Simulation Analysis of the Unconfined Aquifer, Raft River Geothermal Area, Idaho-Utah

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[Factors for converting U.S. customary units to the International System of Units (SI) are given below in four significant figures. However, in the text the metric equivalents are shown only to the number of significant figures consistent with the values for the U.S. customary units]

ft (feet)	0.3048	m (meters)
ft/d (feet per day)	.3048	m/d (meters per day)
ft/mi (feet per mile)	.1894	m/km (meters per kilometer
ft/s (feet per second)	.3048	m/s (meters per second)
ft²/d (feet squared per day)	.09290	m²/d (meters squared per day)
(ft³/d)/mi² (cubic feet per day per square mile	.01093	[(m³/d)/km²] (cubic meters per day per square kilometer
in (inches)	25.40	mm (millimeters)
mi (miles)	1.609	km (kilometers)
mi² (square miles)	2.590	km² (square kilometers)
acre-ft/yr (acre-feet per year)	1.233×10^{-3}	hm³/yr (cubic hectometers per year)

SIMULATION ANALYSIS OF THE UNCONFINED AQUIFER, RAFT RIVER GEOTHERMAL AREA, IDAHO-UTAH

By WILLIAM D. NICHOLS

ABSTRACT

This study covers about 1,000 mi² (2,600 km²) of the southern Raft River drainage basin in south-central Idaho and northwest Utah. The main area of interest, approximately 200 mi² (520 km²) of semiarid agricultural and rangeland in the southern Raft River Valley that includes the known Geothermal Resource Area near Bridge, Idaho, was modelled numerically to evaluate the hydrodynamics of the unconfined aquifer. Computed and estimated transmissivity values range from 1,200 feet squared per day (110 meters squared per day) to 73,500 feet squared per day (6,830 meters squared per day). Water budgets, including ground-water recharge and discharge for approximate equilibrium conditions, have been computed by several previous investigators; their estimates of available ground-water recharge range from about 46,000 acre-feet per year (57 cubic hectometers per year) to 100,000 acre-feet per year (123 cubic hectometers per year).

Simulation modeling of equilibrium conditions represented by 1952 water levels suggests: (1) recharge to the water-table aquifer is about 63,000 acre-feet per year (77 cubic hectometers per year); (2) a significant volume of ground water is discharged through evapotranspiration by phreatophytes growing on the valley bottomlands; (3) the major source of recharge may be from upward leakage of water from a deeper confined reservoir; and (4) the aquifer transmissivity probably does not exceed about 12,000 feet squared per day (3,100 meters squared per day). Additional analysis carried out by simulating transient conditions from 1952 to 1965 strongly suggests that aquifer transmissivity does not exceed about 7,700 feet squared per day (700 meters squared per day). The model was calibrated using slightly modified published pumpage data; it satisfactorily reproduced the historic water-level decline over the period 1952–65.

INTRODUCTION

Several proposals have been advanced for the development of geothermal resources in the upper Raft River Basin, Idaho-Utah (fig. 1). One proposal (Dart and others, 1975) recommends the generation of 10 MW (megawatts) of electric power using an estimated 7,100 acre-ft/yr (8.7 hm³/yr) geothermal fluid at 140°C. The temperature of this fluid will be reduced by heat loss to an organic liquid (probably isobutane) in the proposed system heat exchanger. The cooled geothermal water then would be returned to either the geothermal reservoir or a confined aquifer at an intermediate depth.

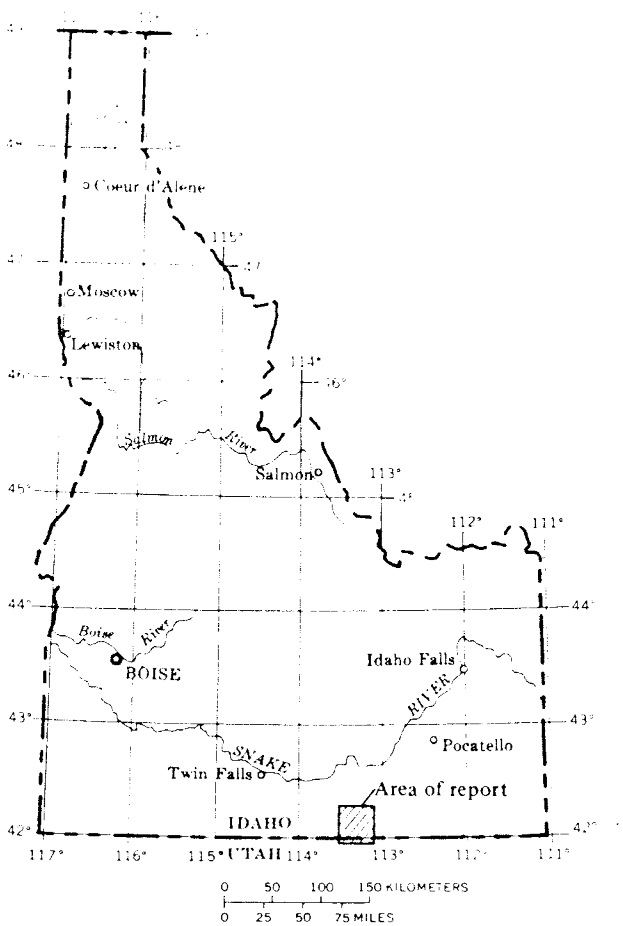


FIGURE 1.—Index map of Idaho and northern Utah showing area covered by this report.

Electric power production using geothermal fluid could require the use of shallow ground water for cooling purposes. One proposal (Dart and others, 1975) estimated that about 32,300 to 43,500 acre-ft/yr (39.8 to 53.6 hm³/yr) of cooling fluid may be required. It is anticipated that ground water used for cooling will be returned to the aquifer, thus providing that use of the water will be nonconsumptive except for that evaporated in the cooling process. Final design and operating criteria, determined in light of legal and environmental constraints, will determine the volume of water needed from the shallow ground-water system for cooling.

PURPOSE AND SCOPE

This study was undertaken to define quantitatively the geohydrologic properties and hydrodynamics of the shallow aquifer system in the southern Raft River Valley and to determine if there is any significant hydrodynamic interdependence between this system and the deeper geothermal system known to underlie at least part of the area. Any such interdependence between the two systems would be a significant factor in concurrent development of the geothermal reservoir and the shallow ground-water system.

The scope of this investigation includes the following:

(1) Assemble geohydrologic data developed during several previous studies in the Raft River Basin.

(2) Construct and calibrate a computer-simulation model of the shallow ground-water system with emphasis on the area of anticipated geothermal development. Calibrate the model through use of water-budget and water-yield values computed during previous investigations.

(3) Collect new water-level data and compute estimated pumpage from newly collected electrical power-consumption data for the years 1966–75 to extend, if possible, the period of model calibration beyond 1965, the last year for which goehydrologic data have already been published.

(4) Using the simulation model, determine the volume and distribution of predevelopment recharge and discharge within the limits of previously determined water-budget estimates.

(5) Provide simulation methods for quantitative evaluation of the effect of increased development of the shallow ground-water system.

STUDY AREA

The area of interest covers about 1,000 mi² (2,600 km²) and encompasses all the Raft River drainage basin upstream from Malta, Idaho (fig. 2). This includes the southern part of the main valley of the Raft River between Malta and The Narrows (fig. 2), hereafter referred to as the southern Raft River Valley subbasin, as well as the Yost-Almo subbasin, the Elba subbasin, and Junction Valley (fig. 2). The area of principal hydrologic interest comprises the southern Raft River Valley subbasin and the Yost-Almo subbasin. Geothermal-resource exploration and evaluation is now centered on an area in the upper (southern) end of the southern Raft River Valley subbasin about 14 miles (23 km) south of Malta, Idaho, and about 6 miles (10 km) northeast of The Narrows near Bridge, Idaho (fig. 2).

The modeled area covers 200 mi² (20 km²) and is in the southern Raft River Valley subbasin (fig. 2). It extends from Malta on the north

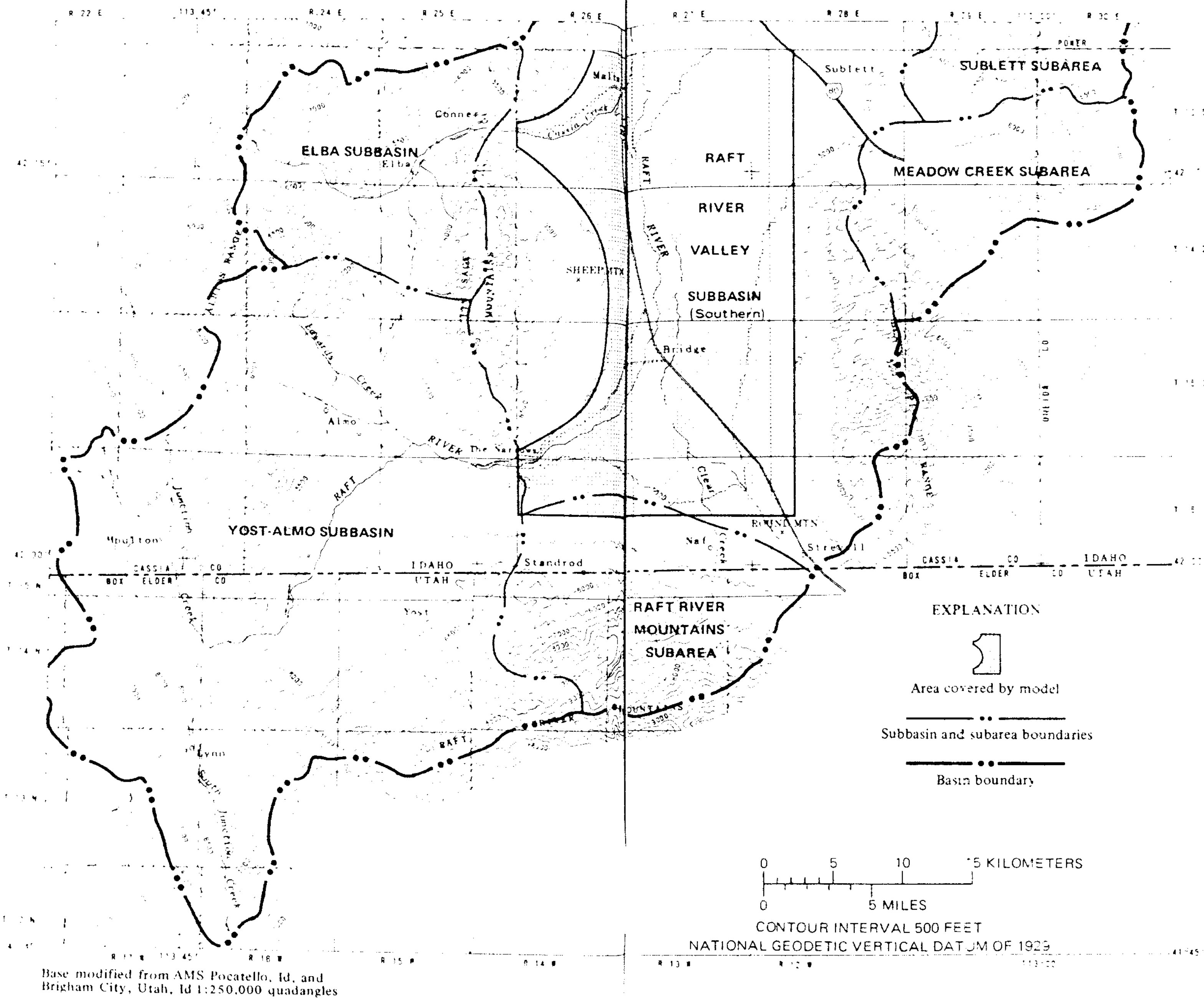


FIGURE 2.—Subbasins and subareas of the Raft River Basin and the area covered by the simulation model.

almost to Naf, Idaho, at the base of the Raft River Mountains on the south. The western boundary extends northeastward from The Narrows to the area about a mile (1.6 km) east of Sheep Mountain and then northwestward almost to Connor on Cassia Creek. The eastern boundary is about 2 miles (3 km) west of the base of the more rugged parts of the Black Pine Mountains; it extends due northward from the vicinity of Round Mountain near Strevell (fig. 2). The modeled area includes the area of current geothermal exploration and development.

PRINCIPAL PREVIOUS INVESTIGATIONS

Several previous geologic and hydrologic studies include all or parts of the area of present interest. Two of these studies summarized or compiled most of the data needed for the development and calibration of a simulation model, and these constitute the primary previous investigations upon which the present study is based. The first comprehensive study of the water resources of the Raft River Basin (Nace and other, 1961) covered the period 1948–55 and provided well data and estimates of all elements of the hydrologic budget. By 1967, significant new geologic and hydrologic data had become available, and a second comprehensive report was prepared (Walker and others, 1970). That study redescribed and redefined the geohydrologic framework of the basin using newly acquired data on ground-water pumping, change in water level, and use of irrigation water, and it covered the period 1948–66. It also presented new data for the reevaluation and refinement of elements of the hydrologic budget and independently developed a new budget for the basin. Both reports (Nace and others, 1961; Walker and others, 1970) contain references to previous investigations.

Recent deep test drilling in the southern part of the Raft River Basin indicates that the basin is underlain by as much as 5,250 feet (1,600 m) of consolidated and unconsolidated sedimentary deposits. Basement rocks beneath most of the valley are igneous and metamorphic rocks of Precambrian age and sedimentary rocks of early Paleozoic age. These rocks are overlain by as much as 5,000 feet (1,500 m) of fluvial and lacustrine tuffaceous siltstone, sandstone, and conglomerate of the Salt Lake Formation. This formation, which has been divided into several units, is of Miocene and Pliocene age. The Salt Lake Formation is overlain by as much as 750 feet (230 m) of sand and gravel beds with some intercalated layers of silt and clay that compose the Raft Formation of Pleistocene age. This formation is confined largely to the valley trough and thins rapidly on the west side of the valley axis; it also thins rapidly south of Bridge, pinching out several miles north of the Raft River Mountains. Overlying the

Raft Formation, or where this unit is absent (the Salt Lake Formation), are Pleistocene and Holocene alluvial deposits of sand, gravel, silt, and loess as much as 200 feet (60 m) thick.

The alluvial deposits, the Raft Formation where present, and the upper part of the Salt Lake Formation make up the principal aquifer in the southern Raft River Valley. This aquifer, discussed in detail in the following section of this report, can be considered a water-table aquifer for practical purposes even though the water is locally confined. Data obtained from test wells, and geothermal exploration wells drilled west and southwest of Bridge, and the interpretation of borehole geophysical logs obtained from these wells indicate the presence of at least one, and probably several, deeper confined aquifers beneath the water-table aquifer. The unconfined aquifer extends to a depth of about 800 to 850 feet (240 to 260 m). At greater depths are confined aquifers: from about 1,100 to 1,400 feet (335 to 425 m), from 1,800 to 2,200 feet (550 to 670 m), and perhaps from 2,700 to 3,300 feet (820 to 1,000 m). Underlying these aquifers at a depth of 4,400 to 4,500 feet (1,340 to 1,370 m) in the area southwest of Bridge is the approximate upper boundary of the geothermal reservoir. The depth to the upper boundary under other areas of the valley is not known but is thought to be about the same.

The relations among the several aquifers and between the aquifer systems and the geothermal reservoir are poorly known. The scant data available for the aquifer at 1,100 feet (335 m) indicate that the head in this aquifer is higher than the head in the overlying unconfined aquifer; it is inferred from this that water in the shallow confined aquifer leaks upward and recharges the water-table aquifer. Data (on file with U.S. Geological Survey, Boise, Id.) obtained from the geothermal exploration holes southwest of Bridge indicate the head in the geothermal reservoir is considerably higher than the head in either the water-table aquifer or the confined aquifer at 1,100 to 1,400 feet (335 to 425 m). On the basis of this evidence, it is inferred that the head is greatest on the geothermal reservoir and decreases progressively to the minimum observed in the water-table aquifer. In other words, the head in any given aquifer or water-bearing zone is lower than it is in the aquifer beneath and higher than in the aquifer above. Each of the aquifers in the southern Raft River Basin is therefore recharged, in part at least, by upward leakage from underlying aquifers.

ACKNOWLEDGMENTS

The Raft River Rural Electric Cooperative furnished records of power consumption for irrigation wells and maps showing the distribution of the wells. Aerojet Nuclear Company furnished data and

reports concerning specific proposals and plans for geothermal-resource development and powerplant operation, as well as borehole geophysical data. Aquifer-test data were provided by Alberto G. Morilla, University of Idaho, Moscow, Idaho, and by Dale R. Ralston of the Idaho Bureau of Mines and Geology. The cooperation of all these agencies and associated individuals is gratefully acknowledged.

GEOHYDROLOGY OF THE SHALLOW GROUND-WATER SYSTEM

The following discussion concentrates on the specific details of the geohydrology of the shallow ground-water system in the southern Raft River Valley subbasin as they apply to the simulation model.

BOUNDARIES, GEOMETRY, AND HYDRAULIC PROPERTIES

The shallow ground-water system underlies the entire area modeled. Within this area no natural geologic or hydrologic lateral boundaries of the shallow aquifer can be defined on the basis of data now available. Arbitrary boundaries as shown in figure 2 were selected for operational purposes and generally coincide with aquifer limits suggested in Walker and others (1970); they are 2 to 6 miles (3 to 10 km) from areas of significant irrigation pumpage. Lack of data regarding the nature of the aquifer made it impracticable to extend the boundaries any greater distance east, south, or west into or beneath the bordering mountain ranges. Northward extension of the modeled area beyond the boundary shown was avoided because of pumping centers 3 to 5 miles (5 to 8 km) north of Malta. On the basis of the configuration of the 1952 water table shown in figure 3, the east, south, and west boundaries are believed to be recharge boundaries, and the north boundary is believed to be a discharge boundary.

The thickness of the shallow aquifer is poorly known throughout the area of interest including the southern Raft River Valley subbasin. Nace and others (1961, p. 96) suggest that it may be as much as 1,200 feet (370 m) thick near Malta. Walker and others (1970, fig. 8) suggest that the shallow aquifer may range in thickness from about 200 feet (60 m) on the east side of the valley along the base of the Black Pine Mountains to about 1,400 feet (430 m) near Malta. Aquifer thickness used in the present study is shown in figure 3; values for thickness are based mostly on the isopach map of Walker and others (1970, fig. 8) showing the combined thickness of alluvium, basalt, and Raft Formation, and on recent test-hole data and borehole geophysical logs in the part of the area south of Bridge. The aquifer as shown

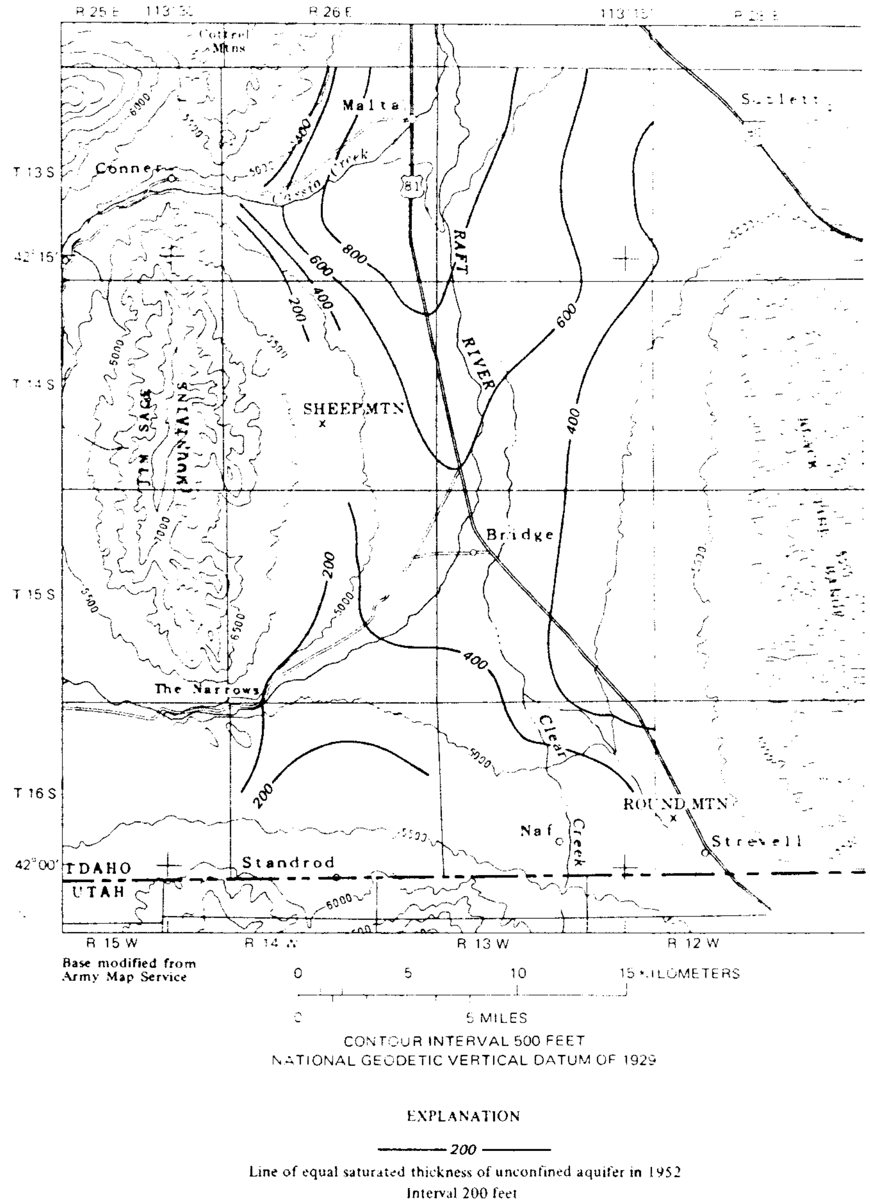


FIGURE 3.—Saturated thickness of the unconfined aquifer in 1952, southern Raft River Valley subbasin.

in figure 3 may include part of the upper unit of the Salt Lake Formation.

Ground water in the shallow aquifer occurs mainly under unconfined or water-table conditions. Even in parts of the aquifer penetrated by the deepest wells in the subbasin, the water is at most poorly confined (Walker and others, 1970, p. 58), although ground water is confined to semiconfined locally where the primary water-bearing zone is capped by small local and discontinuous deposits of low permeability. Perched water occurs locally in small areas of the valley bottom. These conditions exist during the irrigation season and in some cases persits for several months afterward (Walker and others, 1970, p. 58).

Aquifer permeability, transmissivity, and specific yield have been computed or estimated by several investigators (table 1). All methods used by Nace and others (1961, p. 83–95) to compute or estimate transmissivity give a range of values from about 7,000 feet squared per day (620 m²/d) to about 70,000 ft²/d (6,200 m²/d) An average hydraulic conductivity of about 130 feet per day (40 m/d) is suggested for the uppermost 200 feet (60 m) of the aquifer (Nace and others, 1961, p. 96). Values of specific yield ranging from 0.11 to 0.15 were computed from data obtained from an aquifer test near Malta (Nace and others, 1961, p. 87–91); computed transmissivity values range from 22,700 ft²/d (2,110 m²/d) to 26,700 ft²/d (2,480 m²/d). Walker and others (1970, p. 61, 63) suggested an average hydraulic conductivity of 130 ft/d (40 m/d) for the upper 200 feet (60 m) of the alluvial aquifer in the Raft River Valley subbasin. They also indicated (1970, p. 63) that the entire thickness of the aquifer may have an average hydraulic conductivity of about 40 ft/d (12 m/d), and they estimated (1970, p. 77 and fig. 19) that the specific yield of sediments in the Raft River Valley subbasin ranges from 0.02 to 0.2.

Morrilla and Ralston (1976) conducted several aquifer tests in the

TABLE 1.—*Summary of aquifer test results*

Source	T = transmissivity (ft²/d)		K = hydraulic conductivity (ft/day)		Specific yield (percent)	
	Minimum	Maximum	0–200 ft	Entire thickness	Minimum	Maximum
Nace and others:						
All methods	7,000	70,000	130			
Malta Land and Irrigation Co. test	22,700	26,700			0.11	0.15
Walker and others			130	40	0.02	0.22
Morrillia and Ralston:						
Test in Raft River Valley subbasin	13,400	73,500			1.8×10^{-3}	2.8×10^{-2}
Test in Yost-Almo subbasin*	3,200					
This study:						
From Raft of River Valley subbasin data of Morrilla and Ralston	1,300	110,000		33–180		
From Yost-Almo subbasin data	*3,000			*10		
From Malta Land and Irrigation Co. test data of Nace and others	23,900					
From specific-capacity data	1,200	12,000				

*Only one observation well used in test.

southern end of the Raft River Valley subbasin and one test in the Yost-Almo subbasin and obtained transmissivity values that ranged from 13,400 ft²/d (1,240 m²/d) to 73,500 ft²/d (6,830 m²/d) for the Raft River Valley subbasin and a value of 3,200 ft²/d (300 m²/d) for the Yost-Almo subbasin. Values for the storage coefficient range from 1.8 x 10 $^{-3}$ to 2.8 x 10 $^{-2}$. Recently obtained borehole geophysical data suggest that the aquifer may be as thick as 400 feet (120 m) where the aquifer tests were made. Thus, the average hydraulic conductivity of the entire thickness of the aquifer may range from about 33 ft/d (10 m/d) to 180 ft/d (55 m/d).

Aquifer test data collected by Nace and others (1961) and Morrilla and Ralston (1976) were reevaluated during the present study. It was determined that, with two exceptions, the data collected from each observation well during the tests do not define a unique time-drawdown curve that can be used to compute a value of transmissivity. Using equations and type-curves developed by N. S. Boulton and R. W. Stallman (Lohman, 1972, p. 34–40) for the analysis of aquifer tests in water-table aquifers, one can compute transmissivity values that are as much as one order of magnitude smaller to about one-half order of magnitude larger than the values derived by Morrilla and Ralston (1976). Similar analyses of two of the three data sets collected by Nace and others (1961) are completely nondiagnostic because the duration of the tests was too short.

Data collected by Nace and others (1961) from one observation well during their test in the southern Raft River Basin were found to describe uniquely a time-drawdown type curve from which a transmissivity value of 23,900 ft²/d (2,220 m²/d) was calculated using the Boulton curves given by Lohman (1972). The value is virtually identical with that computed by Nace and others (1961) with the same data using different methods. The observation well from which these data were obtained is 31 feet (9.4 m) deep and 4 feet (1.2 m) in diameter, the same depth and diameter as the pumped well. Both wells penetrate about 3 feet (0.9 m) of soil and 28 feet (8.5 m) of gravel; at least 2 feet (0.6 m) of clayey and silty sediment underlies the gravel. These deposits are probably alluvial deposits of Holocene age. The significance of the computed value of transmissivity with respect to the entire aquifer thickness is uncertain because the full circumstances of the test are not known. The pumping well is very near the Raft River, which may have been a source of recharge during the test.

The test conducted by Morrilla and Ralston (1976) in the Yost-Almo subbasin also yielded data that uniquely defined a type curve from which the transmissivity could be calculated using the method for water-table aquifers outlined by Lohman (1972). Again, the recalculated value was virtually the same as the 3,200 ft²/d (300 m²/d) computed by Morrilla and Ralston (1976). The stratigraphy of the test site was not described by Morilla and Ralston, but the wells, which are 263 feet (80 m) and 253 feet (77 m) deep, are believed to penetrate only the Salt Lake Formation, probably the lower member.

Transmissivity values also were estimated during this study from specific-capacity data using a method developed by Hurr (1966). The estimated values range from about 1,200 ft²/d (110 m²/d) to 12,000 ft²/d (1,120 m²/d) and average 4,200 ft²/d (390 m²/d). These are significantly lower than most values computed by other investigators using test data obtained from observation wells.

The variations between values of estimated and computed transmissivity reflect not only areal differences in aquifer thickness but also horizontal and vertical differences in hydraulic conductivity. The upper 50 to 200 feet (15 to 61 m) of aquifer in the Raft River Valley may have a hydraulic conductivity as high as 130 ft/d (40 m/d). Below these depths, hydraulic conductivity may range as low as 5 to 10 ft/d (1.5 to 3 m/d), as is suggested by the aquifer test in the Yost-Almo subbasin. The average transmissivity of the entire thickness of water-bearing material included in the unconfined aquifer, as used in this study, is thought to be significantly lower than that suggested by results of tests using observations from wells, nearly all of which are less than 300 feet (91 m) deep. This interpretation is supported, though not proven, by mathematical simulation analysis. (See the sections "Transmissivity" and "Credibility of Results.")

THE WATER TABLE, 1952–76

Most of the wells in the southern Raft River Valley subbasin are in a narrow belt extending about 1 or 2 miles (1.6 to 3.2 km) on either side of the Raft River in the central part of its valley. Stearns and others (1938, pl. 1) showed a water-table map for part of the Raft River Valley for 1928–29, but Nace and others (1961, pl. 5) showed a generalized configuration of the water table for the period October-November 1952 in a large part of the subbasin (fig. 4). Because water-level altitudes based on depth-to-water measurements made in 1952 are in general agreement with scattered measurements made intermittently since 1928, the 1952 surface is assumed to approximate closely the altitude and configuration of the predevelopment water table. The slope of the water table in the lowland part of the southern Raft River Valley subbasin ranges from about 30 to 40 ft/mi (5.6 to 7.6 m/km) near The Narrows, to about 20 ft/mi (3.8 m/km) directly north of Bridge, and to about 15 ft/mi (2.8 m/km) directly south of Malta.

Water levels were measured in the Raft River Valley subbasin again in 1961 (Mundorff and Sisco, 1963, pl. 1) and in the Elba, Yost-

hm³) in 1948 to an estimated 10,200 acre-ft (12.6 hm³) in 1952; it increased to about 59,100 acre-ft (72.9 hm³) in 1961, then decreased to an estimated 55,200 acre-ft (68.1 hm³) in 1965 (table 2). Yearly pumpage for irrigation in the southern Raft River Valley subbasin was anomalously high in 1966, amounting to an estimated 84,100 acre-ft (103 hm³), because precipitation in that year was only about 60 percent of the long-term mean annual precipitation. The estimated pumpage for irrigation shown in table 2 for the period 1952–65 is taken from Walker and others (1970). The estimates are based on power-consumption and unit-power-consumption data developed by Nace and others (1961), Mundorff and Sisco (1963), Haight (1965), and Walker and others (1970).

Records for electric power used by irrigation-well pumps were obtained for the years 1967 through 1974 through the courtesy of the Raft River Rural Electric Cooperative. Attempts were made to compute irrigation pumpage using these data, but reliable associated data related to well efficiency, unit-power consumption, and lift were not available. In addition, sprinklers have gradually been introduced randomly into the area since about 1969, increasing the lift requirements of individual wells by 90 to 220 feet (27 to 67 m) but decreasing the quantities of water used by unknown amounts. These added lift demands increased power consumption without a concurrent increase in irrigation pumpage. Excessively large increases in power consumption for individual wells or given areas could not be identified, and the impact of sprinklers on the analysis could be compensated for only in a very general way. Consequently, estimated irrigation pumpage shown in table 3 may be too large.

WATER BUDGET

Predevelopment water-budget analyses of the Raft River Basin by previous investigators (Nace and others, 1961; Walker and others, 1970) have been reevaluated and are considered to provide reasonable estimates of upper and lower limits of net recharge of ground water before development. The estimates of water yield and ground-water outflow have also been revised to reflect discharge by significant phreatophyte growth in areas of shallow ground water in the Raft River Valley subbasin before development of the shallow aquifer. The result has been a large decrease in the estimated volume of water leaving the basin as underflow and a proportional decrease in the estimate of transmissivity required to accommodate the estimated underflow. The revised estimates of evapotranspiration and underflow have decreased the water yield, as defined by Nace and others (1961, p. 45) and Walker and others (1970, p. 40), in the lowland area of the Raft River Valley subbasin. The corresponding change in estimated

TABLE 2.—*Pumpage, in acre-feet, for 1952 through 1965*

[From Walker and others, 1970, and unpublished records on file in District office; totals rounded to three significant figures]

Location	1952	1953	1954	1955	1956	1957	1958	1959	1960	1961	1962	1963	1964	1965
13S/25E-32	0	0	0	0	0	0	0	0	0	0	0	70	180	[illegible]
13S/26E- 1	80	97	85	128	124	92	90	126	267	286	365	514	487	[illegible]
13S/26E-12	17	14	0	24	27	24	0	16	20	180	353	612	658	[illegible]
13S/26E-13	280	384	629	1,415	1,323	939	847	974	1,961	1,367	1,180	1,126	705	[illegible]
13S/26E-14	13	25	36	49	56	52	0	35	47	39	41	50	42	[illegible]
13S/26E-20	0	0	0	0	0	0	0	0	48	32	27	24	13	[illegible]
13S/26E-22	0	0	0	406	427	356	407	588	1,258	946	900	974	724	[illegible]
13S/26E-23	0	0	0	214	251	235	304	301	576	426	396	419	302	[illegible]
13S/26E-24	0	0	0	0	0	0	0	0	944	0	0	0	0	[illegible]
13S/26E-26	0	0	0	64	92	102	152	170	339	244	218	221	149	[illegible]
13S/27E- 2	0	0	0	0	3,181	3,042	3,156	2,164	1,798	1,656	1,850	1,657	1,866	[illegible]
13S/27E- 6	403	454	783	593	511	646	810	1,257	1,741	1,265	1,093	733	585	[illegible]
13S/27E- 7	34	69	543	710	357	303	272	935	1,520	1,201	1,153	888	859	[illegible]
13S/27E- 9	0	0	0	0	0	0	0	0	0	273	433	553	895	[illegible]
13S/27E-10	0	0	0	0	0	0	0	0	0	1,089	1,329	1,308	1,810	[illegible]
13S/27E-11	0	0	0	0	0	0	0	0	0	1,935	2,070	1,785	1,947	[illegible]
13S/27E-14	0	0	0	0	0	0	0	2,152	2,272	2,493	3,163	3,111	3,796	[illegible]
13S/27E-16	0	0	0	0	601	676	840	1,879	2,856	2,155	1,958	1,408	1,249	[illegible]
13S/27E-17	0	0	0	0	0	0	0	0	0	0	0	0	0	[illegible]
13S/27E-18	366	410	167	378	220	216	226	259	318	322	387	367	434	[illegible]
13S/27E-19	101	0	105	107	64	64	68	116	165	303	482	541	717	[illegible]
13S/27E-20	402	398	131	1,530	698	520	389	163	32	40	55	57	72	[illegible]
13S/27E-23	0	0	0	0	0	0	0	0	0	0	0	1,139	1,500	[illegible]
13S/27E-29	411	385	911	880	409	313	245	158	121	157	218	228	290	[illegible]
13S/27E-30	762	772	1,275	1,811	1,180	1,267	1,428	1,270	1,344	1,245	1,397	1,256	1,419	[illegible]
31S/27E-31	0	302	523	375	298	361	443	420	465	406	433	372	403	[illegible]
13S/27E-32	929	1,125	2,260	2,240	1,558	1,750	2,038	1,590	1,513	1,420	1,611	1,465	1,667	[illegible]
13S/27E-33	792	785	1,502	676	423	442	485	387	375	370	438	411	481	[illegible]
13S/28E- 3	0	0	0	0	80	80	80	100	570	1,970	2,360	2,330	1,545	[illegible]
13S/28E- 7	0	0	0	0	0	0	0	0	0	0	0	0	239	[illegible]
13S/28E-12	0	0	0	0	0	0	0	0	0	0	0	0	576	[illegible]
14S/27E- 3	0	0	0	0	0	0	168	0	0	0	0	0	0	[illegible]
14S/27E- 4	1,303	0	2,053	1,808	2,253	2,488	3,063	2,756	2,223	2,451	1,970	1,772	1,435	[illegible]
14S/27E- 5	0	0	0	0	427	409	423	494	508	641	594	623	595	[illegible]
14S/27E- 6	95	169	374	550	586	580	659	1,186	1,528	2,086	2,068	2,296	2,308	[illegible]
14S/27E- 7	219	341	1,002	1,052	934	778	748	1,113	1,322	1,654	1,521	1,581	1,500	[illegible]
14S/27E- 8	268	416	484	476	882	1,168	1,598	2,589	3,199	3,692	3,130	2,994	2,609	[illegible]
14S/27E- 9	214	355	1,543	2,036	2,964	3,560	4,620	4,912	4,688	5,276	4,347	4,026	3,380	[illegible]
14S/27E-16	0	249	598	1,049	831	598	472	597	647	796	720	737	689	[illegible]
14S/27E-17	334	557	0	1,158	896	622	462	833	1,075	1,180	944	843	677	[illegible]
14S/27E-18	92	129	360	482	484	457	497	687	786	876	714	653	540	[illegible]
14S/27E-20	0	86	429	436	1,111	1,608	2,295	2,950	3,229	3,553	2,850	2,556	2,062	[illegible]
14S/27E-28	0	0	0	0	0	0	0	0	0	0	0	0	64	[illegible]
14S/27E-29	151	155	319	901	830	720	723	781	756	1,057	1,067	1,202	1,223	[illegible]
14S/27E-30	144	202	366	274	290	286	0	346	332	341	250	198	132	[illegible]
14S/27E-32	108	301	510	1,044	860	653	559	632	635	712	585	539	451	[illegible]
14S/27E-33	72	140	162	234	152	73	13	114	192	285	300	350	365	[illegible]
15S/24E-13	0	0	0	0	0	0	0	0	1,430	1,024	628	728	421	[illegible]
15S/24E-25	0	0	0	0	0	0	0	0	0	0	0	0	177	[illegible]
15S/24E-27	0	0	0	0	650	800	1,000	1,230	460	336	212	252	152	[illegible]
15S/25E- 7	0	0	0	0	0	0	0	0	620	830	630	630	310	[illegible]
15S/25E-33	0	0	0	0	0	0	0	0	0	0	0	0	180	[illegible]
15S/26E-13	0	0	0	0	0	0	0	0	810	0	0	0	0	[illegible]
15S/26E-23	0	0	0	0	0	0	0	67	301	220	166	130	55	[illegible]
15S/26E-24	0	910	2,057	1,844	1,973	2,018	2,064	2,442	1,951	1,845	1,948	2,493	2,642	[illegible]
15S/26E-26	0	0	0	0	0	0	0	0	0	0	0	0	184	[illegible]
15S/26E-27	1,100	0	143	256	327	382	436	551	460	386	355	390	349	[illegible]
15S/26E-33	0	0	0	0	0	0	0	0	8	9	11	17	0	[illegible]
15S/27E- 6	0	0	148	142	683	1,149	1,976	1,395	866	888	984	1,524	1,764	[illegible]
15S/27E- 7	385	0	353	308	252	221	249	265	0	262	290	451	0	[illegible]
15S/27E- 8	0	0	805	2,037	1,311	812	388	1,304	1,785	1,293	973	939	551	[illegible]
15S/27E-18	0	803	1,435	2,352	1,974	1,787	2,072	1,975	1,747	1,603	1,613	2,297	2,473	[illegible]
15S/27E-19	599	493	663	837	574	339	315	388	409	406	438	665	756	[illegible]
15S/27E-20	0	433	490	425	399	401	0	483	422	338	294	355	316	[illegible]
15S/27E-29	116	71	306	299	307	331	0	440	401	290	218	209	120	[illegible]
16S/24E-12	200	220	235	275	300	350	400	490	1,230	1,030	720	1,330	1,417	[illegible]
16S/24E-23	0	0	0	0	0	0	0	0	0	0	0	0	523	[illegible]

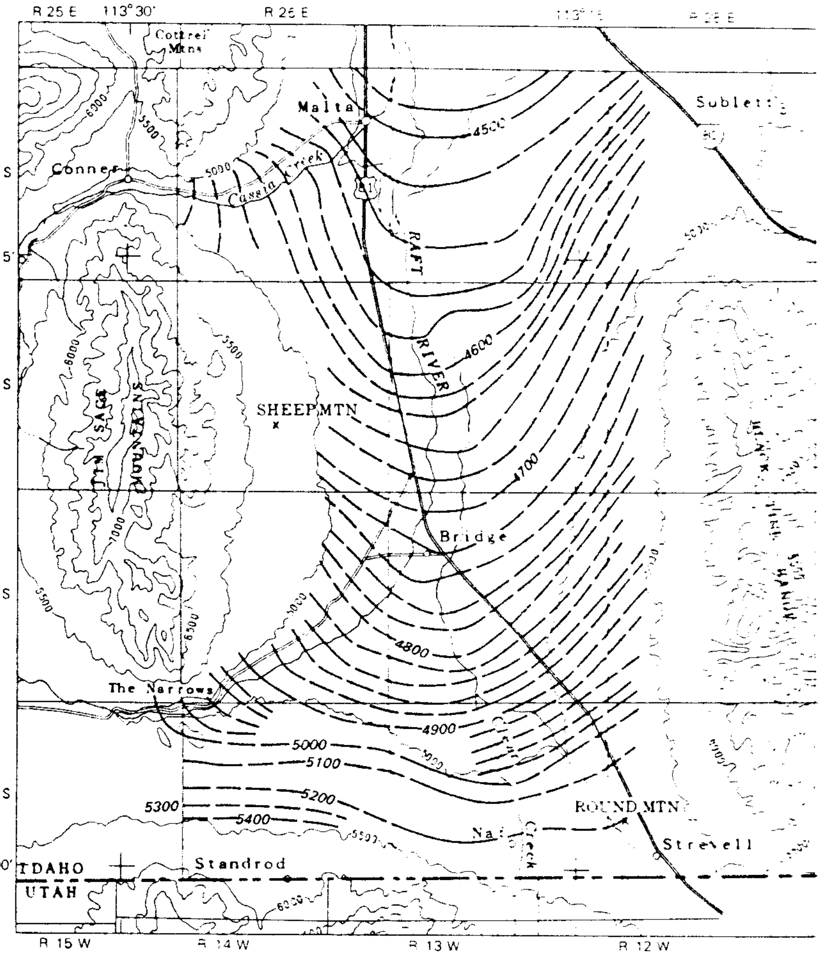
TABLE 2.—*Pumpage, in acre-feet,*

Location	1952	1953	1954	1955	1956
16S/25E- 5	0	0	0	0	0
16S/25E- 7	0	0	0	0	0
16S/25E- 8	0	0	0	0	0
16S/25E-11	70	70	70	70	400
16S/26E-11	0	0	0	0	370
16S/27E- 3	0	250	250	300	200
Rounded totals	10,200	11,600	24,100	32,200	35,100

TABLE 3.—*Estimated pumpage, in acre-feet, for 1966 through 1974*

[Totals rounded to three significant figures]

Location	1966	1967	1968	1969	1970	1971	1972	1973	1974
13S/25E-21	0	330	751	0	0	0	0	0	0
13S/25E-22	797	602	1,059	600	371	1	253	569	574
18S/25E-24	0	0	0	0	0	0	59	96	71
13S/25E-32	483	133	213	203	105	96	112	209	280
13S/26E- 1	1,555	1,034	507	609	397	491	477	329	529
13S/26E- 2	0	1,254	680	1,496	972	1,268	1,503	1,516	1,068
13S/26E-12	1,413	2,315	1,738	788	144	107	140	222	256
13S/26E-13	610	206	401	217	133	104	122	130	185
13S/26E-14	2,145	686	670	847	175	149	192	578	645
13S/26E-20	0	131	107	91	66	14	65	49	52
13S/26E-21	0	0	0	0	0	0	0	0	6
13S/26E-22	1,177	808	1,151	750	491	254	510	398	458
13S/26E-23	591	326	465	335	234	228	362	378	403
13S/26E-24	535	340	399	636	145	178	190	154	265
13S/26E-25	0	0	0	0	0	0	0	0	669
13S/26E-26	464	380	333	370	158	47	272	186	212
13S/27E- 2	2,103	1,817	1,964	2,090	2,090	1,806	1,805	1,229	735
13S/27E- 3	0	0	0	0	0	0	0	0	1,085
13S/27E- 4	0	0	0	0	0	0	0	0	282
13S/27E- 5	0	0	0	0	0	0	0	0	1,696
13S/27E- 6	1,019	556	726	560	417	211	364	452	535
13S/27E- 7	988	1,045	815	758	755	469	417	608	551
13S/27E- 8	525	317	314	253	405	583	728	509	552
13S/27E- 9	0	0	0	0	0	0	1,789	1,981	1,855
13S/27E-10	2,420	2,055	2,284	2,235	2,422	2,456	2,078	1,408	573
13S/27E-11	2,001	1,551	1,814	2,588	2,406	2,029	1,684	1,166	694
13S/27E-14	4,643	3,849	3,858	3,421	3,724	3,575	3,102	2,757	3,019
13S/27E-15	0	0	0	0	1,491	1,404	1,326	915	1,091
13S/27E-16	2,187	1,866	2,102	1,753	1,880	1,843	1,739	1,260	1,797
13S/27E-17	78	0	0	516	1,220	1,731	1,688	1,199	1,207
13S/27E-18	270	173	249	262	102	31	47	41	75
13S/27E-19	1,552	1,288	900	916	641	438	468	989	1,191
13S/27E-20	216	244	244	233	354	687	1,377	1,611	1,781
13S/27E-22	0	0	0	0	0	0	0	1,213	2,850
13S/27E-23	2,016	1,473	1,526	2,226	2,796	1,613	1,813	1,271	1,525
13S/27E-26	0	0	0	0	0	9	0	0	12
13S/27E-28	211	209	194	240	230	225	184	192	193
13S/27E-29	0	0	566	712	387	323	260	331	300
13S/27E-30	1,657	1,477	1,610	1,261	991	747	611	659	596
13S/27E-31	832	623	1,116	1,174	1,177	951	978	966	1,479
13S/27E-32	1,823	1,627	1,353	1,680	1,349	1,100	1,054	1,029	1,278
13S/27E-33	787	865	492	346	210	362	302	299	369
13S/27E-34	0	0	112	638	277	638	704	924	969
13S/28E- 3	2,143	2,129	2,084	1,877	1,727	1,643	1,249	1,461	1,305
13S/28E- 7	125	121	87	65	80	69	92	87	52
13S/28E-11	0	129	166	345	279	281	386	486	638
13S/28E-12	865	921	0	1,335	1,163	842	1,002	1,252	1,503
13S/28E-26	86	110	160	125	146	188	148	185	126
14S/27E- 4	602	430	1,164	1,058	949	682	825	889	1,234



Base modified from Army Map Service

0 5 10 15 KILOMETERS

0 5 MILES

CONTOUR INTERVAL 500 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929

EXPLANATION

—— 5000 ——
Line of equal water level altitude. Dashed where assumed
Datum is mean sea level
Interval 20 feet and 100 feet
(modified after Nace and others, 1961)

FIGURE 4—Altitude of water levels in the unconfined aquifer, 1952.

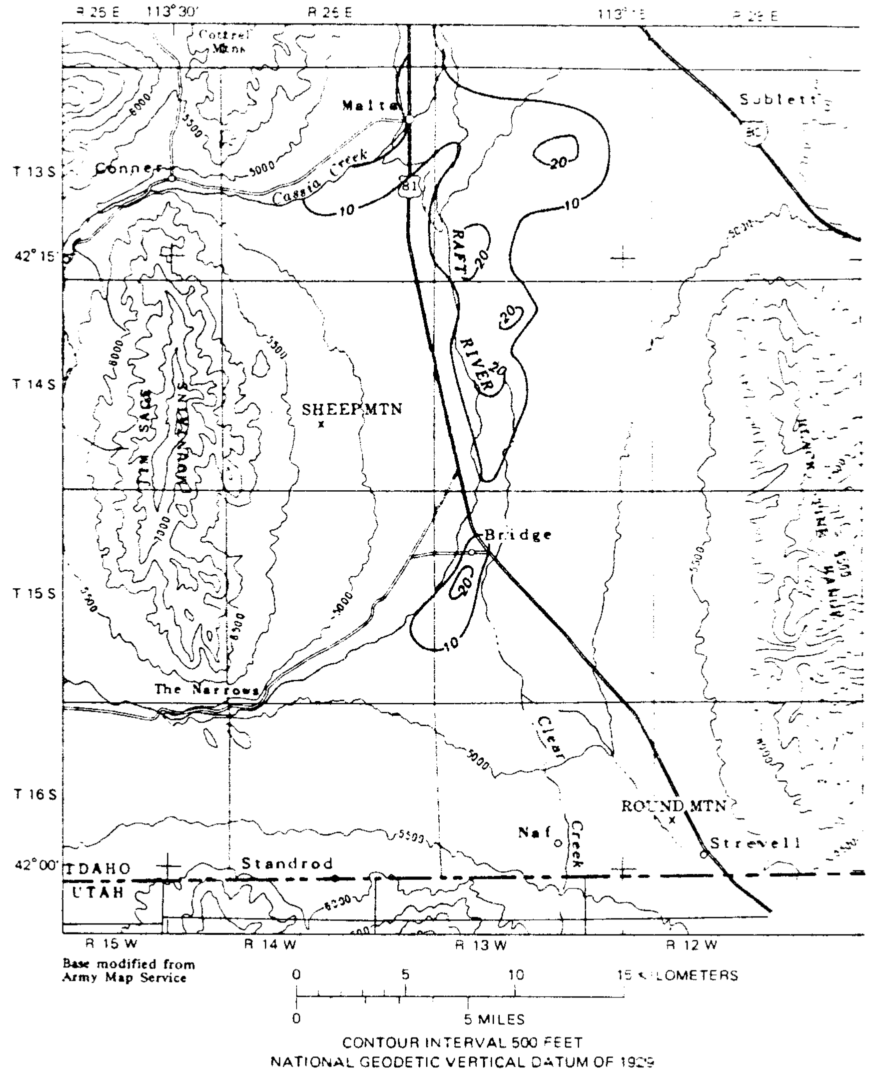
Almo, and Raft River Valley subbasins in 1966 (Walker and others, 1970, fig. 14). In April 1976, water levels were measured in the Yost-Almo and Raft River Valley subbasins (unpub. data, U.S. Geological

Survey, Boise, Idaho). Mundorff and Sisco (1963, pl. 1) showed water-level change only for the period 1952–61. During this time, the water level declined about 10 feet (3 m) near Bridge and about 15 feet (4.6 m) in an area midway between Bridge and Malta.

Walker and others (1970, fig. 14) extended the area in which water-level measurements were made and showed the shape and slope of the water table in the Yost-Almo and Elba subbasins for the first time. They reported (p. 60) that the slope of the water table was about 25 ft/mi (4.7 m/km) near Bridge and about 17 ft/mi (3.2 m/km) north of Malta, a probable slight increase in gradient since 1952. The change in water level from 1952 to the spring of 1966 (Walker and others, 1970, fig. 20), as summarized here in figure 5, shows that water levels generally continued to decline in the same areas as during the period 1952–61 except for the area east of Malta where new development seems to have occurred. Maximum declines are somewhat more than 20 feet (6 m). Available data suggest that water-level decline in the Raft River Valley subbasin continues for the most part in the relatively narrow belt of lowland along the Raft River.

Water-level measurements made in the spring of 1976 (unpub. data, U.S. Geological Survey, Boise, Idaho) indicate that the slope of the water surface is still 30 to 40 ft/mi (5.6 to 7.6 m/km) south of Bridge but that between Bridge and Malta it has increased to about 20 ft/mi (3.8 m/km). The pattern of water-level decline for the period 1952–76 has changed slightly as compared with that for 1952–66. The area south of Bridge over which 20 feet (6 m) or more of decline has occurred is somewhat smaller than in 1966, suggesting a small recovery of water level in this area, probably the result of a decrease in pumpage. Water-level decline in the lowland areas along Raft River between Bridge and Malta is significantly less for the period 1952–76 (Fig. 6) than for 1952–66 (Fig. 6). Water levels in the area east of Malta continued to decline to a maximum of nearly 40 feet (12 m). Annual water-level fluctuations have been discussed extensively by Nace and others (1961, p. 63–73) and by Walker and others (1970, p. 64–68).

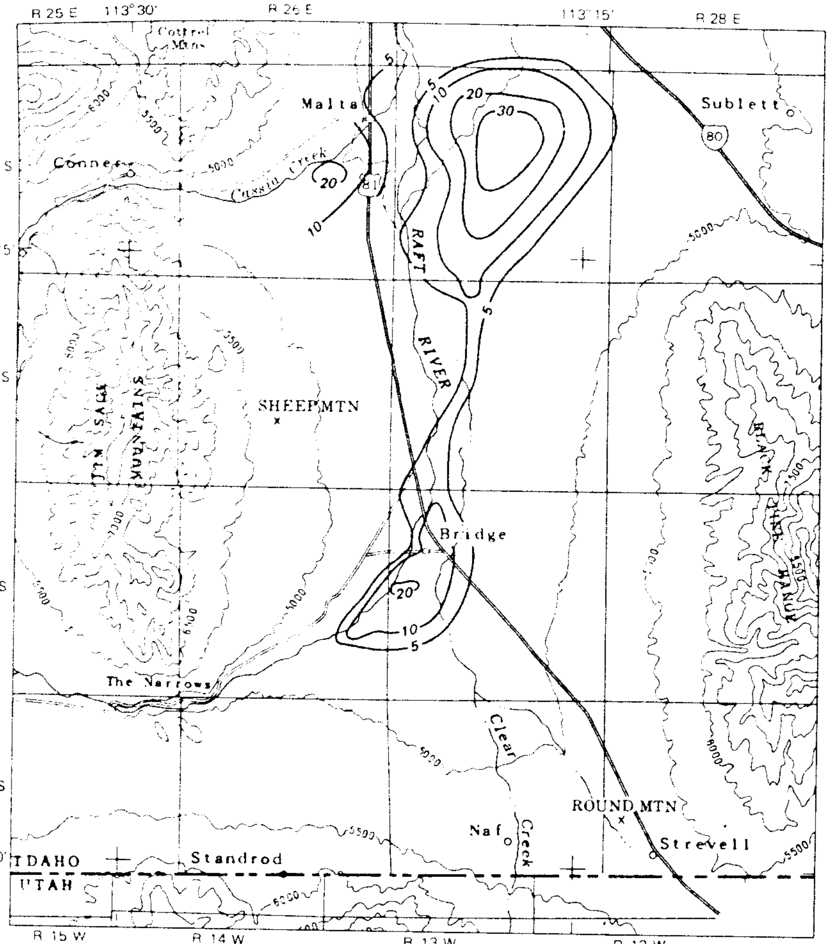
Enough water-level measurements were obtained by Walker and others (1970, fig. 14) to enable them to construct a map of the water table in the Yost-Almo and Elba subbasins for the first time. Ground-water development in both subbasins through 1965 was not large. Irrigation pumpage in the Yost-Almo subbasin was estimated to be about 3,900 acre-ft (4.8 hm^3) in 1965. Only about 480 acre-ft (0.59 hm^3) was pumped in 1965 for irrigation in the Elba subbasin. The water-level surface for spring, 1966, as shown by Walker and others (1970, fig. 14) probably approximates the predevelopment surface in these subbasins. The water level in the Yost-Almo subbasin was measured again in the spring of 1976. The net water-level change over the 10-year period has been slight; at some locations there has been a decline of 10 feet (3 m) or less, and at others there has been a rise of 10 feet (3 m) or less.



EXPLANATION

—— 10 ——
Line of equal water level decline
Interval 10 feet
(modified after Walker and others, 1970)

FIGURE 5.—Water-level change in the unconfined aquifer, 1952–65.



Base modified from Army Map Service

0 5 10 15 KILOMETERS

0 5 MILES

CONTOUR INTERVAL 500 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929

EXPLANATION

——— 20 ———
Line of equal water level decline
Interval 5 and 10 feet

FIGURE 6.—Water-level change in the unconfined aquifer, 1952–76.

PUMPAGE—QUANTITY AND TREND WITH TIME

Annual ground-water pumpage for irrigation in the southern Raft River Valley subbasin increased from an estimated 7,000 acre-ft (8.6

for 1952 through 1965—Continued

1957	1958	1959	1960	1961	1962	1963	1964	1965
0	0	0	0	0	0	0	116	259
0	0	0	178	656	656	841	690	476
0	0	0	0	0	0	0	496	461
600	650	800	302	904	754	819	574	414
370	370	450	40	0	0	0	0	0
200	160	200	210	360	60	80	0	0
35,300	38,800	47,000	55,300	59,100	55,900	58,200	57,900	55,200

TABLE 3.—*Estimated pumpage, in acre-feet, for 1966 through 1974*—Continued

Location	1966	1967	1968	1969	1970	1971	1972	1973	1974
14S/27E- 5	621	0	0	575	321	250	0	355	383
14S/27E- 6	3,719	2,585	2,160	2,994	3,000	962	1,337	1,429	2,222
14S/27E- 7	687	0	773	1,572	1,826	872	902	1,015	1,374
14S/27E- 8	1,290	3,317	4,025	2,958	2,834	1,611	3,038	2,712	3,532
14S/27E- 9	6,363	4,632	4,902	5,564	3,062	6,531	5,903	5,397	5,222
14S/27E-16	1,109	1,262	618	1,205	1,633	1,318	958	1,121	980
14S/27E-17	1,548	848	531	873	556	122	850	790	370
14S/27E-18	1,133	0	795	2,930	2,474	588	442	1,389	1,481
14S/27E-19	1,241	1,006	8,117	1,127	508	165	542	669	678
14S/27E-20	737	1,714	1,775	2,095	2,310	1,680	1,285	1,189	[illegible]
14S/27E-21	458	361	457	362	323	456	819	559	672
14S/27E-28	498	444	658	360	89	78	104	230	365
14S/27E-29	595	1,271	1,333	1,236	956	288	637	731	574
14S/27E-31	340	272	0	479	230	210	0	406	649
14S/27E-32	0	308	0	305	284	55	238	139	159
14S/27E-33	246	0	180	0	149	0	188	275	[illegible]
15S/24E-13	793	561	435	255	198	140	178	192	252
15S/24E-22	0	13	79	9	10	18	22	14	22
15S/24E-25	390	277	256	168	85	55	127	197	254
15S/24E-26	218	378	375	271	197	122	245	246	196
15S/24E-27	0	0	0	0	0	0	0	39	144
15S/25E- 6	407	215	374	148	0	0	35	132	301
15S/25E- 7	294	254	276	194	127	96	75	206	209
15S/26E-23	0	277	517	412	307	191	504	531	699
15S/26E-24	3,892	5,005	4,066	3,991	3,068	2,289	1,814	2,897	4,503
15S/26E-26	0	486	547	0	0	0	810	792	794
15S/26E-27	1,541	1,452	933	1,323	387	1,247	1,649	1,584	1,429
15S/27E- 5	1,612	1,844	1,810	1,668	919	2,138	2,592	2,144	1,794
15S/27E- 6	1,008	1,324	1,550	1,203	1,188	967	921	972	1,012
15S/27E- 7	755	373	167	500	593	376	314	479	312
15S/27E- 8	2,057	305	0	1,051	523	1,693	1,978	1,486	2,344
15S/27E-17	786	0	0	0	117	82	472	572	393
15S/27E-18	3,796	4,890	4,045	2,951	2,819	1,950	2,153	2,238	2,226
15S/27E-19	2,550	1,946	2,421	2,173	1,715	1,061	1,089	1,644	2,094
15S/27E-20	0	0	0	0	0	0	0	0	569
15S/27E-29	2,117	2,353	1,376	1,551	1,234	1,204	1,417	1,211	1,544
16S/24E-12	599	809	657	427	359	339	296	435	579
16S/24E-13	0	0	70	124	75	59	48	55	[illegible]
16S/24E-14	0	0	247	172	606	0	252	146	164
16S/24E-23	512	297	354	360	477	319	458	401	510
16S/25E- 4	219	157	115	161	94	51	88	91	176
16S/25E- 7	274	265	235	248	237	241	315	272	505
16S/25E- 8	427	344	349	372	380	276	425	495	567
16S/25E-11	367	343	268	411	158	115	203	227	293
16S/25E-31	2	8	17	16	12	6	14	13	[illegible]
16S/26E-20	0	0	0	0	45	54	0	35	47
16S/27E- 3	22	0	33	0	0	0	27	0	25
16S/27E- 4	0	0	0	0	0	0	0	0	381
16S/28E-29	0	0	0	0	0	0	0	0	10
Rounded totals	84,100	75,600	75,000	80,400	69,500	60,500	68,200	70,500	[illegible]

evapotranspiration represents an increase of only about 4 percent over previous estimates (Nace and others, 1961, table 5; Walker and others, 1970, p. 75) for the entire basin.

SIMULATION OF THE SHALLOW GROUND-WATER SYSTEM

Detailed discussions of simulation model theory and the theoretical basis of model development have been given by Pinder and Bredehoeft (1968) and Trescott and others (1976). The method involves solving finite-difference approximations of the partial differential equation of two-dimensional ground-water flow (Trescott and others, 1976). This technique represents the aquifer as a two-dimensional grid or network of rectangular elements; at the center of each element is a node where aquifer data are given. The grid for the southern Raft River subbasin model was designed to be coincident with the township, range, and section grid, and each element covers 1 mi^2 (2.6 km^2).

Any model is, at best, an approximation of the real hydrologic system. All the complexities of the actual system cannot be included. Simplifying assumptions are required to make the problem manageable. The present analysis is based on the following assumptions:

1. All flow in the aquifer is unconfined and two-dimensional with no vertical component of flow. Flow across the boundaries is perpendicular to the boundary.
2. The aquifer is homogeneous within a given element of the finite-difference grid.
3. The Raft River is not a significant hydrologic boundary.
4. 1952 water levels represent steady-state water levels.
5. Water levels along the margins of the basin are maintained by recharge from the immediately adjacent mountain ranges.
6. Previous estimates of the magnitude and distribution of available recharge are reasonable.
7. The pumping rate, as averaged over the period of calibration, adequately represents the stress on the aquifer; net pumpage is 60 percent of the total withdrawal and irrigation return flow equals 40 percent of the withdrawal. This is based on the suggestion by Walker and others (1970) that about 60 percent of irrigation pumpage is consumed.

DISCUSSION OF ASSUMPTIONS MADE IN THE MODEL

BOUNDARY CONDITIONS

The locations of the model boundaries are shown in figure 2. All boundaries were arbitrarily located because their nature is unknown, but they generally coincide with the aquifer limits suggested by Walker and others (1970). It has been assumed that water-table conditions extend up to and beyond the boundaries shown, especially those on the east, south, and west. The northern border is represented in the model as a discharge boundary; all others are represented as recharge boundaries. These representations are based on the configuration of the 1952 water table.

INITIAL CONDITIONS

INITIAL HEAD

The water-level data given by Nace and others (1961, plate 5) were used as the basis for estimated initial head conditions in the lowland part of the southern Raft River Valley subbasin before development. These data were supplemented by extrapolating water-level data from measurements in 1965 and 1976 back to 1952 and by estimating water-level values beyond the valley bottom so that the entire modeled area was covered (fig. 4). Only in this way could the boundaries be moved far enough away from the areas of major pumping and anticipated future development so as to be unaffected by head changes caused by current and future simulated pumpage. The head condition shown in figure 4 is assumed to represent steady-state conditions because it is in general agreement with scattered measurements made intermittently since 1928.

STEADY-STATE FLUXES

The estimated volume and distribution of predevelopment recharge, or water yield (Nace and others, 1961; Walker and others, 1970) available to the ground-water system, in the southern Raft River Valley subbasin have been the principal constraints in determining steady-state recharge volumes and distribution in the simulation model. The amount of recharge estimated for the Yost-Almo subbasin, Elba subbasin, Raft River Mountains, Jim Sage Mountains, Black Pine Mountains, and Meadow Creek subbasin is assumed to represent the long-term average annual recharge. Implicit in this assumption is the further assumption that there is no recharge by underflow from adjacent basins. This is believed to be reasonable in view of the geology of the Raft River Basin. The entire basin is surrounded by either fault-block mountain ranges of Paleozoic rocks or by mountain ranges consisting of gneiss-dome complexes of Precambrian quartz monzonite mantled by Precambrian and lower Paleozoic metasedimentary rocks. Significant interbasin underflow is unlikely under these circumstances.

The distribution of flux crossing the eastern boundary of the model is assumed to be determined by the general distribution of estimated recharge available as underflow from the Meadow Creek subarea and the Black Pine Mountains; the total volume of flux cannot exceed the estimated total recharge available from these two areas. Initial fluxes crossing the southern boundary of the model are assumed to be equal to or less than the volume estimated to be available as underflow from the Raft River Mountains subarea. Along the western boundary, from about 1.5 miles (2.4 km) south of Cassia Creek to about 0.5 mile (0.8 km) north of the Raft River at The Narrows, the boundary flux is assumed to be supplied by infiltration and deep percolation on the eastern slope of the Jim Sage Mountains. The small segment of the boundary north of the Jim Sage Mountains is assumed to receive some recharge by shallow underflow through the valley of Cassia Creek from the Elba subbasin. South of the Jim Sage Mountains the boundary flux is assumed to come from the Yost-Almo subbasin, partly by shallow underflow through The Narrows, and partly by underflow along the entire length of the boundary south of The Narrows. The flux across the northern boundary of the modeled area is ground-water outflow from the southern Raft River Valley subbasin; volume estimates have not been made previously for this location. All boundary fluxes in the model are assumed to represent the "horizontal" movement of water in the aquifer across the vertical plane of the arbitrary boundaries.

Additional steady-state recharge and discharge fluxes occur throughout the area encompassed by the model boundaries. The distribution, magnitude, and character of these fluxes are determined by the transmissivity and relative head distribution in the modeled area. The ultimate constraint on the total magnitude of recharge is the estimated recharge still available from intrabasin sources, in other words, the amount of estimated total available recharge that has not been committed to satisfy initial head conditions on the boundaries. Steady-state fluxes needed to maintain steady-state head distribution over the modeled area within the limits of the boundaries represent the net vertical movement of water into or out of the aquifer. The total net discharge, including discharge along the north boundary, must equal the net recharge including all recharge along the boundaries on the east, south, and west.

MODEL DEVELOPMENT AND CALIBRATION

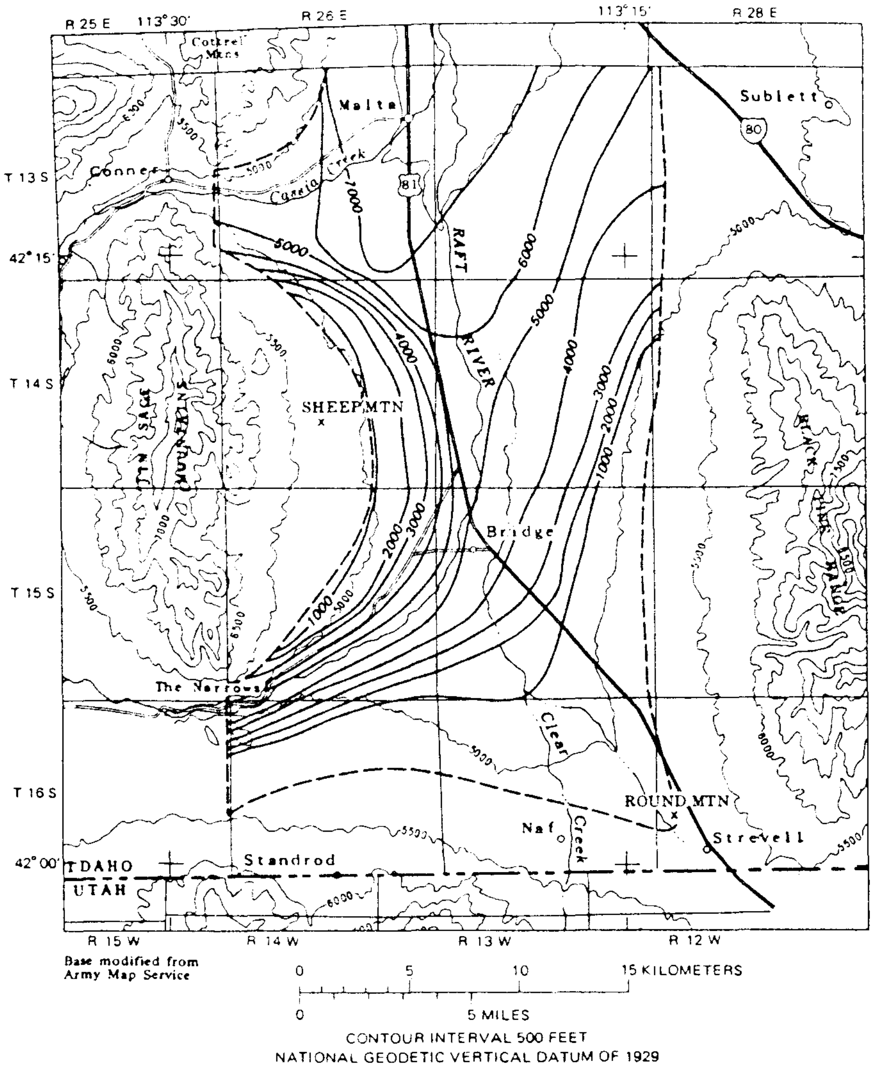
Model development and calibration were carried out in two stages. The first stage was the simulation and preliminary calibration of steady-state conditions in the shallow aquifer. This stage was used to test the conceptual model of the system and to evaluate the hydrodynamics of the aquifer under steady-state conditions. Calibration of

supplied by inflow across lateral boundaries and by water from "deep" infiltration originating high in the Raft River Mountains. Both forms of recharge are believed to occur at relatively shallow depths in this area because close correlation has been observed between high and low streamflows in Clear Creek, near Naf, and high and low ground-water levels, respectively, in an observation well near Naf. The relationship between stream discharge and ground-water level was pointed out by Nace and others (1961, p. 66, fig. 17) and by Walker and others (1970, p. 66, fig.16). The available data and calculated values are consistent with the interpretation that recharge to the north-sloping pediment from the Raft River Mountains is mostly by shallow underflow and that the water is almost entirely consumed through evapotranspiration by phreatophytes growing on the pediment surface. Little of the recharge from the Raft River Mountains subarea reaches the unconfined ground-water system north of the pediment.

TRANSMISSIVITY

The transmissivity distribution used in the final version of the steady-state model is shown in figure 8. Values computed by the model range from less than 1,000 ft²/d (90 m²/d) to about 7,000 ft²/d (650 m²/d). These values are about one order of magnitude smaller than those computed from aquifer-test data by Nace and others (1961) and by Morrilla and Ralston (1976). They are consistent with values of transmissivity estimated from specific-capacity data during this study. Larger values of transmissivity were used during the early stages of steady-state model development. These initial values ranged from about 2,500 ft²/d (230 m²/d) to 17,400 ft²/d (1,620 m²/d) but required a recharge rate of about 117,000 acre-ft/yr (144 hm³/yr) to maintain the 1952 head distribution. This rate is about 125 percent of the maximum estimated available recharge. Downward revision of transmissivity by 30 percent lowered the recharge demand to about 94,000 acre-ft/yr (116 hm³/yr), only 1 percent larger than the maximum estimated. Additional analysis during calibration of the nonsteady-state model indicated that transmissivity values were still too large because the cones of depression caused by simulated pumping were much too flat, or widespread, as compared with field data. Transmissivity was then reduced by another 35 percent, through trial and error, so that observed decline gradients could be reproduced.

Transmissivity distribution was determined early in the development of the steady-state model on the basis of the availability of, and demand for, recharge. This distribution remained unchanged during subsequent reduction of transmissivity.



EXPLANATION

——5000—— Line of equal transmissivity in ft²/d. Interval 1000 ft²/d

FIGURE 8.—Transmissivity, in ft²/d, of the unconfined aquifer in 1952.

CREDIBILITY OF RESULTS

The general validity of the computed transmissivity and recharge-discharge values is demonstrated by the mutually limiting constraints imposed on the problem. Water levels measured by Nace and

others (1961) in the central part of the valley in 1952 were considered unalterable and were assumed to represent long-term average steady-state conditions. Estimated water levels along the valley margins were modified slightly on the basis of measurements made in 1965 (Walker and others, 1970) and data collected in 1975 and 1976. Once determined, water level was a rigid constraint on further analysis.

Given the 1952 head distribution, the problem of steady-state analysis is reduced to determining mutually dependent values of flux and transmissivity. Furthermore, given limiting values and an approximate distribution of flux, the problem becomes almost trivial because under steady-state conditions, there is no change in storage in the aquifer and the transmissivity multiplied by the hydraulic gradient minus flux must equal zero. It follows that if recharge-discharge rates and distribution computed by the model are reasonable, then transmissivity values used with those fluxes to maintain the given hydraulic heads and gradients must also be reasonable.

Estimates of available recharge, determined during two previous studies (Nace and others, 1961; Walker and others, 1970), were used as the limiting minimum and maximum constraints for steady-state recharge in the simulation analysis. If the limiting estimates of total available recharge are acceptable, then the computed value, which is within the limits of the two extremes, must be considered reasonable. The distribution and magnitude of available recharge along the model boundaries appear credible on the basis of earlier estimates (table 4 and fig. 7). The greatest discrepancy between estimated and model-computed boundary inflow is at The Narrows. Recent test drilling in The Narrows has suggested that the cross-sectional area of the channel fill in The Narrows is as small as 80,000 ft² (7,000 m²)—one-sixth of the previously estimated 500,000 ft² (50,000 m²). Using the reduced cross-sectional area, a gradient of 40 ft/mi (7.6 m/km), and a rather large hydraulic conductivity of 133 ft/d (40 m/d), only about 680 acre-ft/yr (0.84 hm³/yr) is estimated to move through The Narrows as underflow. The computed boundary flux is about 400 acre-ft/yr (0.5 hm³/yr), close to the revised estimated flux based on field data.

The present interpretation of recharge distribution within the boundaries of the model differs substantially from those of previous investigators, who assumed that major recharge to the southern Raft River Valley subbasin is by shallow underflow through The Narrows. The present conceptual model of hydrodynamics in the subbasin assumes that the principal recharge is by upward leakage from deeper confined aquifers. This concept has been used successfully in mathematically modeling the system as described in the following sections. On the basis of what is now known of the system, the distribution of recharge and discharge developed during the present study is believed to be logical, reasonable, and hydrologically sound.

NONSTEADY-STATE ANALYSIS

Nonsteady-state, or transient, calibration was carried out to refine initial estimates of transmissivity, recharge-discharge rates, specific yield, net pumping rates, and evapotranspiration capture until the simulation model reproduced observed historic water-level changes (fig. 5) as given by Walker and others (1970, fig. 20). The period 1952–65 was selected as the initial calibration period because it is the only one for which all the necessary data are available.

The initial transmissivity and flux distribution selected are the same as those computed earlier by the preliminary steady-state model; initial water levels for the transient model are the 1952 heads. The distribution of specific yield (fig. 9) is the same as that estimated by Walker and others (1970, fig. 19). Pumping rates for each township in the modeled area have been given by Walker and others (1970, table 11) for 1948 to 1966; pumpage for each section within a township was determined proportionally from the electrical power used for pumping in the section and the total estimated pumpage in the township as given by Walker and others (1970). A net pumping rate equal to 60 percent of the 14-year average pumping rate was used to simulate the average stress on the aquifer from 1952 through 1965. At the beginning of the calibration process, no capture of evapotranspiration was assumed; this parameter was added later. Observed water-level decline over the period 1952–1965 was reproduced by trial-and-error changes in all the above parameters except the initial head.

INITIAL RESULTS AND PARAMETER MODIFICATION

The preliminary versions of the transient model,using the initial high estimated transmissivity values ranging from 1,700 ft²/d (160 m²/d) to 12,000 ft²d (1,100 m²/d), computed declines of 10 to 15 feet (3 to 4 m) over the entire modeled area; there was little definition of any of the observed pumping depressions. Changes made in specific yield, average pumping rates, and captured discharge did not improve the solution. The only way to reduce declines outside pumping areas and to steepen the observed decline gradient was to reduce the magnitude of tramsmissivity. There was a corresponding reduction in predevelopment recharge-discharge rates. The best solution (fig. 10) was obtained after reducing the preliminary transmissivity estimates by about 35 percent. The resulting distribution is that shown in figure 8. The final inferred natural recharge-discharge distribution is that shown in figure 7. Average pumping rates were revised somewhat

this model consisted of determining, within given constraints and by trail and error, reasonable values of transmissivities and recharge-discharge fluxes. The second stage of development and calibration analyzed the response of the simulated aquifer to historic pumping stress for the period 1952–65 with the objective of reproducing observed historic water-level changes during the same time period. Calibration was carried out by making changes or adjustments (mostly small) in transmissivity, recharge-discharge fluxes, specific yield, and average pumping rates for 1952–65.

STEADY-STATE ANALYSIS

Steady-state analysis was done to determine reasonable values of transmissivity and boundary fluxes and to determine the location and volume of recharge and discharge within the model boundaries. Water levels in 1952 (fig. 4) were assumed to represent long-term average steady-state conditions. Total available long-term average net recharge was assumed to be somewhere between the minimum and maximum estimates of Walker and others (1970) and Nace and others (1961). Their estimates range from about 46,000 acre-ft/yr (57 hm³/yr) to about 100,000 acre-ft/yr (123 hm³/yr). Table 4 shows the general areal distribution of these estimates. Limited by the constraints of fixed water levels and a range for total recharge, the steady-state analysis was used to determine a plausible distribution of recharge and discharge fluxes inside the model boundaries and a compatible distribution and magnitude of transmissivity.

TABLE 4.—*Estimated maximum and minimum water available for ground-water recharge*

Area	Estimated ground-water recharge (in acre-feet/year)	
	Nace and others	Walker and others
Elba subbasin	[1]1,900	[2]600
Yost-Almo subbasin	[3]60,000	[4]16,900
Southern Raft River Valley subbasin	[5]30,000(?)	20,900
Jim Sage Mountains		[6]5,300
Black Pine Mountains		[6]6,200
Raft River Mountains		[7]9,400
Meadow Creek subbasin	[8]8,200	[9]7,700
Total	100,100	46,100

[1]Nace and others, 1961, p. 31, 47, and 49.
[2]Walker and others, 1970, p. 61.
[3]Nace and others, 1961, p. 31, 47-48.
[4]Walker and others, 1970, p. 55.
[5]Estimated for this study.
[6]Estimated for this study from Walker and others, 1970, figure 13.
[7]Walker and others, 1970, p. 46 and 55.
[8]Nace and others, 1961, p. 31.
[9]Walker and others, 1970, p. 46.

RECHARGE AND DISCHARGE

The long-term average steady-state net flux distribution computed by simulation modeling is shown in figure 7. These values were computed empirically as a function of 1952 head and transmissivity in the unconfined aquifer. The total net computed recharge is 63,300 acre-ft/yr (78.1 hm^3/yr). This is about 37 percent larger than the minimum estimate of net recharge made by Walker and others (1970) and 36 percent smaller than the maximum estimated net recharge of Nace and others (1961). Recharge along those boundaries of the model corresponding to areas bordered by mountain ranges is assumed to be derived from shallow underflow from ground-water sources in the mountains. The magnitude and distribution of recharge along these (the mountainous) parts of the model boundaries compare favorably with estimated available recharge from the respective areas. Recharge along about 6 miles (10 km) of the northeast border of the model is assumed to be underflow from the Meadow Creek subbasin. Small amounts of recharge cross the boundary as shallow underflow from the Elba subbasin and the Yost-Almo subbasin. Table 5 gives the source and magnitude of boundary recharge determined by steady-state simulation analysis. These data can be compared with estimates given in table 4.

Computed areal recharge fluxes within the boundaries of the model (fig. 7) are assumed to represent vertical leakage upward through low-permeability confining beds underlying the shallow aquifer, although the confining beds and underlying source are not simulated explicitly in the model. The immediate source of recharge is believed to be deeper confined aquifers. The ultimate source of recharge to the confined aquifers is probably deep infiltration in the Yost-Almo subbasin, with lateral flow beneath the Jim Sage Mountains, but part of it may be from deep infiltration of water in the surrounding mountain ranges.

TABLE 5.—*Steady-state boundary recharge rates computed by simulation model*

Source area	Rate	
	Acre-feet per year	Cubic hectometers per year
North of Cassia Creek	2,000	2.47
Elba subbasin:		
Cassia Creek Valley	1,330	1.64
Yost-Almo subbasin:		
The Narrows	400	.49
South of The Narrows	1,050	1.30
Raft River Valley subbasin:		
Jim Sage Mountains	5,050	6.23
Raft River Mountains	4,950	6.11
Black Pine Mountains	3,600	4.44
Meadow Creek subbasin	2,020	2.49

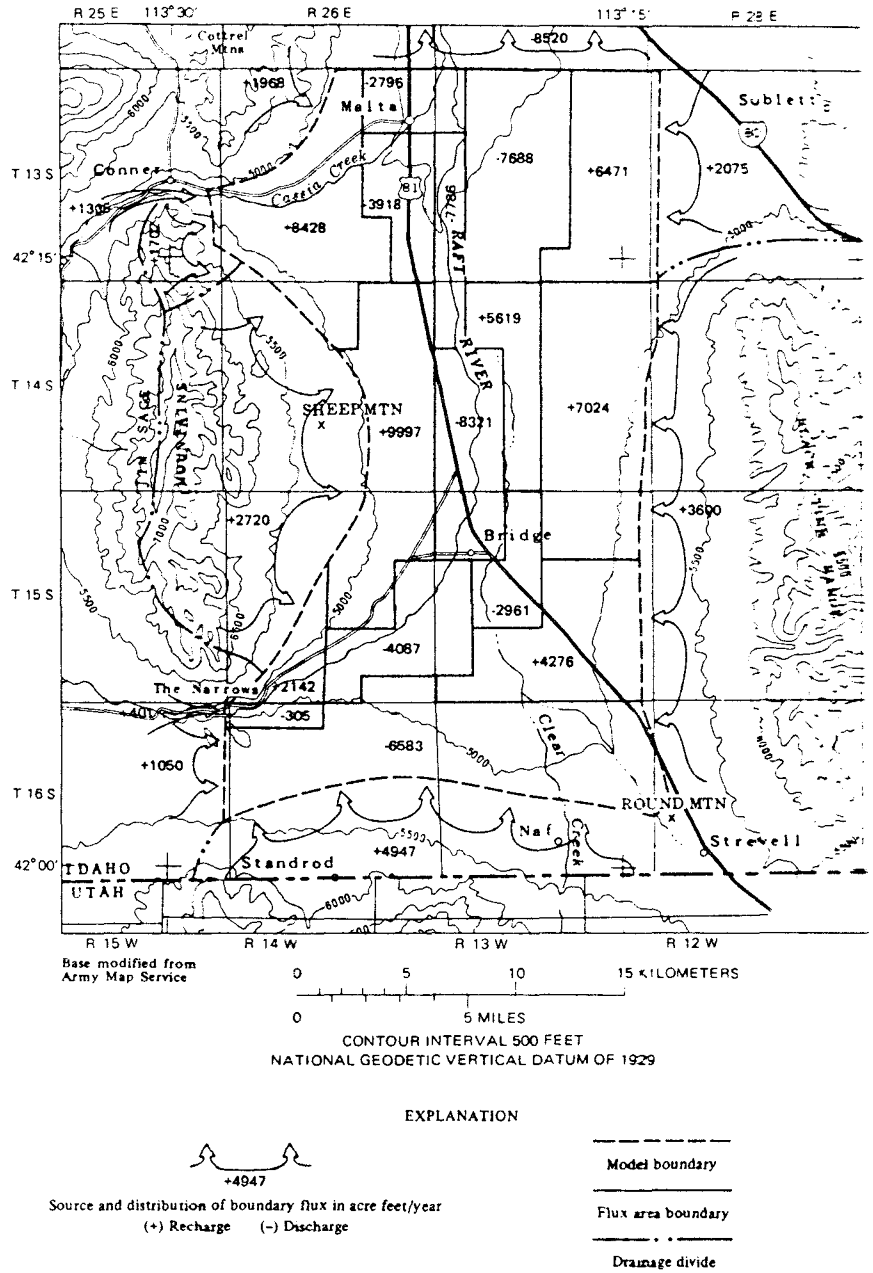
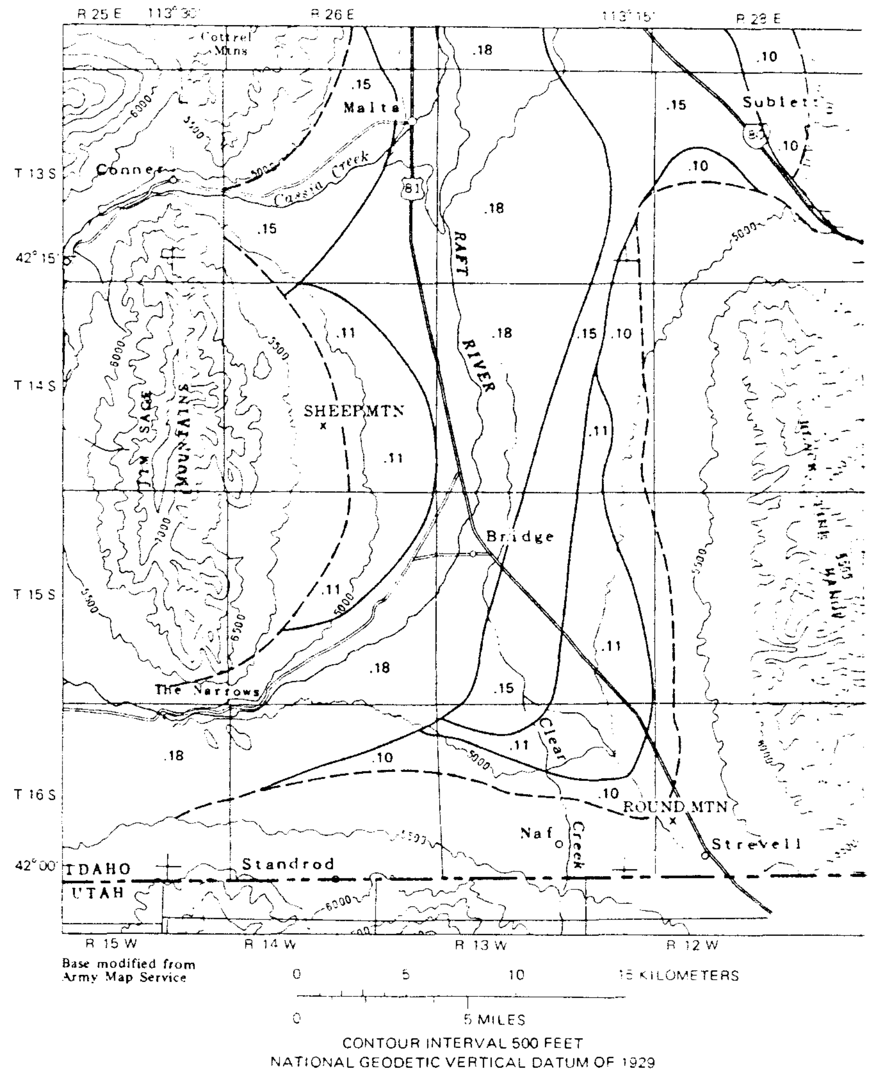


FIGURE 7.—Distribution of long-term average steady-state recharge and discharge, in acre-feet/year, based on 1952 water levels.

The maximum estimated net "vertical" recharge rate used in the simulation model is about 6.7 x 10^{-3} ft/d (2.0 x 10^{-3} m/d). Data collected during test drilling in 1974 and 1975 in an area about 5 miles (8 km) northeast of The Narrows suggest that a confining layer about 600 feet (180 m) thick underlies the shallow aquifer. The head in the aquifer beneath the confining layer may be as much as 50 feet (15 m) higher than the head in the overlying unconfined aquifer. The maximum computed vertical flux rate could be realized under these conditions of confining-layer thickness and head difference with an average vertical hydraulic conductivity of the confining bed of about 8.0 x 10^{-2} ft/d (2.4 x 10^{-2}m/d). This value compares favorably with hydraulic-conductivity values ranging from 5.3 x 10^{-6} ft/d (1.6 x 10^{-6} m/d) to 8.5 x 10^{-2} ft/d (2.6 x 10^{-2} m/d) determined in the laboratory for rock types similar to the siltstone, sandstone, and conglomerate that form the confining bed (Aerojet Nuclear Co., 1976, written commun). The core samples used for laboratory analysis were obtained from various depths in one of the geothermal production wells.

Calculated areal discharge of ground water is confined largely to the valley bottom (fig. 7), except in the southern part of the basin where the presence of phreatophytes shows that the entire pediment surface extending northward from the Raft River Mountains is also a discharge area. The discharge rates computed by the model range from 3.8 in/yr (97 mm/yr) to 24 in/yr (610 mm/yr). The maximum computed discharge rate occurs along a 6-mi (10-km) reach of the Raft River immediately upstream from Malta. Nace and others (1961, p. 52) indicated this was gaining reach of the stream. Streamflow measurements made at 11 sites along the river between the Narrows and Malta in 1949 and 1950 indicated a net increase in discharge, most of which occurred along the 8- to 10-mile (13- to 16-km) reach above (south of) the junction with Cassia Creek (Nace and others, 1961, p. 52). The data are for such a short period of time and are so variable that conclusions about long-term average changes in streamflow are tenuous. Nevertheless, an increase in streamflow from ground-water sources of only 5 ft^3/sec (0.14 m^3/s), which is within the limits of short-term observations made by Nace and others (1961), reduces ground-water discharge through other means to 13.2 in/yr (335 mm/yr) in this part of the valley. Empirically computed discharge rates for other areas of the valley bottom range from 8.4 in/yr (213 mm/yr) to 14.4 in/yr (366 mm/yr). All these rates are within the expected evapotranspiration rates for the phreatophytes growing in the area.

Discharge from the pediment extending northward from the Raft River Mountains in the southern end of the basin is believed to be



EXPLANATION

Aquifer boundary of Walker and others, 1970 (after Walker and others, 1970)

FIGURE 9.—Specific yield of the unconfined aquifer.

from the preliminary estimates. The original and final average rates used in the model are given in table 6.

The distribution of specific yield was changed uniformly and equally over the area by a factor ranging from 0.8 to 1.3 times the

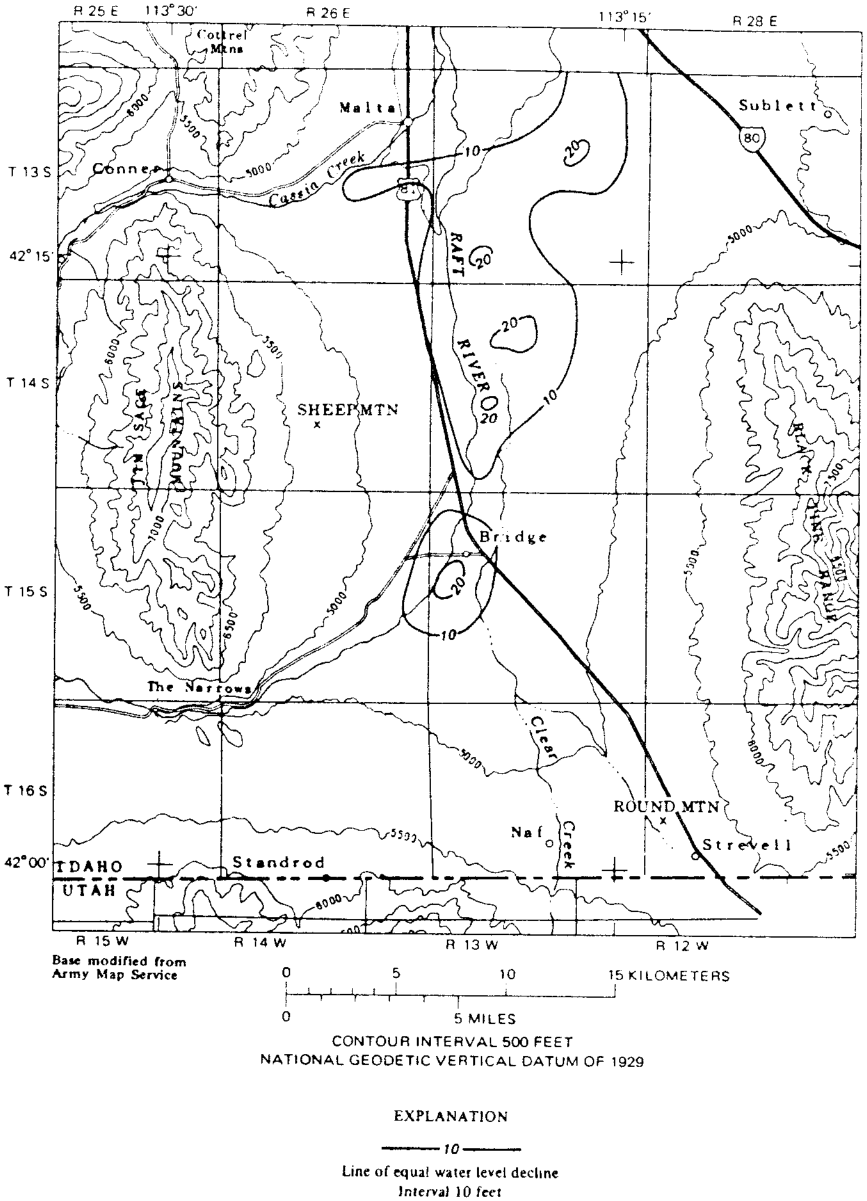


FIGURE 10.—Computed water-level change in the unconfined aquifer, 1952–65.

original distribution shown in figure 9. The original values give the best solution, and the calibrated model used the original estimates of specific yield (fig. 9).

TABLE 6.—*Average annual pumping rates for 1952 through 1965 as used in the calibrated simulation model based on previously published estimates*

Township Range	Section	From estimates by Walker and others (1970): Acre-feet per year	From estimates by Walker and others (1970): Cubic hectometers per year	Used in model: Acre-feet per year	Used in model: Cubic hectometers per year	Percent change
13S26E	1	210	0.259	210	0.259	----
	12	137	.169	137	.169	----
	13	970	1.196	970	1.196	----
	14	50	.062	50	.062	----
	20	7	.009	7	.009	----
	22	528	.651	*695	.857	32
	23	260	.321	*471	.581	81
	24	65	.080	65	.080	----
	26	123	.152	123	.152	----
Sum		2,350	2.899	2,728	3.365	16
13S27E	2	1,600	1.974	1,600	1.974	----
	6	789	.973	789	.973	----
	7	673	.830	673	.830	----
	9	152	.187	152	.187	----
	10	507	.625	507	.625	----
	11	702	.866	702	.866	----
	14	1,484	1.830	*760	.937	−49
	15	0	0	*724	.893	----
	16	1,086	1.340	1,086	1.340	----
	17	80	.099	80	.099	----
	18	311	.383	311	.384	----
	19	253	.312	253	.312	----
	20	340	.419	340	.419	----
	23	297	.366	297	.366	----
	29	355	.438	355	.438	----
	30	1,231	1.518	*869	1.072	−29
	31	369	.455	369	.455	----
	32	1,614	1.991	*1,882	2.321	17
	33	579	.714	579	.714	----
Sum		12,420	15.32	12,330	15.21	−0.7
14S27E	4	1,955	2.411	*1,086	1.340	−44
	5	369	.455	369	.455	----
	6	1,202	1.483	*833	1.027	−31
	7	1,064	1.312	615	.759	−42
	8	1,890	2.331	*1,013	1.250	−46
	9	3,178	3.920	2,534	3.126	−20
	16	615	.759	615	.759	----
	17	724	.893	*615	.75	−15
	18	514	.634	514	.634	----
	20	1,781	2.197	*2,063	2.545	16
	28	7	.009	7	.009	----
	29	782	.965	782	.965	----
	30	233	.287	233	.287	----
	32	565	.697	565	.697	----
	33	188	.232	188	.232	----
Sum		15,070	18.59	12,030	14.84	−20
15S26E	13	58	0.072	58	0.072	----
	23	72	.089	72	.089	----
	24	1,933	2.383	*977	1.205	−49
	26	36	.044	36	.044	----
	27	456	.562	456	.562	----
	32	7	.009	7	.009	----
Sum		2,562	3.159	1,606	1.981	−37
15S27E	6	992	1.223	*434	0.535	−56
	7	261	.322	261	322	----
	8	977	1.205	977	1.205	----
	18	1,745	2.152	*1,889	2.339	8
	19	543	.670	*1,339	1.651	147
	20	333	.411	333	.441	----
	29	232	.286	232	.286	----
Sum		5,083	6.269	5,465	6.739	8
Total		37,480	46.21	34,200	42.17	−9

*Value changed from published estimate.

CAPTURE OF NATURAL DISCHARGE

Ground water is discharged by evapotranspiration over most of the valley bottom and by seepage to the Raft River along some reaches of that stream. This discharge is included in the transient model as the negative component of predevelopment flux (fig. 7) computed during steady-state analysis. The areas of discharge generally coincide with areas of shallow water table (2 to 15 ft below land surface, or 0.6 to 4.6 m) and high density of phreatophyte growth. As the ground-water level is lowered by pumping, discharge to the river is reduced and it may even cease if the decline is large enough. More significantly, evapotranspiration losses decrease as the water level declines beneath areas of phreatic vegetation. Provision is made in the transient model for the capture of natural discharge as a function of computed water-level decline and depth to the water table. Where no decline is computed, it is assumed there is no change in discharge, and no capture occurs.

A water-table depth of 30 feet (9 m) was selected as that at which maximum capture is realized. The principal phreatophytes in the valley are greasewood, whose roots may extend more than 30 feet (9 m) below land surface, and rabbitbrush, whose root depth is as deep as 20 feet (6 m) (Robinson, 1958). In the model, discharge is captured as the ground-water level declines. The rate of capture increases in linear fashion as the water level approaches 30 feet (9 m); at that level the maximum rate of capture is reached, and this rate remains constant thereafter. The maximum rate of capture, estimated to be about 400 (acre-ft/yr)/mi^2 (1.3 hm^3/yr/km^2), was determined by trial and error; it is well within the limits of evapotranspiration expected for the types of phreatophytes growing in the valley bottom lands.

CALIBRATED SOLUTION, 1952–65

The calibrated solution for the period 1952–65 is shown in figure 10. It compares reasonably well with the observed decline shown in figure 5. Some pumping depressions are shifted slightly from their observed positions because all simulated pumping in the model must be assigned to a node of the finite-difference grid located at the center of the grid element, and the nodes commonly do not correspond to the locations of wells.

Simulated net pumpage over the 14-year calibration period is 301,000 acre-ft (371 hm^3), 60 percent of the total discharge of 502,000 acre-ft (619 hm^3). This is 88 percent of the 573,000 acre-ft (707 hm^3) total pumpage estimated by Walker and others (1970) from power-consumption data. The average net pumping rate in the model for 1952–65 is 21,500 acre-ft/yr (26.5 hm^3/yr).

The computed volume of water removed from storage in the aquifer from 1952 through 1965 is 176,000 acre-ft (217 hm^3), representing an average rate of withdrawal from storage of 12,500 acre-ft/yr (15.4 hm^3/yr). This is 58 percent of the volume pumped. The remaining 42 percent of the ground water pumped, or 126,000 acre-ft (155 hm^3), was supplied by captured evapotranspiration. The average annual rate of captured natural discharge equals 8,970 acre-ft/yr (11.1 hm^3/yr).

PREDICTED EFFECTS OF INCREASED PUMPING OF GROUND WATER

The main purpose for which the simulation model was constructed and calibrated was to analyze the regional hydrodynamics of the unconfined aquifer in the southern Raft River Valley. The model also can be used to predict water-level changes caused by artificial recharge or by changes in pumping. These predictions can be made only on a regional or semiregional scale because node spacing is 1 mile (1.6 km) and grid elements cover 1 mi^2 (2.6 km^2). The scale of the model is not appropriate for making detailed predictions of water-level change in areas of only 1–3 mi^2 (2.6 to 8 km^2); these predictions require a larger scale model with node spacing of one-half mile (0.8 km) or less.

Recognizing the limitations of the existing model, water-level change predictions can be made on a semiregional scale covering areas of 10 to 30 mi^2 (26 to 78 km^2). Predictions on this scale will give some indication of the general response of the aquifer to increased pumping or artificial recharge.

Two cases have been selected to demonstrate the use of simulation-modeling techniques for predicting the effects of concurrent pumping and artificial recharge of the unconfined aquifer. It is assumed, for purposes of these predictions, that all increased pumpage will be used for cooling purposes at the proposed geothermal powerplant; there is no increase in the rate of pumping for irrigation. The rate of pumping of the cooling water selected for the predictions is the minimum proposed one of 32,300 acre-ft/yr (39.8 hm^3/yr). For prediction case number one, 50 percent of the pumped water is artificially recharged to the unconfined aquifer, and the rest is consumptively used. For case number two, all pumpage is recharged.

The simulation method requires that only the net recharge or discharge rate be used to represent the flux at any given node in the model. If recharge equals discharge at a node, then the net flux is zero, implying that there would be no change in water level throughout the area represented by the node. This implication represents an inadequate and unreasonable interpretation of the actual effects of pumping and recharge, even on a local scale.

Pumping and artificial recharge in a water-table aquifer leads to problems of nonlinear cause-and-effect relationships. A cone of depresssion forms around each discharging well, reducing the saturated thickness of the aquifer and thereby diminishing its transmissivity. A recharge mound is likely to develop around each recharge well, thereby increasing the saturated thickness and transmissivity of the aquifer. The increase in transmissivity around recharge sites may be more theoretical than real, however, because of the possibility of deterioration of the hydraulic properties of the aquifer as a result of chemical reaction between the native and injected waters.

The rise and decline of water levels may also bring about noncompensating changes in natural recharge-discharge relationships. Water-level decline may reduce evapotranspiration loss, and thus it may reduce discharge to surface streams or induce infiltration of water from streams. Rising water levels may increase evapotranspiration and increase discharge to surface streams. These effects of pumping and artificial recharge do not necessarily balance, and there may thus be a net gain or loss of ground water in storage.

In the following discussion, only the changes caused by pumping and recharging for 10 years are considered. The effects of agricultural pumping during the same 10-year period have been removed. Recharge and discharge sites are the same in both predictions (fig. 11). Discharge rates are the same at each site (2,000 gpm, 125 L/s) and the total rate of 20,000 gpm (1,250 L/s) is the same for each prediction. Recharge rates are equal at each site; the rate is 1,000 gpm (65L/s) for case one and 2,000 gpm (125 L/s) for case two. Each site represents a single well during either recharge or discharge. Pumping rates are near the maximum that might be expected for the area considered. The reasonableness of recharge rates is not known. Most recharge sites were selected so that they would be downgradient from their corresponding discharge sites, in order to minimize increases in ground-water temperature in the areas of pumping. Several recharge sites are located between pumping sites, however, to reduce the impact of pumping.

Figure 12 shows the predicted change in water level if only 50 percent of the water pumped is used to recharge the aquifer. Declines of as much as 75 feet (23 m) occure in the area southeast of Bridge. Recharge wells north of Bridge stop the cone of depression from expanding in that direction. The water level rises as much as 22 feet (7 m) west and southwest of Bridge, suggesting that the recharge rate is too high for the area. The average annual rate of pumping and artificial recharge are given in table 7. The average annual rate of captured natural discharge, principally evapotranspiration, and of water

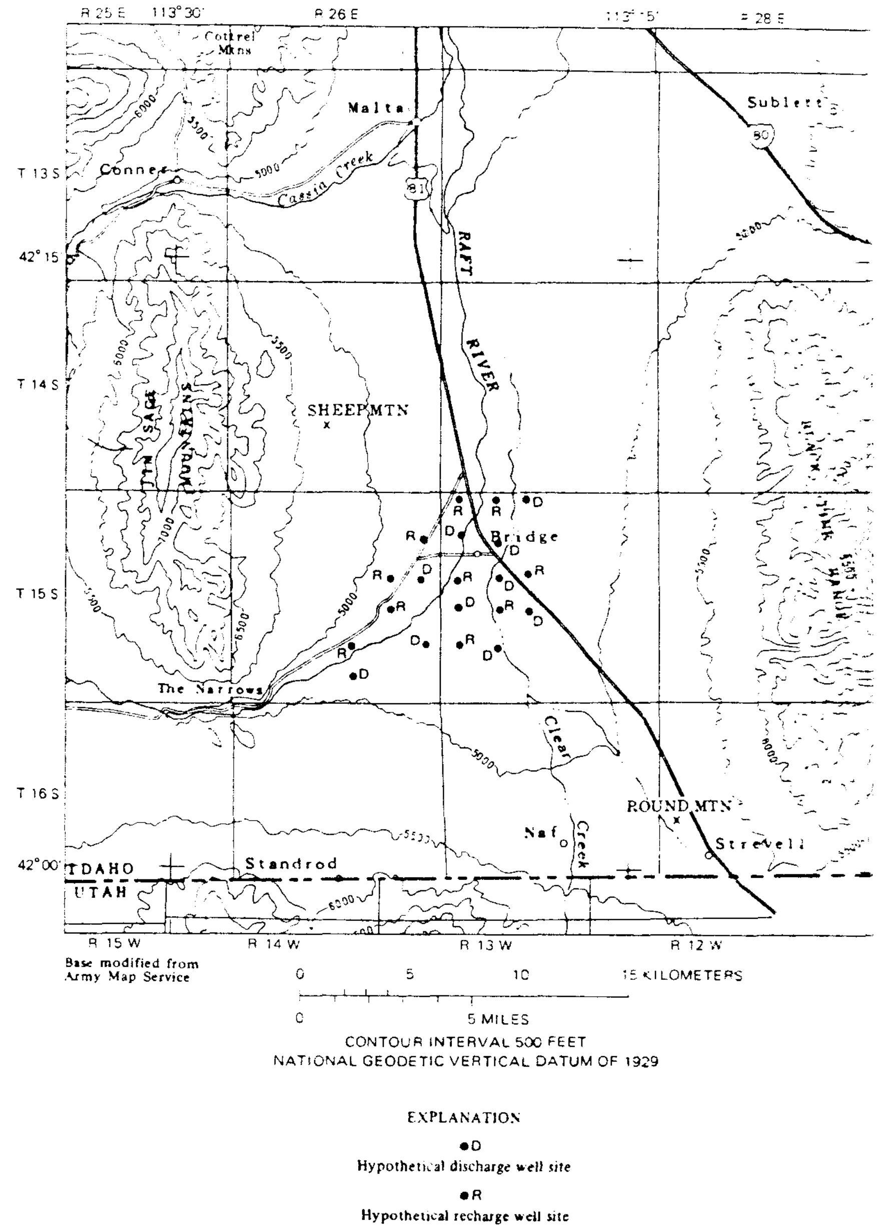
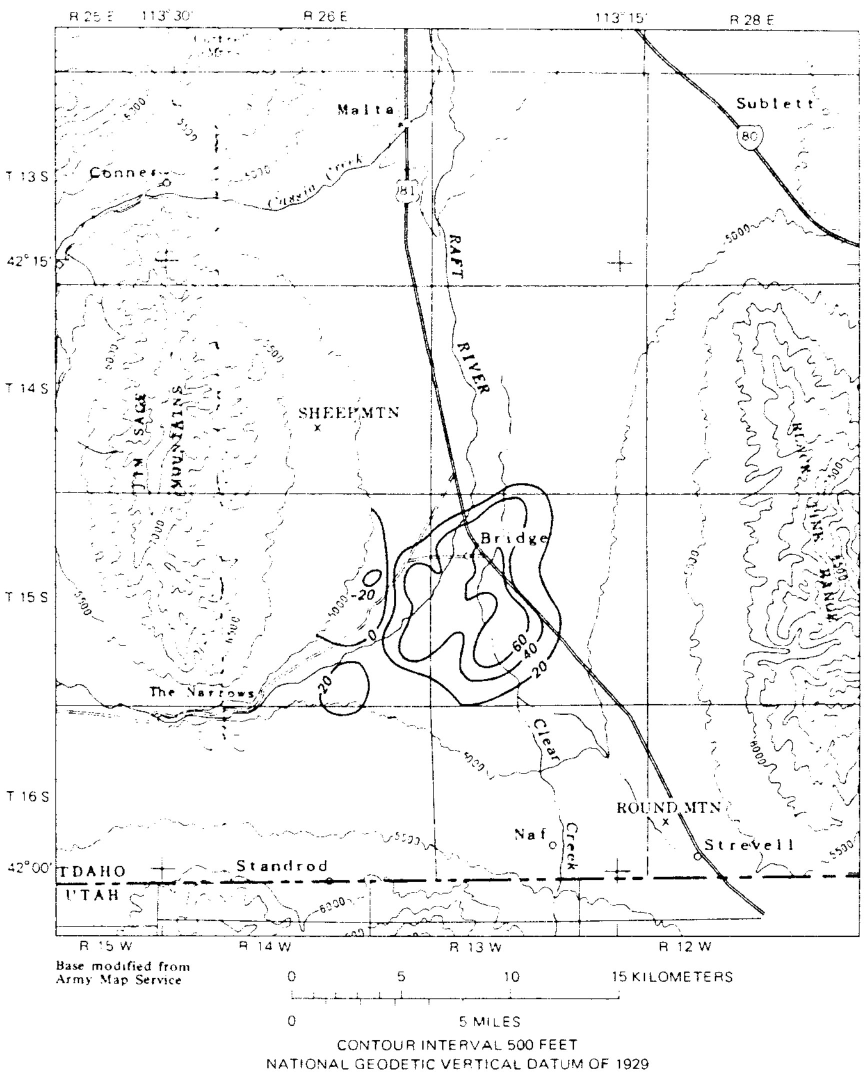


FIGURE 11.—Location of the hypothetical recharge and discharge wells used to predict effects of increased development of the unconfined aquifer.



EXPLANATION

—— -20 ——

Line of equal water level change
(-) indicates rise
Interval 20 feet

FIGURE 12.—Predicted water-level change after 10 years caused by pumping 3,230 acre-ft/yr from each of 10 sites and recharging 1,615 acre-ft/yr at each of 10 other sites.

TABLE 7.—*Sources and average annual volume of pumpage and artificial recharge of cooling water, based on computer simulation prediction*

Average annual volume	100 percent of annual pumpage recharged			50 percent of annual pumpage recharged		
	Acre-feet	Cubic hectometers	Percentage of total	Acre-feet	Cubic hectometers	Percentage of total
Pumped from unconfined aquifer	32,300	39.84	100	32,300	39.84	100
Recharged to unconfined aquifer	30,360	37.45	94	15,180	18.72	47
Removed from aquifer storage	646	.80	2	13,240	16.33	41
Captured from natural discharge	1,292	1.59	4	3,876	4.78	12
Total	32,300	39.84	100	32,300	39.84	100

removed from storage in the aquifer caused by this pumping-recharge scheme also are given in table 7. During the 10-year period, 41 percent of the water pumped for cooling came from storage in the aquifer, and 12 percent was obtained from captured natural discharge.

Figure 13 shows the predicted water-level change after 10 years of artificial recharge at the same total rate as the rate of withdrawal of cooling water. The net recharge rate is slightly less because of existing irrigation pumping at some artificial-recharge sites. Water-level declines of as much as 55 feet (17 m) occur in the area southeast of Bridge. This is about 20 feet (6 m) less than predicted for the previously described recharge-discharge scheme. Considerably less decline is predicted throughout the Bridge area under the 100-percent recharge condition than for the case of 50-percent recharge. Water-level rises of as much as 70 feet (21 m) occur in the area west and southwest of Bridge. This rise is excessive but is not enough to cause water-logging during the 10-year test period. Continuation of this trend for a much longer period would, however, lead to widespread waterlogging conditions west, northwest, and north of Bridge. Further detailed analysis of appropriate recharge rates and sites could be undertaken if a particular development scenario were assumed. Increase in water temperature in the unconfined aquifer is also a consideration.

Average annual rates of pumping and of artificial recharge for the 100-percent recharge scheme are given in table 7. The average annual rate of capture of natural discharge and of removal of water from storage in the aquifer also are given in table 7. During the 10-year period only 2 percent of the volume of cooling water was obtained from aquifer storage, and only 4 percent from captured natural dis-

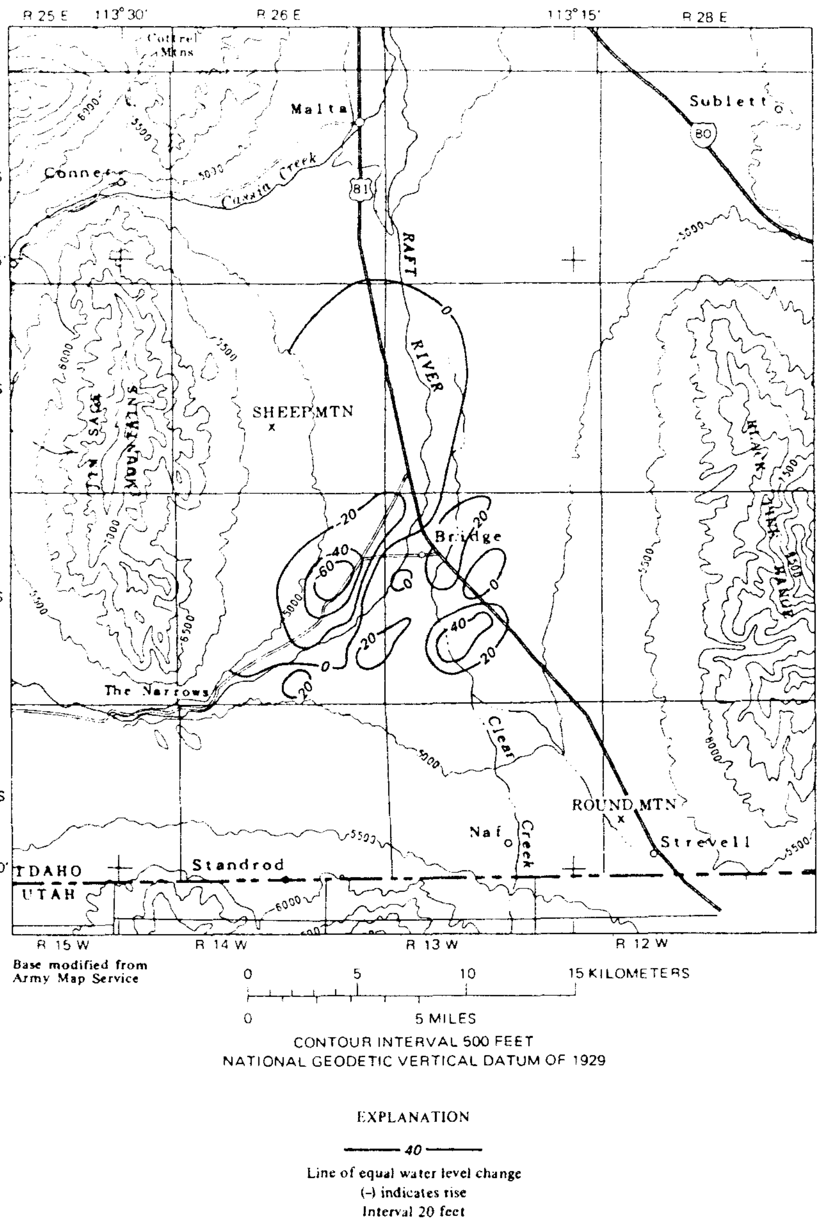


FIGURE 13.—Predicted water-level change after 10 years caused by pumping or recharging 3,230 acre-ft/yr at sites shown in figure 11.

charge. Even so, the pumping-recharge regime had a considerable impact on the distribution and magnitude of water-level change in the aquifer.

CONCLUSIONS

Steady-state and transient simulation analyses have been used to evaluate and modify earlier concepts of the hydrodynamics of the unconfined aquifer in the southern Raft River Valley subbasin. On the basis of these analyses, the average transmissivity for the entire aquifer thickness is believed to be considerably lower than that previously estimated. The modeling, in conjunction with recently obtained subsurface data, shows that shallow underflow through The Narrows is not the principal means of recharge to the aquifer, as had previously been suggested. The model results suggest that only about 20,400 acre-ft/yr (25.2 hm^3/yr) enters the aquifer as recharge by lateral inflow through the entire outer boundary of the subbasin; this represents about one-third of the total recharge. The remaining two-thirds, or about 42,900 acre-ft/yr 52.9hm^3/yr), must enter the aquifer as vertical leakage from below, a source not considered in earlier studies.

The primary mechanism of ground-water discharge under steady-state conditions appears to be evapotranspiration by phreatophytes growing on the valley bottom lands and on their bordering pediments, rather than by downvalley underflow. Discharge by evapotranspiration in the southern Raft River Valley subbasin alone is about 51,000 acre-ft/yr (63 hm^3/yr), with an additional 3,800 acre-ft/yr (4.7 hm^3/yr) lost to surface discharge: Evapotranspiration losses from the entire Raft River Valley subbasin may be as much as 100,000 acre-ft/yr (120 hm^3/yr). Subsurface outflow from the entire basin may be as little as 13,000 acre-ft/yr (16 hm^3/yr).

The volume of underflow out of the southern Raft River Valley subbasin was not significantly reduced by pumpage between 1952 and 1965. Much more significant is the volume of natural discharge salvaged by the capture of evapotranspiration and the reduction of ground-water discharge to streams. For the 14-year period of simulation, the model computed a total captured natural discharge of 126,000 acre-ft/yr (155 hm^3/yr). This amounted to 42 percent of the net volume pumped. The removal of phreatic vegetation from agricultural land might lead to additional reduction in natural discharge, but the net effect would depend on the volume of water pumped for irrigation and the introduction of phreatophytic crops such as alfalfa which can develop deep root systems. The environmental effects of the removal of natural phreatophytic flora are beyond the scope of this report.

Areas or zones of anomalously high or low recharge, discharge, or transmissivity were not required in the model to reproduce 1952 head conditions, nor were any special data manipulations required during transient modeling to reproduce 1952–65 declines. This, together with the absence of any field evidence, strongly suggests that there is little or no significant direct interconnection between the shallow aquifer and the geothermal reservoir—that is, that there is no line or point source or sink connecting the two aquifers. The two aquifers are indirectly connected by leakage upward through relatively thick confining layers with low hydraulic conductivity. That limited direct interconnection exists is indicated by the several warm- and hot-water wells in the eastern half of T. 15 S., R. 26 E., and in the vicinity of The Narrows. For the most part, however, the occurrence of warm water wells in the southern Raft River Valley subbasin is localized and probably results from wells penetrating a nearly sealed fracture system of limited extent through which hot water is circulating. There is little surface expression of this phenomenon.

Principles and theory of the hydrodynamics of aquifer-confining layer interrelationships (Hanshaw and Bredehoeft, 1968; Bredehoeft and Pinder, 1970) can be used to demonstrate the long time required for head or pressure change in the deep geothermal reservoir, in shallower confined aquifers, or in the shallow unconfined aquifer to be transmitted through the thick intervening confining layers after application of a new stress. During the time when newly applied stress changes are taking place in adjacent confining beds, the stressed aquifer is effectively isolated from the effects of pumping in underlying or overlying aquifers, and head or pressure changes within it are caused only by the stress (such as pumping) applied to that aquifer itself. Over the short to intermediate term, development of the geothermal reservoir would have a negligible effect on vertical recharge to, and water levels in, the shallow unconfined aquifer. Using the equations of Hanshaw and Bredehoeft (1968) and reasonable estimates of confining layer parameters, it can be calculated that theoretically the effects of development of the geothermal reservoir might not be felt in the unconfined aquifer for 100 years or more. At the end of this estimated period, the unconfined aquifer would begin to feel the effects of initial pressure change, initiated 100 years earlier.

The effects of increased development of the unconfined aquifer for cooling water can be simulated on a regional or gross scale with the simulation model described in this report. Detailed analysis showing the effects of an individual discharging or recharging well would

require a revised model, based on the original one, using a much more refined finite-difference grid and arrangement of nodes. The area of simulation would have to be limited to the region within 5 or 6 miles (8 to 10 km) of the probable powerplant site in order to reduce the size of the problem and of the data requirements to manageable proportions. Such detailed analysis is beyond the scope of the present report.

The present simulation model was used to demonstrate the semiregional cause-and-effect relationship of the pumping and recharging of powerplant cooling water. The area in which the wells were located at 1 mile (1.6 km) intervals covered about 30 mi² (78 km²), but water levels were affected over a considerably larger area. The predictions, although based on hypothetical conditions and projections, amply demonstrate the effects of increased ground-water pumping in the extreme southern end of the Raft River Valley. If only 50 percent of the pumped water is returned to the aquifer, then significant net volumes of ground water are removed from storage, and large water-level declines occur over areas as large as 15–20 mi² (39–52 km²). Recharging 100 percent of the cooling water causes little change in ground-water storage but still results in significant local water-level declines over about 10 mi² (26 km²); large water-level rises occur over about 2 mi² (5 km²).

Further study is needed to evaluate the full impact of increased development of the unconfined aquifer in the southern Raft River Valley. Consideration should be given to a quantitative investigation of aquifer capabilities and ground-water availability in the Yost-Almo subbasin, because development there will intercept recharge to the Raft River Valley subbasin. Detailed models covering small areas would be needed to determine the impact of pumping and recharging in areas of 1 mi² (2.6 km²) or less. Finally, the effect of the recharge of heated cooling water on local and semiregional ground-water temperatures, which was not considered in this study, needs investigation. The present analysis and simulation model can serve as quantitative guides in developing and conducting these additional studies.

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