

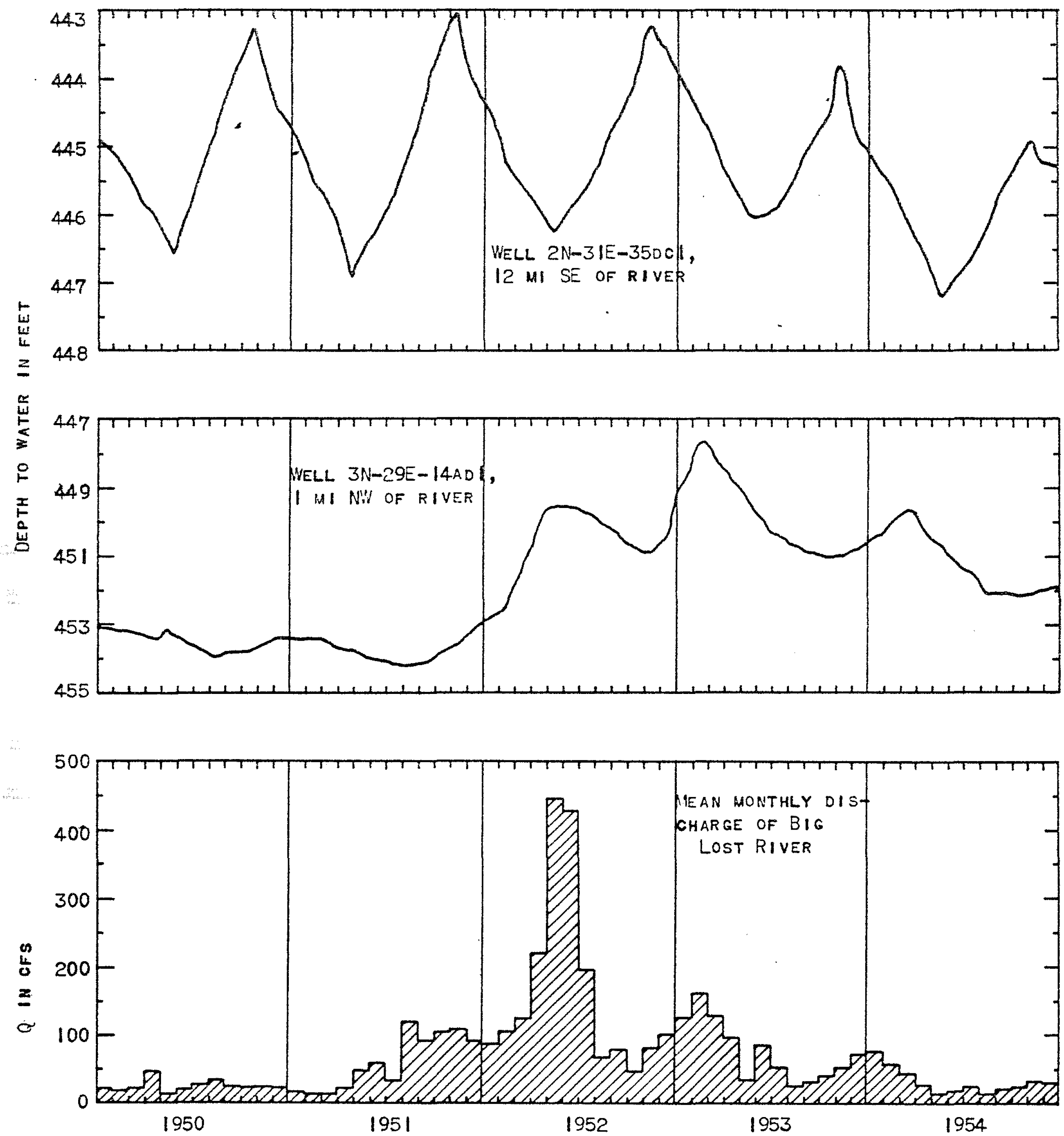


FIGURE 7.—COMPARISON OF GENERALIZED HYDROGRAPHS OF BIG LOST RIVER AND WELLS 2N-31E-35DC1 AND 3N-29E-14AD1, 1950-54.

friction of turbulent flow in the aquifer in the immediate vicinity of the well, through the well screen and in the well casing. Well loss, in feet, may be represented approximately by the following relationship (Jacob, C. E., 1947):

$$\text{well loss} = CQ^2 \qquad (1)$$

where C is the "well-loss" constant, its dimension being in sec^2/ft^5, and Q is the rate of pumping in cubic feet per second. A decrease in the specific capacity of a well with an increase in the pumping rate may indicate that well loss is appreciable, but other factors may cause the same result. Examples are dewatering of some of the saturated zone, pump intake pipe breaking suction, and perhaps others.

Specific-capacity data for wells in the NRTS and vicinity are summarized in table 10. The specific capacities range from 12 to 2,860 gpm/ft. The tabulation indicates that, in the NRTS area, the average specific capacity of a well 16 inches in diameter during a 24-hour test is about 900 gpm/ft. The average depth of the wells is 586 ft, and average depth of penetration below the regional water table is 206 ft.

Wells NRF-1 and -2 and MTR-1 have the largest specific capacities on the NRTS, averaging about 2,800 gpm/ft. Wells ANP-1 and -2 in the northern part of the NRTS, and wells CF-2 and EBR-1 in the southern part, have moderately low specific capacities. The specific capacity of well CF-2 might be about of the same magnitude as that for well EBR-1 (47 gpm/ft) if the depth of penetration were as great (480 feet).

On the other hand, the record for 1950-54 shows that the water level in well -14adl normally rises shortly after August in some years (fig. 7). The rise was cumulative during the three high-runoff years of 1951-53, culminating in the record-high water level in 1953. If the correlations shown in figure 7 are valid, the lag between infiltration and arrival of the water at the water table may be on the order of 6 to 8 months.

Suppose that percolating water from the river spread to a belt 500 to 1,000 feet wide beneath a 10 to 20 mile reach of the river. The area in which percolation recharge occurred would be about 600 to 2,400 acres. At 3.5 pct average formational porosity, it would require 21 to 84 ac-ft of water to raise the water table 1 ft. A rise of about 4 ft occurred in 230 days from September 10 to the end of April, representing about 84 to 330 acre-feet. This increment would be equivalent to a continuous recharge flow of about 0.18 to 0.72 cfs. These figures have no standing whatever as fact. They merely show that it is reasonable to expect that recharge from the river would cause a detectable rise in the water table. The matter requires further study and no conclusions are drawn now.

SPECIFIC CAPACITIES OF WELLS

The performance of a well may be expressed in terms of its specific capacity, which is defined as the yield in gallons per minute per foot of drawdown. The specific capacity of a well varies both with the duration of pumping and with the pumping rate.

The drawdown in a pumped well has two components. One is the head loss from the friction of laminar flow of water in the aquifer towards the well. The other is the head loss (well loss) resulting from the was unusually great (see p. 17-21), and much water entered the ground by infiltration along the river channel on the NRTS. If a rise of water levels occurred, it would appear first in wells near the river and later or not at all in distant wells. Therefore, we studied well hydrographs and compared them with the hydrograph of the Big Lost River in 1951-52 (fig. 6).

On September 10, 1951 the water level in several wells near the river on the NRTS rose sharply about 0.35 ft — a large rise in this aquifer and locality. The residual rise after adjustment for a correlative drop in barometric pressure was 0.15 ft. Perhaps this rise was not related to infiltration from the Big Lost River, because a similar rise was recorded on the same day in other wells many miles from the river. Recharge from the river may have reached the water table about concurrently with a temporary rise caused by some other factor. Nevertheless, the water levels in some wells near the river continued to rise after September 10.

Infiltration from the river was along a narrow belt, the wetted channel, but the percolating water spread to a wider belt before reaching the water table. Owing to that spread, the rise of the water table caused by increased infiltration was too small to be detected at its beginning or in daily increments, but infiltration may have caused part of the cumulative upward trend of the water-table in well -14ad1 (see fig. 6). If so, the percolation time through nearly 500 ft of basalt and sediments to the water table was on the order of four to seven weeks.

distinguish "rock porosity" from "formational porosity." We have no practical method for measuring or estimating accurately the formational porosity. It seems certain, from study of outcrops and from drilling records, that the average effective porosity is at least 2 pct and probably is about 5 pct. For the necessary purposes of this report, we assume conservatively that the average effective formational porosity of the Snake River basalt is 3.5 pct.

INFILTRATION RATE

No means have been devised for direct measurements of infiltration rates in the basalt. The rate ranges from high to nil, depending on physical variations in the rock. Wherever there are impermeable layers in the basalt, the percolating water tends to form a perched saturated zone on each impermeable layer. Water in these perched zones moves laterally till it finds openings through which it can move downward. Water disposed at a point or small area on the surface, therefore, tends to spread through a wider area by the time it reaches the water table.

Perching of percolating water is well illustrated by the artificial zone of saturation at the MTR site, which was built up with water released to the environment (see p. 47).

PERCOLATION RATE

In connection with waste disposal and numerous other problems, it would be helpful to know how long water takes to percolate from the land surface to the water table. During 1951-53 runoff in the Big Lost River

PORE WATER AND POROSITY

The presence of pore water in the basalt is easily verified by direct inspection and by simple laboratory tests. Pieces of basalt were collected from under water and from the zone of saturation at springs in the valley of the Snake River. Freshly broken pieces were damp along the fresh breaks. When the specimens were collected, superficial water was dried off and they were weighed immediately on a chemical balance. In the laboratory they were heated in an oven at 105°C and were weighed periodically until they ceased to lose weight. The larger samples required about 100 hours for drying. The twenty-five samples of Snake River basalt listed in table 9 ranged in porosity from about 4 to 25 pct and averaged 10 pct. The water content ranged from 1.4 to 11.6 pct by weight and averaged 4.2 pct. The porosity of two samples of Banbury basalt averaged 18.6 pct and the water content was 8 pct.

The basalt is permeable to air and "breathes" freely. Dried samples exposed to the atmosphere change weight in response to changes in relative humidity.

Effective Porosity

Though the total porosity from pores and vesicles is relatively high, most of the pores are too small to transmit water under the pressures that normally prevail in the aquifer, and most of the vesicles are isolated. The effective porosity of the rock ranges from about nil to a small percentage. Nevertheless, the considerable void space of other kinds in a mass of lava is hydraulically effective. Therefore, we

Table 9.—Loss of pore water from samples of basalt at 105°C

Location of source of sample	Weight of wet sample (gr)	Weight of dry sample (gr)	Volume of material (cm^3)	Volume of water (cm^3)	Porosity (pct)	Water content (pct)	Remarks
Blue Lake Spring	99.851	97.144	34.206	2.707	7.3	2.8	Vesicular
Sec. 28, T. 9 S., R. 17 E.	108.889	103.503	36.445	5.386	12.9	5.2	Sparsely vesicular.
Jerome County	50.508	49.807	17.538	.701	3.8	1.4	Very vesicular
Devil's Washbowl Spring	49.787	45.202	15.916	3.585	18.4	7.9	Scoriaceous
Sec. 4, T. 10 S., R. 18 E.	64.008	60.617	21.344	3.391	13.7	5.6	Vesicular
Jerome County	200.173	188.391	66.335	11.782	15.1	6.3	Scoriaceous
Malad Springs	42.243	40.987	14.432	1.256	8.0	3.1	Slightly vesicular
Sec. 36, T. 6 S., R. 13 E.	108.350	106.576	37.527	1.774	4.5	1.7	Dense
Gooding County	69.379	68.038	23.957	1.341	5.3	2.0	Vesicular
	151.329	145.421	51.205	5.908	10.3	4.1	Vesicular
	162.851	145.878	51.365	16.973	24.8	11.6	Vesicular
	82.578	80.775	28.442	2.800	9.0	3.5	Vesicular
	108.957	102.745	36.178	6.212	14.7	6.0	Very vesicular to scoriaceous
	122.121	116.224	40.924	5.897	12.6	5.1	Vesicular
	106.053	98.336	34.625	7.717	18.2	7.8	Dense
	89.260	84.523	29.762	4.737	13.7	5.6	Dense; fractured
Thousand Springs	177.168	174.518	61.472	2.650	4.1	1.5	Dense
Sec. 8, T. 8 S., R. 14 E.	80.779	78.328	27.580	2.451	8.2	3.1	Vesicular
Gooding County	121.502	119.899	42.218	1.703	3.9	0.8	Dense
	107.733	105.302	37.078	2.431	6.2	2.3	Slightly vesicular
	64.877	63.618	22.401	1.259	5.3	2.0	Very vesicular
	101.947	100.526	35.396	1.421	3.9	1.4	Dense
	174.081	169.941	59.838	4.140	6.5	2.4	Very vesicular
	123.951	114.852	40.441	9.099	18.4	7.9	Banbury basalt
	93.754	86.687	30.524	7.067	18.8	8.2	Banbury basalt

Pores

Much of the Snake River basalt -- even that which seems dense and tight -- is permeated by small pores between its mineral grains. Pores of that type occupy as much as 25 pct of some rock. Laboratory measurements of the porosity of 8 random samples of drill cores from foundation borings on the NRTS were as follows:

Sample	Porosity (pct)	Sample	Porosity (pct)
1	13.4	5	16.0
2	14.3	6	10.1
3	22.3	7	13.7
4	16.7	8	25.2

Tests were made of 25 additional samples in a study of pore water (table 9). The median value for the 33 samples is about 13 pct and the average is 10 pct.

Most of the pores are capillary or subcapillary in size and store water but do not transmit it. In the zone of saturation they are filled with water, but the hydraulic head on the water is not sufficient to overcome the capillary force that holds it in the pores. The water stored in capillary pores cannot be withdrawn through wells because it will not drain from the rock by gravity.

Tension joints that were produced by differential movement of the hardened crust of a lava flow are more widely open than shrinkage joints, and the voids observed range up to 10 or 12 ft wide, a few feet to hundreds of feet long, and at least 20 ft deep. These openings, commonly reported in well logs as crevices, are copious water bearers.

Lava Tubes

Breached lava tubes were found at three places on the Station. Other tubes, breached and unbreached, undoubtedly occur because they are not rare on the plain. No subsurface tubes have been identified in the zone of saturation beneath the Station. Underground cavities were reported in the logs of some test holes, but none was identified as a lava tube. Lava tubes store and transmit very large amounts of ground water. Their permeability is practically infinite.

Vesicles

Most vesicles in basalt contribute but little to its permeability and their practical hydrologic importance is minor compared to that of joints, crevices, flow-contact zones, and interstices in scoria and in cinders. Vesicles are so abundant in some layers that the rock is spongy in appearance. Some well drillers believe that spongy basalt is a copious source of water. Despite their abundance, however, the vesicles usually are not widely interconnected and do not transmit much ground water. The most spongy material generally occurs at the tops of flows, where the main source of water actually is the voids adjacent to the flow contact.

vesicular or scoriaceous basalt which has been pulverized by the drill. These kinds of materials are common at the tops of some flows. Some reported thick beds of cinders may be scoriaceous parts of aa flows.

Fresh aa — broken, blocky basalt — is among the most permeable of all types of rock. Some aa weathers readily when exposed to the atmosphere, and decomposes when acted on by subterranean water and gases. Therefore, it becomes less permeable as it ages. The aa in the Snake River basalt is geologically very young and but little altered. The physical characteristics of aa were described briefly on page 113 of part 2.

Tension Joints

Open tension joints are very common in the Snake River basalt. Shrinkage during cooling is a principal cause of tension joints in basalt (see part 2, p. 108-111), but these joints are less widely open than those produced by other causes. The vertical and lateral extent of single shrinkage joints ranges up to a few tens of feet. Shrinkage joints undoubtedly are developed in nearly all the Snake River basalt because they are typical of pahoehoe. The systems of shrinkage joints in a flow are interconnected and thus provide irregular channels for the storage and movement of ground water. The average shrinkage joint is quite narrow — on the order of a few millimeters to practically nil — and shrinkage joints alone would not transmit the large yields commonly obtained from wells in the basalt unless they extended far into the zone of saturation.

forward part of the flow moves forward with the flow and is rolled under it, much like the tread of a track-laying vehicle. When the "track" is overrun it forms a highly permeable layer. The combination of factors produces a porous, permeable zone adjacent to the contacts between flows.

Flow-contact zones are especially important water-bearers in the Snake River Plain because the Snake River basalt is made up of many thin flows. The average contact zone is thin and may transmit relatively little water, but the aggregate capacity of all the zones is large.

Interstitial Large Voids

Some Snake River basalt flows moved into place and cooled on marshy ground or under water in streams, ponds and lakes. The cooling lava formed rounded, pillow-like masses among which there are interstices because the pillows, though plastic, did not mold completely against each other before they hardened. Pillow lava was observed at a few exposures in the canyon of the Snake River between Milner and Bliss. Geologic and physical conditions obviously favored its development there. In most of the Snake River Plain, however, only a few surficial layers of basalt are exposed and pillows have not been observed in them. We have no direct evidence that pillow lava is an important component of the main mass of the Snake River basalt, though some geologists believe that pillow lava is widespread.

Volcanic cinders generally are about equal in permeability to poorly sorted sand and gravel. Cinders were found in many test holes and wells in and near the NRTS (Appendix 1, part 3). Material reported as cinders in some drillers' logs, however, actually is reddish

3. Open tension joints formed by shrinkage of the basalt during cooling, or by differential movements of the crust of a hardening basalt flow.

4. Tunnels and cavities produced by liquid lava draining out from under a hardened crust (lava tubes).

5. Vesicles and cavities formed by expanding gas bubbles during cooling of the basalt.

6. Pores, chiefly interstitial (between mineral grains).

7. Tree molds. These are largely of academic interest and have little hydrologic importance. They will not be discussed further.

Voids Adjacent to Flow Contacts

Voids along or near the contacts between flows are among the most important of the water-bearing openings in basalt. Most larger voids of this sort are interconnected and they are distributed about as widely as the flows in which they occur. The upper surface of a flow commonly is somewhat brecciated, extensively fractured, or highly irregular in configuration. A later flow that spreads over an earlier one tends to chill and solidify rapidly at its base, without completely filling the surficial voids in the underlying flow. Many flows also are somewhat brecciated along their bases. It is characteristic of slowly moving pahoehoe flows that the stiffened and fractured upper crust along the

Evidently, complexity of ground-water conditions is normal on the NRTS. The complexities at the CPP site seem unusual because we have more test holes there and know more local details than are available for other areas. The natural complexity is compounded by heavy pumping of the CPP wells, which causes drawdown of the water table there, and by the disposal of liquid waste at a well a few hundred feet distant, causing some build-up. The water level in the disposal well cannot be measured, but the build-up around it and the drawdown nearby probably cause the water-table contours to "wander" in that vicinity.

WATER-BEARING PROPERTIES OF THE SNAKE RIVER BASALT

Dense, massive basalt is nearly impermeable to water. Most flows of the Snake River basalt, however, contain primary voids which store and transmit ground water. Certain secondary voids, as in talus breccia buried beneath younger sediments or basalt, are copious water bearers. We have no evidence of occurrences of water-bearing talus breccia in or near the NRTS, so this material will not be described.

PRIMARY TYPES OF WATER-BEARING OPENINGS

The principal kinds of voids, listed in the approximate order of their relative importance, are as follows:

1. Voids adjacent to the contacts between flows or between a flow and an underlying formation.
2. Interstitial openings formed during emplacement of pillows, cinders and aa.

than that in the northern well. These water levels imply a reverse gradient of the water table (northeastward more than 10 ft/mi). The regional gradient is southwestward about 3 ft/mi. Releveling to measuring points and repeated remeasurement of the water levels failed to remove the apparent discrepancy, which must be accepted as a fact. Theories to explain this phenomenon have been entertained, but there is little evidence directly supporting them, so discussion seems pointless at this time. The anomaly probably arises partly from quasi-artesian conditions and partly from differences in the construction of the two wells.

The detailed configuration of the water table and the direction of ground-water underflow are important considerations in the vicinity of the CPP area and southwestward from there. Regional contours on the water table indicate a general southwestward gradient of the water table. We tried to trace the course of saline liquid-waste materials in the zone of saturation by drilling test holes where interception of the waste seemed likely. Lacking direct evidence of the true direction of movement, flow directly down the apparent maximum gradient had to be assumed. Six test holes were drilled at 500-foot intervals along a 2,500-foot line normal to the direction of the apparent maximum gradient of the water table. The saline waste was not found.[1/] This problem is discussed more fully in part 4 (waste disposal).

1/ After this report was written the waste liquid was found by further exploration. It was not following the apparent maximum gradient.

Owing to the prevalence of quasi-artesian conditions, predictions of the depth at which water will be struck in wells at some places on the Snake River Plain are apt to be in error by a few feet. Prediction is more accurate when phrased as a prognosis of static water level in the finished well, rather than of the depth at which water will be encountered.

In some wells in the Snake River Plain the water-level fluctuations resemble those in artesian aquifers because the water levels respond to fluctuations in barometric pressure. Noticeable barometric fluctuations do not occur in water-table aquifers where there is free communication between ground air and the atmosphere. The barometric fluctuations in basalt wells confirm that there are tight zones in the basalt above the zone of saturation, and these impede equalization of pressure between ground air and the atmosphere. Other kinds of water-level fluctuations in basalt wells, such as those caused by earthquakes and by heavy pumping, also simulate phenomena ordinarily seen only in artesian wells. Water-level fluctuations will be discussed fully in later pages.

Unexplained Phenomena

Certain anomalies in the occurrence and behavior of ground water in the NRTS cause critical problems. A case in point is the behavior of water levels in the two wells at the CF area (wells 2N-29E-1aa1 and -1db1). At times the water levels in the wells are at about the same altitude; but on most occasions when the wells were measured the water in well -1db1, the southernmost of the two, was 0.1 to 4.0 ft higher immersed in the zone of saturation in a water-bearing gravel. A drill, striking the boulder above the level of the water table, would not strike water until it broke through the bottom of the boulder. Water then would rise in the boulder, but it would not be called artesian. In an extreme case, such as that of the basalt well, it is difficult to determine whether the aquifer is artesian or quasi-artesian.

Test well 4N-30E-7ad1, illustrates a more common situation on the Snake River Plain. Water was first recognized in the hole at the depth of 366 ft. The water rose to 332 ft within about four minutes after it was struck. Twenty-four hours later the depth to water was 328 ft. The water occurs in broken basalt overlain by solid, nonpermeable basalt. The latter seemingly confined the water under quasi-artesian pressure in the permeable bed, the pressure surface of which coincides with the regional water table. Other better-than-average water-bearing zones in this well were struck in basalt at depths of 400 to 405 ft, in sandstone at 500-504 ft, and in creviced basalt at 655 ft. Table 7 of Appendix 2 shows the depth-to-water measurements made in this well during drilling.

Another example is test well 4N-30E-6ab1 for which water level measurements are shown in Appendix 2 table 6. Water was first recognized at the depth of 331 ft and rose promptly to 315 ft. The water occurred in broken basalt overlain by a local confining bed of solid, nonpermeable basalt. Additional water was struck in basalt at 605 ft and in sand at 1,400 ft. Water with a foul odor was found at a depth of 1,010 ft. In this and the previously noted test hole, the water level was quite stable during drilling except when the holes were bailed rapidly or when water-bearing zones were cased off.

complex, ground water occurs under perched, water-table, and artesian conditions. Test hole 5N-34E-9bd1, 322 ft deep, encountered water at a depth of 256 ft, but artesian water seemingly was not struck. The hole was deepened to 553 ft without finding artesian water. A private well (5N-35E-4bd2) 6 miles to the east, however, reportedly struck artesian water at a depth of about 750 ft. The water rose to about 50 ft below the land surface.

Flowing artesian wells occur in the Aberdeen-Springfield irrigation district, north of American Falls Reservoir. Other artesian areas probably will be found in the plain.

Exploratory deep drilling on the NRTS did not disclose unquestioned occurrences of artesian water. If artesian water is present, it is at depths from which recovery probably would not be economical. Also, the deep water is apt to be inferior in chemical quality, as was the case with unconfined water from the lower part of the deepest test hole (4N-30E-6ab1), which tapped ground water at numerous depths down to nearly 1,500 ft. The water from permeable zones rose no higher than the regional water table.

Test hole 2N-27E-33ac2 reached a depth of 1,065 ft before water was struck and the water rose to a static level 981 ft below the surface. By literal definition, the water is artesian. In this hole extremely tight basalt extends to a considerable depth below the level of the regional water table, and the hole actually was bone dry until water was struck. After a depth of 1,000 ft was reached the hole was undisturbed for 13 months, but it contained no water when drilling was resumed. The situation may be analogous to that created by a large boulder partly voids and extremely low hydraulic permeability. The zone of saturation often is not recognized in a drill hole in such material, because the influx of water to the well is small and does not noticeably decrease the amount of water that must be added for drilling. When a permeable zone finally is struck the hole fills to the level of the water table. This rise of the water level simulates that caused by artesian pressure, though the conditions are not truly artesian. Nonpermeable sediments in the zone of saturation sometimes cause effects similar to those of tight basalt.

The conditions that produce simulated artesian effects are here called quasi-artesian. Quasi-artesian pressure is very common in the Snake River Plain. In many instances the rise of water in a well is less than 5 ft, but 25 ft is not unusual (Appendix 2, table 5), and rises as great as 50 ft are reported by drillers.

Water occurs in some parts of the Snake River Plain under truly artesian conditions, most commonly in areas where widespread sheets of impermeable sediments overlie permeable basalt. At Rupert, for example, a thick blanket of nonpermeable lacustrine sediments overlies basalt, in which water is confined under pressure. Water was struck in basalt at a depth of 500 ft in one well and rose to a static level about 90 ft below the land surface. In a few localities on the plain, water rises more than 1,000 ft above the level at which the aquifer is tapped. At some of these places the confining material is tight basalt.

Water is confined under pressure beneath impermeable basalt in parts of the Mud Lake basin also, and flowing and nonflowing artesian wells are numerous there. In that basin, which is hydrologically northeastward into the southern part of T. 6 N., Rs 30-31 E. Owing to the low permeability of the perched aquifer and the small amount of water in storage, the perched water is not a likely source of supply for facilities on the NRTS.

In the MTR area, low-level radioactive liquid waste is discharged to the ground through an infiltration pond. The waste water built up a perched zone of saturation that may have underlain all or much of the MTR site and some adjacent land. In an excavation in the adjacent ETR area, perched water was struck in a foundation boring in basalt at a depth of 19 feet. The driller reported that, with two feet of water in the hole, he could not bail the hole dry. The water evidently is perched on impermeable basalt.

The perched zone of saturation beneath the waste-disposal pond at the MTR site spread outward to an unknown distance beyond the borders of the pond and appeared in shallow test holes 25 and 50 feet distant. The occurrence of artificial perched water has special interest because, on the basis of geologic conditions, the writers had predicted that a perched zone of saturation would develop at the MTR site if much waste water were disposed of in the gravel. Similar developments may be expected at other places where the geology is similar.

Quasi-Artesian and Artesian Water

The ground water in the Snake River basalt and in sediments associated with it behaves at some places as though it were under artesian pressure. Some layers of the basalt are dense and massive, having few the water-table maps.[1/] On that basis, underflow in the northern NRTS is generally southward. The southward bulging of the contour that passes through the Birch Creek playa probably is caused by underflow increments from the Birch Creek valley. A similar southward bulge south of the mouth of the Little Lost River valley is caused by underflow from that basin.

South of T. 4 N. the regional trend of underflow is west of south, the western component being most pronounced in the southwestern part of the Station. Along the eastern part of the Station, south of Circular Butte, the contours show that water from the Mud Lake basin enters the NRTS, moving southwestward across the Station boundary. Along the entire southern boundary of the Station, underflow is southwestward.

Perched Water

Perched ground water was struck in well 6N-31E-27bal, in the northern NRTS, where it occurs in alluvium or lake beds. The depth to the water table was about 60 ft and the saturated thickness was about 30 ft. No perched water was found in other test holes or wells. Well 5N-30E-21bcl, which was drilled many years ago and for which drilling records are not available, probably taps perched water.

Perched water was not given any special study. It is apt to be present near the northwestern edge of the NRTS from T. 5 N., R. 29 E.

1/ The nature of ground-water underflow and of factors that control its speed and direction are discussed in detail on pages 143-151. The principles discussed there should be born in mind when using water-table maps or basing plans on them.

In Tps. 3 and 2 N. the water-table gradient is rather gentle, ranging between about 1 and 5 ft per mi. The general trend of the contours is southeastward.

In most of the area represented by the accompanying maps, wells and test holes are sparse and measurements of the depth to water reveal but few local irregularities in the water table. Information about the local configuration of the water table, however, is important, especially in relation to the disposal of liquid wastes in the ground. This topic will be fully discussed later.

Fluctuations in the position and configuration of a water table are clues to events and factors that influence the aquifer. The position and configuration may be changed, for example, by pumping or by recharge. The accompanying maps show differences in the water-table contours at successive times, and some of these may represent actual changes in the water table. Some differences, however, are only apparent, because the instantaneous conditions at any one time in the whole area cannot be determined. We have no satisfactory means for distinguishing real from apparent changes. At any rate, a precise water-table map, if it could be made, would represent only a transitory, instantaneous condition which never would be repeated.

Ground-Water Underflow

In order to interpret the broader aspects of ground-water underflow beneath the NRTS, it is acceptable to assume that the direction of underflow corresponds roughly to the apparent maximum gradients shown by

The configuration of the water table at the NRTS is quite irregular. The contours on the maps are smooth and gently curving because the depth to water can be measured only at widely scattered points which do not reveal the irregularities. The irregularities are not great enough at most places to cause much error in an estimate of the depth to water at a new site. Where the configuration of the water table is fairly regular and where water-level control points are spaced at intervals of 5 to 10 mi, the error in estimates ranges between a few inches and 5 ft. Where the configuration of the water table is less regular or where control data are poor, errors of estimate may be larger. Additional factors that affect the accuracy and use of water-table maps are discussed in the section on "Ground-water Underflow."

Little is known about ground water in the area immediately north of the IET site in the ANP area. The water table seemingly is very flat there and at places may slope less than a foot per mile. The general slope is southward, but locally the maximum slope departs by at least 90 degrees from that direction.

From Pole-Line Road southward through T. 5 N. the general slope of the water table is southward, ranging between about 7 and 10 ft per mi in most of the row of townships, and the contours trend about eastward. To the west, in T. 5 N., R. 29 E. the gradient is steeper, ranging up to about 12 ft per mi.

In T. 4 N. the water-table gradient becomes flatter toward the south, ranging between about 2 and 10 ft per mi. The configuration of the water table is less simple in this township than it is to the north, as is shown by the broadly sweeping curves of the contours.

both are drawn on a map, the approximate depth to water at any location can be determined readily by inspection. Maps accompanying this report (pl. 1, part 2, and pl. 2, part 3) represent the general configuration and altitude of the piezometric surface in the area studied. They show that the depth to the water table ranges from about 200 ft in the northern NRTS to more than 800 ft in the southeastern part near Twin Buttes (pl. 2).

Position and Form of the Water Table

The altitude of the water table ranges from about 4,580 ft above msl in the northern part of the Station to about 4,405 ft near the southwest corner (pl. 2). The water-table contours on plate 2 represent the generalized altitude and configuration of the water table throughout the NRTS in April 1953. Maps for other times, some months apart, are included in appendix 2 (pls. 1 and 3). The maps are broadly generalized, being based on nonsimultaneous measurements in widely spaced wells and test holes. Each represents an average condition which is reasonably correct for the NRTS as a whole but is not correct in pinpoint detail. The maps are useful and satisfactory so long as their limitations are recognized.

Most of the measurements on which the contour maps are based were made on the days or in the months specified on the maps, but a few were made at other times. The off-date water levels were adjusted to the dates of the maps by interpolation between successive measurements.

GENERAL CONDITIONS OF OCCURRENCE

The Snake River basalt, which underlies all but a few square miles of the northwestern part of the Station, is the only reliable aquifer from which large amounts of water can be obtained perennially on the NRTS. The thickness of the basalt is believed to be at least 2,000 feet, but it may be considerably more.

The ground water beneath the NRTS is largely unconfined in a regional zone of saturation. Locally, the water is quasi-artesian, and true artesian conditions may prevail at some places or at great depth. Perched ground water occurs at some places in the northwestern part of the Station. An artificially created perched zone of saturation is present in the vicinity of the MTR plant.

Depth to Water

The depth to the water table (normal pressure surface) in an aquifer can be measured readily where the aquifer is tapped by wells, test borings, or other openings. The general form of the water table commonly is represented on a map by isopiestic lines. These are lines drawn on the pressure surface, connecting points of equal altitude, and control is obtained by direct measurements of water levels in wells. The accuracy of the map depends upon the number and spacing of the wells and the time and accuracy of the measurements.

The depth to the water or pressure surface at any point is the difference between the land-surface altitude and the altitude of the isopiestic line at that point. If land-surface and water-table contours

Table 8.--Chemical analyses of water from the Big Lost River

(Analyses by U. S. Geological Survey and Idaho Agricultural Experiment Station. Parts per million, except where noted otherwise.)

Location	BUTTE COUNTY 1 mile north of Arco	BUTTE COUNTY Northwest of CPP area	CUSTER COUNTY At Mackay
Date of collection	10/19/49	9/8/51	1948
Temperature (°F)	41	68	-
Silica (SiO_2)	-	14	-
Iron (Fe):			
Dissolved	-	.04	-
Calcium (Ca)	-	45	38
Magnesium (Mg)	-	16	10
Sodium (Na)	-	11	5.7
Potassium (K)	-	4.6	10.9
Bicarbonate (HCO_3)	252	206	139
Sulfate (SO_4)	43	24	19
Chloride (Cl)	7	6.6	9.6
Fluoride (F)	-	.3	-
Nitrate (NO_3)	-	.00	-
Total dissolved solids:			
Ppm	-	216	174
Tons per ac-ft	-	.29	.24
Hardness as $CaCO_3$:			
Total	-	178	125
Noncarbonate	-	9	11
Specific conductance ($K \times 10^5$ at 25°C.)	441	367	290
pH	-	8.1	-