# SIMULATION ANALYSIS OF WATER-LEVEL CHANGES IN THE NAVAJO SANDSTONE DUE TO CHANGES IN THE ALTITUDE OF LAKE POWELL NEAR WAHWEAP BAY, UTAH AND ARIZONA

By Blakemore E. Thomas

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## CONVERSION FACTORS

Values in this report are given in inch-pound units. Conversion factors to metric units are listed below.

| Multiply | By | To obtain |
|---|---|---|
| acre | 0.4047 | square hectometer |
| acre-foot | 0.001233 | cubic hectometer |
| acre-foot per mile | 0.001233 | cubic hectometer per mile |
| acre-foot per year | 0.001233 | cubic hectometer per year |
| cubic foot per second | 0.02832 | cubic meter per second |
| foot | 0.3048 | meter |
| foot per day | 0.3048 | meter per day |
| gallon per minute | 0.06309 | liter per second |
| gallon per minute per foot | 0.207 | liter per second per meter |
| inch | 25.40 | millimeter |
| mile | 1.609 | kilometer |
| square mile | 2.590 | square kilometer |
| square foot per day | 0.0929 | square meter per day |
| square foot per second | 0.0929 | square meter per second |

National Geodetic Vertical Datum of 1929 (NGVD of 1929): A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called mean sea level, is referred to as sea level in this report.

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## ABSTRACT

A two-dimensional, finite-difference, digital-computer model was used to simulate various concepts of ground-water flow in the Navajo Sandstone near Wahweap Bay, Lake Powell, Utah and Arizona. The filling of Lake Powell started in March 1963; and by 1983 the lake had risen almost 550 feet. This resulted in a maximum observed water-level rise of 395 feet in a well in the Navajo Sandstone 1 mile from the lake.

A steady-state model was prepared with subsurface recharge rates of 5,720, 10,440, and 14,820 acre-feet per year, resulting in a range of hydraulic conductivity of 0.25 to 3.38 feet per day. Comparing measured and simulated water-level changes resulted in a range of specific yield of 0.02 to 0.15. Using larger values for hydraulic conductivity in the model area corresponding to the axis of the Wahweap syncline and the Echo monocline was instrumental in attaining a reasonable match for the water-level distribution. This supports previous concepts that areas where rocks are structurally deformed more readily transmit ground water because of the higher degree of fracturing. Using the most likely simulation of the flow system, ground-water storage in the Navajo increased by about 25,000 acre-feet per mile of shoreline from 1963-83, but the flow system will require about 400 years to reach a state of equilibrium.

## INTRODUCTION

This report was prepared as part of the Regional Aquifer Systems Analysis (RASA) of the Upper Colorado River Basin that is being done by the U.S. Geological Survey. Objectives of the RASA include compiling a data base to define existing hydrologic conditions, identifying and describing major aquifers, analyzing regional systems of ground-water flow, and evaluating the impacts of selected developmental alternatives on the hydrologic system.

The purpose of this report is to aid the regional analysis of ground-water flow by examining plausible ranges for values of hydraulic conductivity and recharge that are used to define the flow system in the Navajo Sandstone, the region's principal aquifer. Another objective is to evaluate the existing and future effects of Lake Powell on the aquifer. These objectives are met most realistically by flow modeling, and the area chosen for the model is at the west edge of Lake Powell in Utah and Arizona (fig. 1).

This area was selected for modeling because it includes most of the water-level data that are available for the Lake Powell area. The modeled area, which covers about 600 square miles, is a small part of the 7,400-square-mile study area. The study area includes all the physical boundaries of the ground-water system that are relevant to the model boundaries.

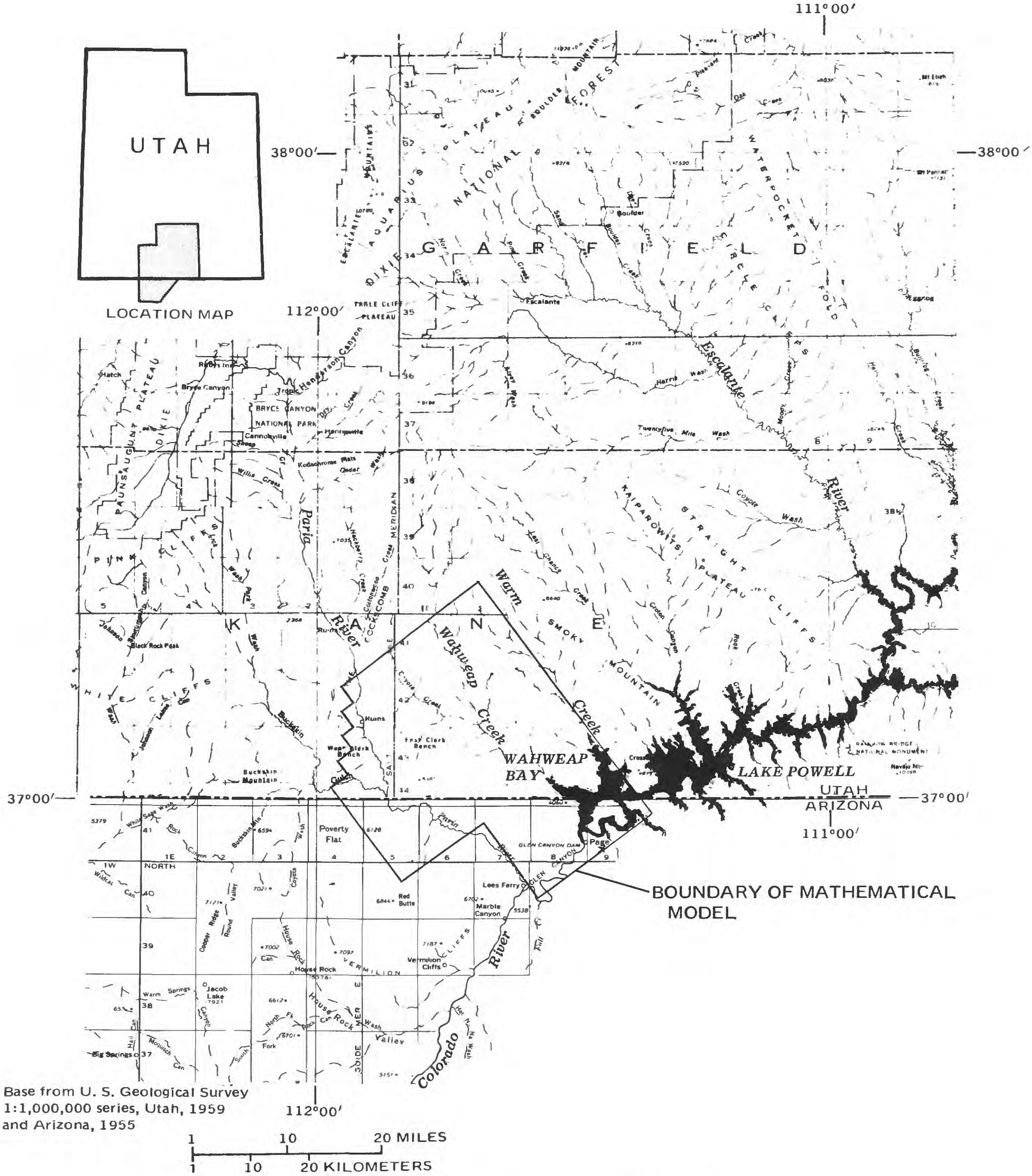



Figure 1.—Location of study and modeled areas.

A two-dimensional, finite-difference, digital-computer model was designed to test estimated and measured values for hydraulic properties and boundary-flow conditions of the aquifer. Transient-state simulations further tested these hydrologic concepts by simulating changes in the ground-water levels that were caused by inflow of water from Lake Powell. The small amount of data available, however, did not allow extensive comparisons with the simulations.

Blanchard (1984) studied the ground-water system near Wahweap Bay, Lake Powell, as part of a reconnaissance of ground water between the Paria and Escalante River basins. Most of the data used in the model in this report were taken from Blanchard's report.

## WELL-NUMBERING SYSTEM

The system of numbering wells in Utah is based on the cadastral land-survey system of the U.S. Government. The number, in addition to designating the well, describes its position in the land net. By the land-survey system, the State is divided into four quadrants by the Salt Lake base line and meridian, and these quadrants are designated by the upper case letters A, B, C, and D, indicating the northeast, northwest, southwest, and southeast quadrants, respectively (fig. 2). Numbers designating the township and range (in that order) follow the quadrant letter, and all three are enclosed in parentheses. The number after the parentheses indicates the section, and it is followed by three letters indicating the quarter section, the quarter-quarter section, and the quarter-quarter-quarter section--generally 10 acres[1]. The letters a, b, c, and d indicate, respectively, the northeast, northwest, southwest, and southeast quarters of each subdivision. The number after the letters is the serial number of the well within the 10-acre tract. If a well cannot be located within a 10-acre tract, one or two location letters are used and the serial number is omitted. Thus, (C-43-1)4bad-1 designates the first well constructed or visited in the SE1/4NE1/4NW1/4 sec. 4, T.43 S., R.1 W. Wells in Arizona are numbered in the same manner, except that the numbering system is based on the Gila and Salt River base line and meridian.

---

[1]Although the basic land unit, the section, is theoretically 1-square mile, many sections are irregular. Such sections are subdivided into 10-acre tracts, generally beginning at the southeast corner, and the surplus or shortage is taken up in the tracts along the north and west sides of the section.

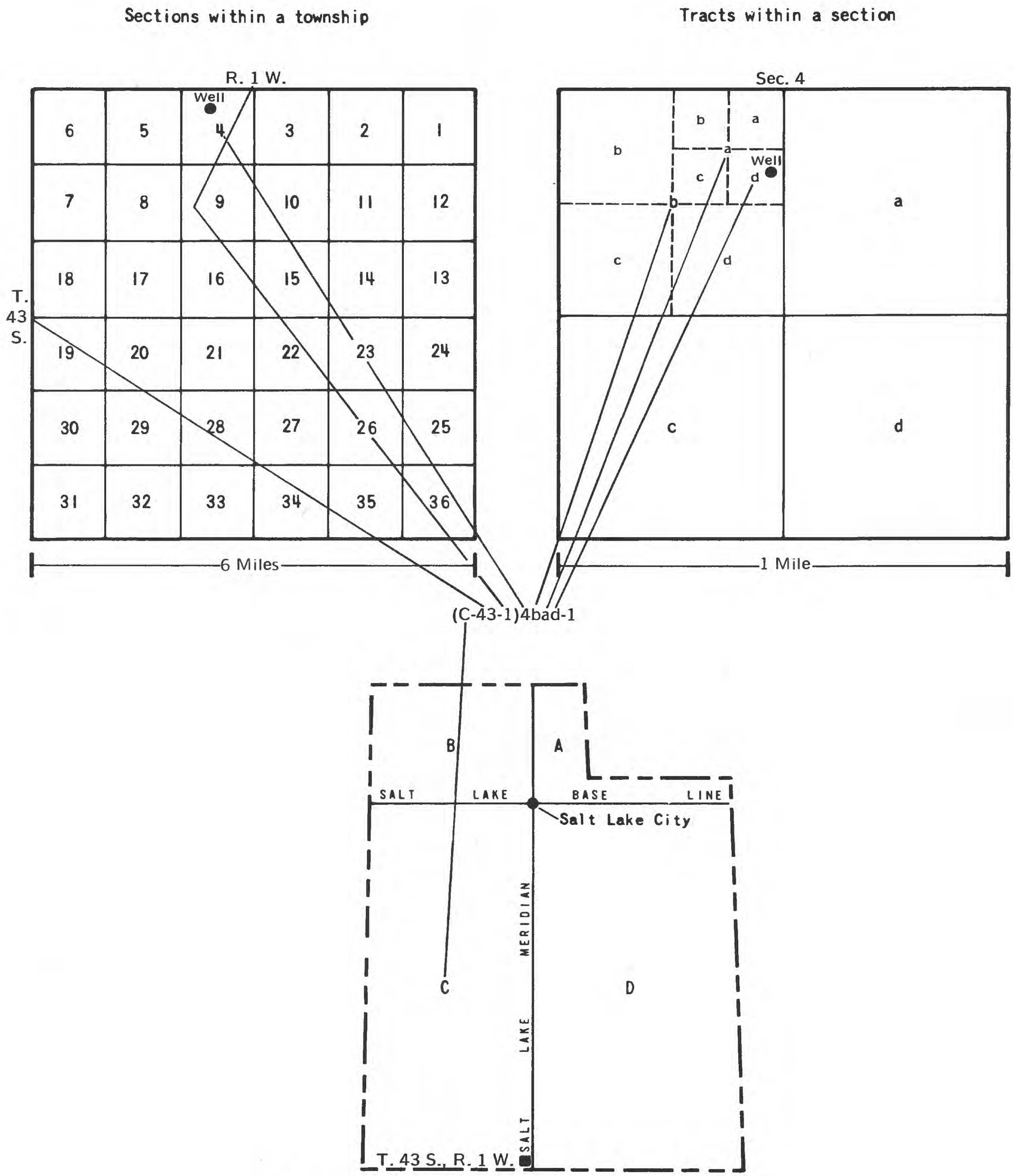



Figure 2. —Well-numbering system used in Utah.

## GEOHYDROLOGIC SETTING

### General Setting

The study area is in the southwest part of the Colorado Plateau physiographic province described by Fennemen (1931, p. 274-325). This province is characterized by broad plateaus and narrow steep-walled canyons. Altitudes of the high plateaus in the north part of the study area (Paunsaugunt, Table Cliff, and Aquarius Plateaus) range from 9,000 to 11,000 feet.

The modeled area is typical of the Plateau province and is a broad plateau cut by canyons of the Paria River, Wahweap Creek, and other streams that head in higher plateaus to the north (fig. 1). Altitudes of the modeled area range from about 3,120 feet at Lees Ferry at the south side, to about 6,500 feet on the Paria Plateau, to about 6,000 feet at the north corner. In its lower reach, the Paria River flows through a deep narrow canyon which has 3,000 feet of relief between the Paria Plateau and the riverbed near Lees Ferry. The bench between Wahweap Creek and Warm Creek is about 1,000 feet above the bottom of the canyons.

The southeast boundary of the modeled area coincides approximately with Glen Canyon, which is about 600 feet deep. Glen Canyon Dam is near the middle of this boundary. Upstream from the dam, the canyon is innundated by as much as 500 feet of water by Lake Powell, and downstream the Colorado River flows through the canyon.

The average annual precipitation in the study area ranges from 6 inches near Lees Ferry in the south to 30 inches in the high plateaus in the north. The model area receives an average of less than 10 inches per year (U.S. Weather Bureau, 1963). The precipitation is variable, with some snow or rain resulting from frontal storms during November through April and the remainder of the precipitation resulting from thunderstorms during May through October.

The major streams that flow through the model area are the Paria River and Wahweap Creek. Warm Creek is along the northeast boundary and the Colorado River and Lake Powell form the southeast boundary. Streamflow in the Paria River is perennial and results mostly from snowmelt and small contributions from discharge of ground water. All other streams in the model area, except the Colorado River, are intermittant or emphemeral.

### Geologic Framework

#### Stratigraphy

The rocks in the study area are mostly sedimentary formations that were deposited under marine or continental conditions. A few igneous intrusions in the northeast part of the study area are not relevant to this study. A generalized section of the sedimentary rock formations in the study area is shown in table 1. The maximum thickness of each formation is given as well as the range in thickness in the model area. Rocks of Permian age are exposed south and west of the Paria Plateau, however, only Triassic and younger rocks are shown in table 1. The ground-water system in the Navajo Sandstone is isolated from water in Permian rocks.

Table 1.--Generalized section of sedimentary rock formations in the study area. [Adapted from Blanchard (1984), Doelling and Graham (1972), and Phoenix (1963)]

| Age | Stratigraphic Unit | Maximum thickness (feet) | Description |
|---|---|---|---|
| Tertiary | Wasatch Formation | 1,700 | Thick-bedded, fine-grained, fluvial or lacustrine, clastic silty limestone containing thin interbeds of mudstone and sandstone. Absent from model area. |
| Cretaceous | Kaiparowits Formation | 3,000 | Fine- to moderately coarse-grained, friable arkosic sandstone, interbedded with thin beds of mudstone. Absent from model area. |
| | Wahweap Sandstone | 1,500 | Alternating thin to thick fluvial beds of sandstone and mudstone. Absent from model area. |
| | Straight Cliffs Formation | 1,900 | Fine- to coarse-grained locally conglomeratic sandstone interbedded with shale and mudstone. Absent from model area. |
| | Tropic Shale | 1,000 | Argillaceous to sandy shale, contains thin sandstone beds at top and base, otherwise uniform. Absent from model area. |
| | Dakota Sandstone | 250 | Coarse-grained sandstone and conglomerate overlain by ripple-bedded sandstone and laminated siltstone. Caps the mesas in northeast part of model area and is about 100 feet thick. |
| Jurassic | Morrison Formation | 700 | Variegated continental beds of sandstone, conglomerate, and bentonitic mudstone. Absent west of Wahweap Creek and about 100 feet thick at east corner of model area. |
| | Summerville equivalent | 200 | Sandstone and shaly siltstone in even, thin alternating beds. Absent from model area. |
| | Entrada Sandstone | 900 | Massive medium- to fine-grained crossbedded sandstone. About 650 feet thick in model area. |
| | Carmel Formation | 900 | Thin beds of limy siltstone, fine-grained friable sandstone, limestone, and gypsum, all of marginal marine origin. Thickness ranges from 400 to 600 feet in model area. |
| Jurassic and Triassic(?) | Navajo Sandstone (includes overlying Page Sandstone) | 2,000 | Massive medium- to fine-grained sandstone, exhibiting large-scale aeolian cross-bedding. Thickness ranges from 1,700 to 2,000 feet in model area. |
| Triassic(?) | Kayenta Formation | 200 | Fluvial sandstone, siltstone, shale, and minor shale-pellet conglomerate and freshwater limestone. Thickness ranges from 100 to 200 feet in model area. |
| | Moenave Formation | 450 | Composed of two members: Springdale Sandstone Member is medium-grained, micaceous sandstone and minor siltstone; and underlying Dinosaur Canyon Sandstone Member is coarse- to fine-grained parallel-bedded sandstone and siltstone. Thickness ranges from 250 to 450 feet in model area. |
| Triassic | Wingate Sandstone | 300 | Fine-grained, thickly crossbedded, calcareous aeolian sandstone. Absent from model area. |
| | Chinle Formation | 1,100 | Varicolored beds of fluvial and lacustrine origin, generally sandy at top; limy, muddy, and bentonitic in the middle; and sandy and conglomeratic near base. Thickness ranges from 900 to 1,100 feet in model area. |

Outcrop areas of the Navajo Sandstone and rocks older and younger than the Navajo are shown in figure 3. Tertiary rocks crop out in the high plateaus in the north part of the study area. Cretaceous rocks comprise most of middle part of the area (Kaiparowits Plateau) and outcrops of the Navajo are east and west of the Plateau. The Dakota Sandstone is the youngest formation that crops out in the model area.

The Navajo Sandstone of Triassic(?) and Jurassic age contains the principal aquifer that is being examined in this study. The Navajo is a crossbedded sandstone of eolian origin which ranges in thickness from about 1,700 feet at the east edge of the model area to about 2,000 feet at the west edge. It is buff to very pale orange, well-rounded, well-sorted, and mostly fine-grained. The Carmel Formation of Jurassic age, which overlies the Navajo, ranges in thickness from 400 to 600 feet in the model area. It consists of thin beds of dusky-red limy siltstone, reddish-brown fine- grained friable sandstone, gray to pink limestone, and thin to thick beds of gypsum, all of marginal marine origin. The Kayenta Formation of Triassic(?) age, which underlies the Navajo, ranges in thickness from 100 to 200 feet in the model area. It consists of red-brown, reddish-orange, pale-gray, and lavender fluvial sandstone, siltstone, shale, and minor amounts of shale- pellet conglomerate and freshwater limestone (Blanchard, 1984, p. 7-12).

## Structure

The structure of the sedimentary beds in the study area is shown using contour lines drawn on the base of the Dakota Sandstone (fig. 4). The thickness of the formations between the Dakota and Navajo varies throughout the study area, however, the regional structure for the Navajo is similar to the Dakota. Two major folds on the east side (Waterpocket monocline) and west- central side (East Kaibab monocline) of the study area trend in a northerly direction and divide the region into two major structural features. The Kaiparowits basin is between the two monoclines and the Kaibab uplift is southwest of the East Kaibab monocline. Cretaceous rocks are exposed in the Kaiparowits basin and Permian rocks form the platform of the Kaibab uplift. The lowest part of the Kaiparowits basin is north of the model area, and this trough extends southward through the east side of the model area.

A more detailed structure map was prepared for the model area (fig. 5). The structure contours have a 250 foot interval and are on the top of the Navajo Sandstone. The altitudes of the top and bottom of the Navajo were estimated from driller's logs of water wells and interpolated from a structure contour map of the Dakota Sandstone, which overlies the Navajo (Hackman and Wyant, 1973). The Navajo Sandstone dips to the northeast, and it has been intensly fractured as a result of three major structures in the area. They are the East Kaibab monocline, which trends north and cuts across the west corner of the model area; the Echo monocline, which trends north in the middle of the area; and the Wahweap syncline, which trends mostly north near the east side of the area.

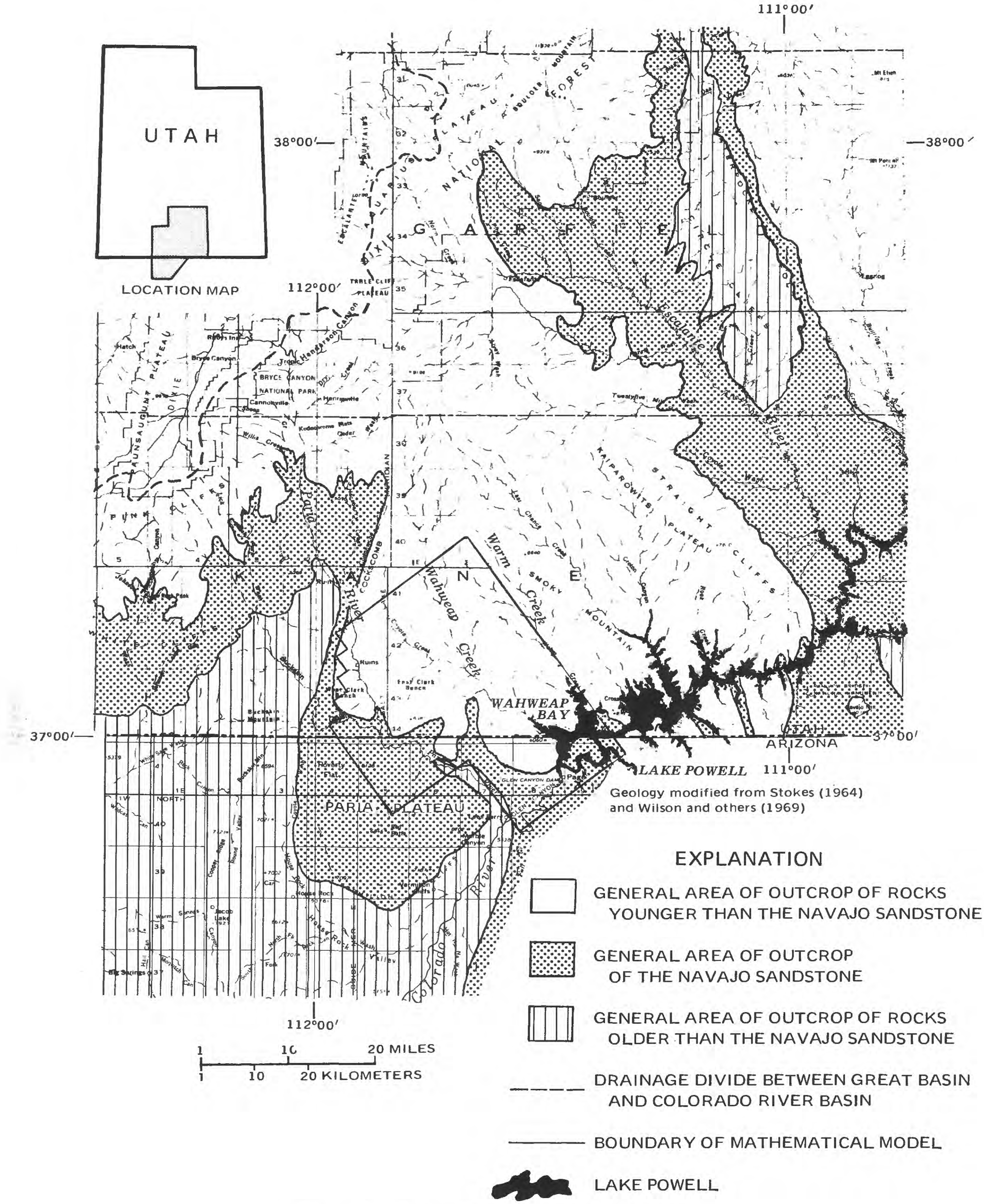



Figure 3.—Generalized geology of study area.

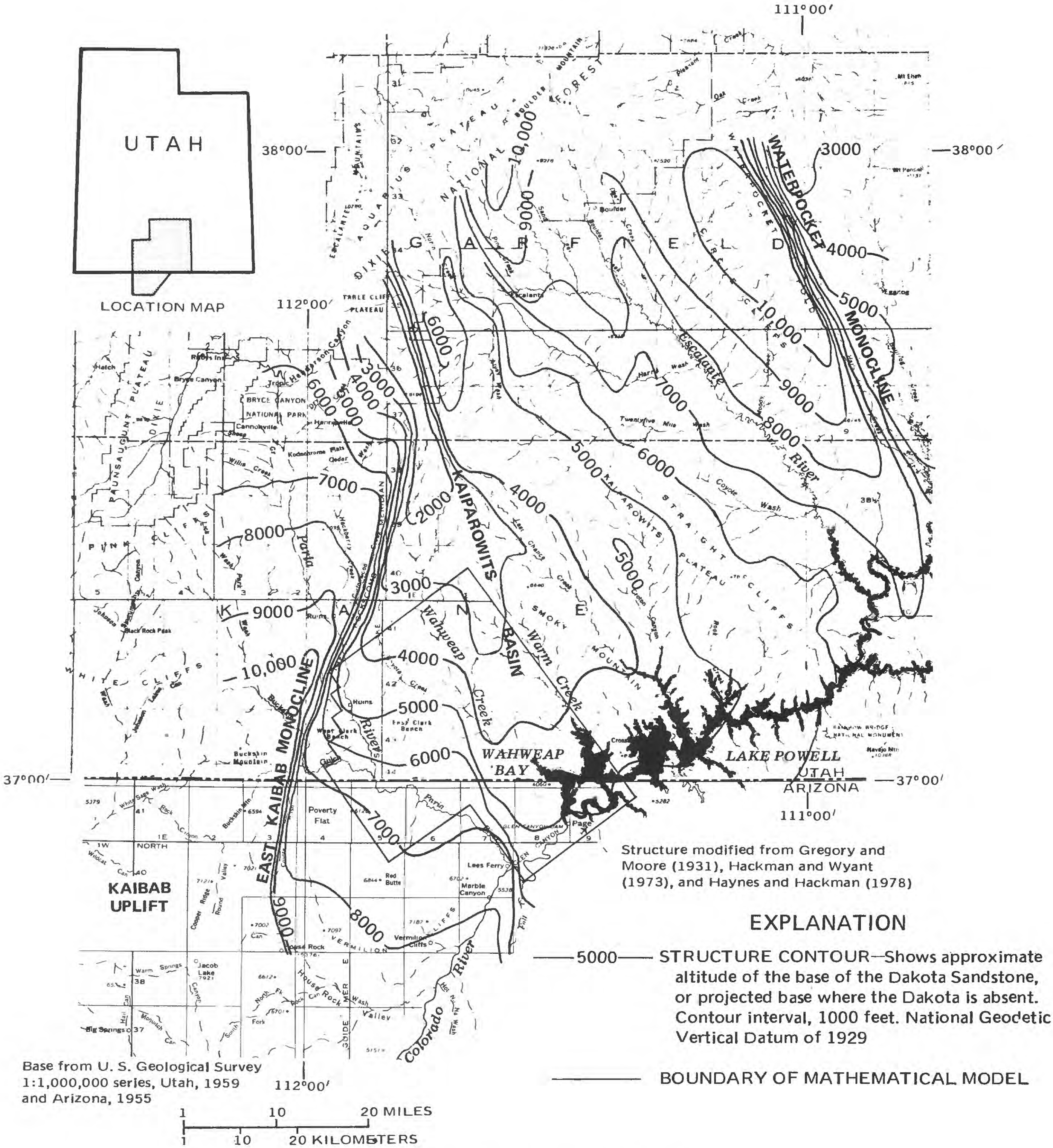



Figure 4.—Generalized structure of study area.

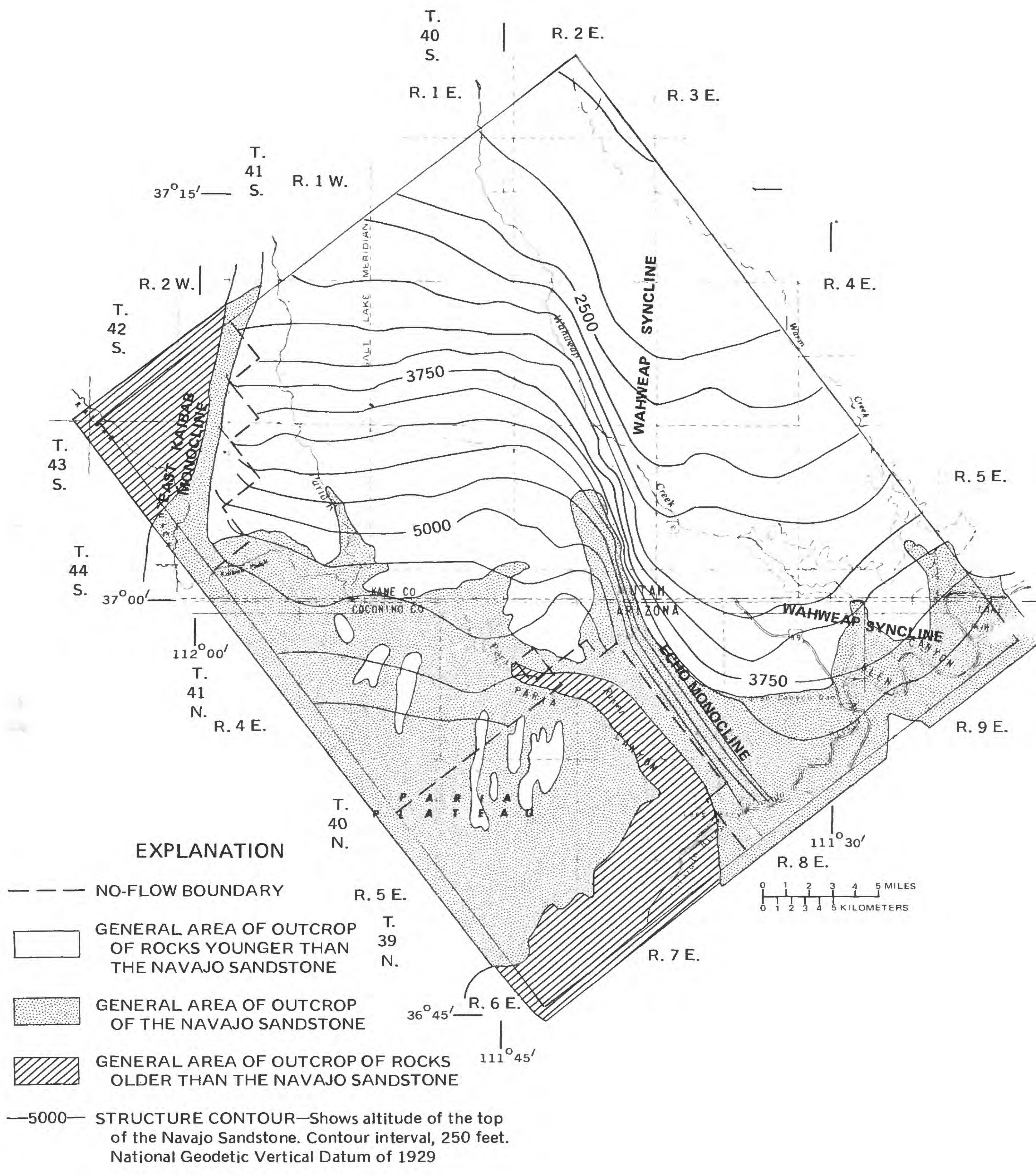



Figure 5.—Altitude of the top of the Navajo Sandstone and areas of outcrop.

## Ground-Water System

The Navajo Sandstone is the principal aquifer in the study area. All other formations contain water, but the Navajo covers the greatest regional extent and has the most extensive contact with Lake Powell. The Navajo constitutes about 40 percent of the shoreline of Lake Powell, and it includes more than 80 percent of the space available in the rocks for bank storage (Jacoby and others, 1977, p. 47).

### Boundaries of System

The boundaries for ground-water flow in the Navajo Sandstone were estimated from the surficial geology (fig. 3), the structure of the sedimentary rocks (fig. 4), and the potentiometric surface (fig. 6). The potentiometric contours and arrows showing direction of flow shown in figure 6 were adapted directly from Blanchard (1984, fig. 5). The boundary of the model area is shown on all these figures and a discussion of the relation between the system and model boundaries is given in the section on boundary conditions of the model.

#### Steady-State Boundaries

The ground-water system in the Navajo Sandstone is assummed to have been in a steady-state condition prior to March 1963. Steady state or equilibrium is a condition where water levels and storage in the aquifer do not change over time, and the quantity of recharge to the system is balanced by an equal quantity of discharge. Since the filling of Lake Powell started in March 1963, water levels in the Navajo Sandstone have been changing and the system has been in a transient-state or nonequilibrium condition.

The Navajo Sandstone is overlain by the Carmel Formation and underlain by the Kayenta Formation. These formations have beds of low permeability, such as shale and siltstone, interbedded with sandstone beds; thus, they act as confining beds that inhibit the vertical movement of water (Blanchard, 1984, p. 7-12). Some water probably does move through these formations where a vertical water-level gradient exists, but the quantity probably is small. Water in the Navajo is under unconfined conditions in the west and south parts of the modeled area and under confined conditions in the east and north parts.

Some of the horizontal boundaries for water in the Navajo Sandstone can be seen on figure 3. The Navajo extends beyond the limits of the study area in all directions except to the southwest where older rocks crop out. The boundary between the Navajo Sandstone and older rocks is a no-flow boundary. Along the steeply dipping East Kaibab monocline (fig. 4) west and southwest of the model area, the Kayenta, Moenave, and Chinle Formations crop out in a narrow band. The Chinle is about 1,000 feet thick and consists mostly of siltstone, mudstone, and claystone. The Chinle and Kayenta form an effective barrier to flow of water between older rocks and the Navajo.

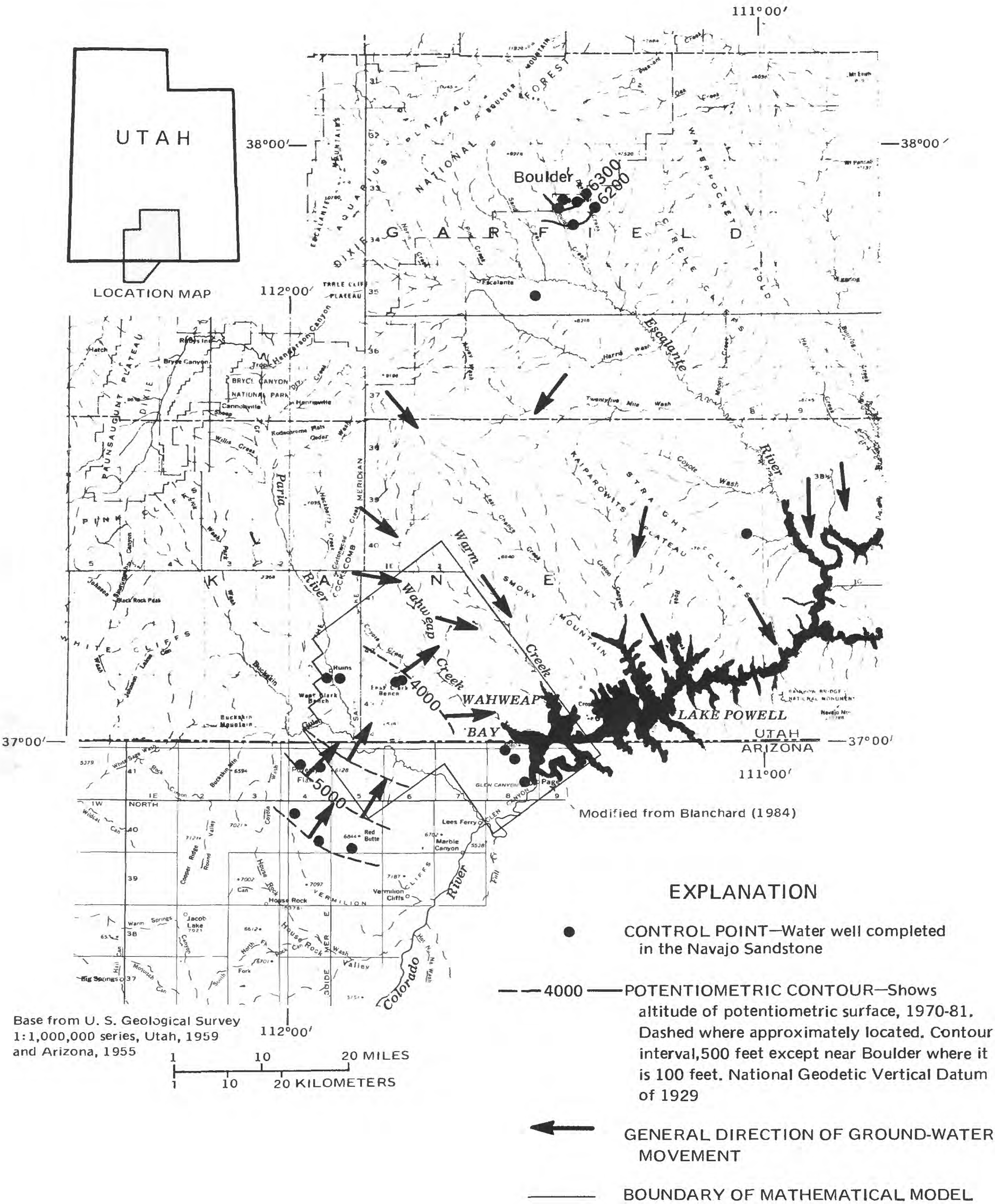



Figure 6.—Approximate potentiometric surface and general direction of movement of water in the Navajo Sandstone.

The Navajo Sandstone is deeply incised by the Colorado River. Cooley and others (1969, p. 44) state that ground water in the Navajo discharges into the Colorado River from both the north and south. The potentiometric map for the study area (fig. 6) shows the regional movement of ground water is toward the south and the Colorado River. Thus, the Colorado River acts as a fully penetrating stream, and it is assumed that no significant quantity of ground water moves under the river.

### Transient-State Boundaries

The ground-water system has been in a transient-state condition since the filling of Lake Powell started in March 1963. The steady-state system and the transient-state system have the same boundaries except the canyon of the Colorado River above Glen Canyon Dam changes from the primary discharge boundary in steady-state conditions to a boundary of mostly recharge in transient-state conditions. It is assummed that the change in boundaries from a river to a lake does not change the fully penetrating nature of the boundary. Ground water still does not move under the lake and stresses on one side of the lake would not affect water levels on the other side.

During the 20-year period represented in this study, 1963-83, the water level of Lake Powell changed because of weather conditions, which primarily affected the inflow, and management decisions, which regulated the outflow. From March 1963 to June 1966, the lake rose from an initial altitude of 3,152 feet (the stage of the Colorado River at Glen Canyon Dam) to 3,545 feet. From June 1966 to July 1973, the lake gradually rose to about 3,650 feet. From July 1973 to March 1983, the lake fluctuated between 3,620 and 3,700 feet (U.S. Geological Survey, 1963-75, 1976-84).

The area of confined conditions for the aquifer has increased as a result of the rise in water levels near the lake. In the model area, the top of the Navajo ranges from about 3,000 feet at the north edge of Lake Powell to about 3,900 feet at Glen Canyon Dam. Thus, the lake has been higher than the top of the Navajo in most of the shoreline area; and this has resulted in the increase in area of confined conditions.

## Hydraulic Properties of the Navajo Sandstone

Few data are available for the hydraulic properties of the Navajo Sandstone in the study area. Blanchard (1984, tables 8 and 12) reported specific capacities of 3.5 and 12.5 gallons per minute per foot for wells (A-42-8)32cdd-1, and (A-42-8)36ccc-1 and a transmissivity of about 7,000 square feet per day for well (A-42-8)35dab-2. The specific capacities convert to hydraulic conductivities of about 0.4 and 1.3 feet per day, and the transmissivity converts to a hydraulic conductivity of about 3.5 feet per day. A study by the U.S. Bureau of Reclamation (Calder, L., written commun., 1965) showed that wells near Wahweap Creek had much larger yields (about 1,000 gallons per minute) than wells near Glen Canyon Dam (about 30 gallons per minute). Calder concluded that fracturing in the Wahweap syncline and Echo monocline increased the permeability of the Navajo Sandstone.

A reasonable range of values for hydraulic conductivity, specific yield, and storage coefficient was assumed on the basis of previous studies of the Navajo Sandstone in southern Utah and Arizona (Blanchard, 1984; Cordova, 1978, 1981; Eychaner, 1983; Hood and Danielson, 1979, 1981). The range for hydraulic conductivity is 0.2 to 10.0 feet per day; for specific yield the range is 0.02 to 0.15; and for storage coefficient it is 0.001 to 0.000001.

## Recharge

Prior to the filling of Lake Powell, recharge to the Navajo Sandstone in the modeled area was primarily from subsurface inflow from outside the modeled area. This recharge is from infiltration of rainfall and snowmelt on outcrops of the Navajo to the northeast, northwest, west, and on the Paria Plateau to the southwest (fig. 3). Also, some recharge probably reaches the Navajo from downward leakage from overlying formations at the higher altitudes north of the model area. The average annual precipitation on the northwest outcrop area and the East Kaibab Monocline is about 12 inches, and on the Paria Plateau it is about 14 inches. Recharge from precipitation directly on the outcrop of the Navajo in the modeled area probably is small because the average annual precipitation on the area is less than 10 inches per year. In the north part of the model area, water levels in the Navajo are below the bed of the Paria River; thus, the stream is a source of recharge to the aquifer.

Subsurface inflow to the modeled area is through the southwest, northwest, and northeast borders. The quantity of subsurface inflow was estimated on the basis of studies by Blanchard (1984), Cordova (1978 and 1981), and Eychaner (1983) to be between 5,000 and 15,000 acre-feet per year. Most of the subsurface inflow probably is from the Paria Plateau and East Kaibab monocline.

During 1963-83, an additional large quantity of recharge entered the Navajo Sandstone from Lake Powell. This is reflected in the rise of water levels in wells near the lake (table 2). A water-budget study by Jacoby and others (1977, p. 64) concluded that water seeping into the rocks around the lake (bank storage) totaled about 6 million acre-feet after 10 years of recharge. R. E. Glover (U.S. Bureau of Reclamation, written commun., 1949) estimated the seepage losses into the rocks around the lake using an equation based on Darcy's Law. Assuming a shoreline of 300 miles, he estimated that the seepage would total 5.6 million acre-feet after 10 years. Thus, the computations by two methods produced similar results. To obtain an estimate of bank storage for the model area, the bank storage can be converted to acre-feet per mile of shoreline. This is a reasonable transfer of values because the Navajo Sandstone comprises over 80 percent of the total space available in the rocks for bank storage at Lake Powell (Jacoby and others, 1977). Assuming a shoreline of 300 miles and a bank storage of 5.8 million acre-feet after 10 years, this converts to about 19,000 acre-feet per mile of shoreline.

## Movement

The approximate altitude of the potentiometric surface and general direction of movement of water in the Navajo Sandstone are shown in figure 6. Water levels used for this map were measured during 1970-81. The general flow pattern shown on this map is the same for steady-state conditions (before March 1963) and transient-state conditions (1963-83).

Table 2.--Measured water levels used in model calibration
Well number and grid location: the nodes are row-column.

| Well number and grid location | Stress period in model | Measured water levels: Date | Measured water levels: Altitude (feet) | Measured water levels: Water-level change (feet) |
|---|---|---|---|---|
| (C-43-1)4bad-1 node 3-4 | 10 | Aug. 1971 | 4,300 | -- |
| (D-43-1)2bdd-1 | steady-state | Sept. 1957 | 4,035 | -- |
| (D-43-1)2cab-1 node 6-10 | 20 | May 1981 | 4,050 | +15 |
| (A-42-8)32cdd-1 node 19-16 | 12 | Oct. 1972 | 3,518 | -- |
| | 13 | Sept. 1973 | 3,547 | +29 |
| | 15 | July 1976 | 3,557 | +10 |
| | 20 | June 1981 | 3,590 | +33 |
| (A-41-8)4dda-1 node 22-19 | 20 | June 1981 | 3,668 | -- |
| | 22 | Apr. 1983 | 3,676 | +8 |
| (A-41-8)14bcb-1 node 26-19 | steady-state | Oct. 1957 | 3,240 | -- |
| | 18 | Sept. 1979 | 3,610 | +370 |
| | 20 | Feb. 1981 | 3,627 | +17 |
| | 22 | Apr. 1983 | 3,635 | +8 |
| (A-42-8)36cbc-1 node 23-24 | steady-state | Jan. 1959 | 3,313 | -- |
| | 3 | June 1964 | 3,351 | +38 |
| | 4 | Mar. 1965 | 3,408 | +57 |
| | 5 | Mar. 1966 | 3.462 | +54 |
| | 6 | Mar. 1967 | 3,485 | +23 |
| | 7 | Mar. 1968 | 3,501 | +16 |
| | 8 | Mar. 1969 | 3,522 | +21 |
| | 9 | Mar. 1970 | 3,553 | +31 |
| | 10 | Mar. 1971 | 3,583 | +30 |
| | 11 | Mar. 1972 | 3,596 | +13 |
| | 12 | Mar. 1973 | 3,593 | -3 |
| | 13 | Mar. 1974 | 3,633 | +40 |
| | 14 | Mar. 1975 | 3,635 | +2 |
| | 15 | Mar. 1976 | 3,653 | +18 |
| | 16 | Mar. 1977 | 3,644 | -9 |
| | 17 | Mar. 1978 | 3,621 | -23 |
| | 18 | Mar. 1979 | 3,626 | +5 |
| | 19 | Mar. 1980 | 3,666 | +40 |
| | 20 | Mar. 1981 | 3,672 | +6 |
| | 21 | Mar. 1982 | 3,660 | -12 |
| | 22 | Mar. 1983 | 3,677 | +17 |
| (A-41-8)23dcd-1 node 30-17 | steady-state | July 1957 | 3,182 | -- |
| (A-41-8)23dac-1 node 30-18 | steady-state | Jan. 1958 | 3,182 | -- |

Water-level data are not available for much of the study area, therefore, the direction of ground-water movement was inferred from the boundaries of the aquifer (fig 3.), the structure of the sedimentary beds (fig. 4), and the location of recharge and discharge areas. Ground water moves from recharge areas surrounding the Kaiparowits basin down the structural trough of the basin, and then southeast to discharge in the Colorado River or Lake Powell. Upstream from Lees Ferry and within a band of about 10 miles on each side of the Colorado River (before 1963) or Lake Powell (after 1963), it is assumed that ground water moves perpendicular to the river or lake.

Since Lake Powell came into existence, the general direction of ground-water movement has not changed. Water from Lake Powell is recharging the Navajo Sandstone near the lake, but the regional flow system is still moving toward the lake. The major changes to the system are within about 20 miles of the lake shoreline. In this area, the water-level gradient toward the lake has flattened as water levels near the lake rise in response to recharge from the lake. Table 2 shows water-level changes measured in wells completed in the Navajo Sandstone in the model area (Blanchard, 1984, tables 12 and 13, and U.S. Bureau of Reclamation, Salt Lake City, Utah, written commun., 1983). Water levels in wells within 1 mile of the lake shoreline indicate that the direction of ground-water movement near the lake reverses following the seasonal fluctuations of the lake level.

## Discharge

Prior to the existence of Lake Powell, most of the discharge of water from the Navajo Sandstone went to the Colorado River. In the model area, some water also discharged to the Paria River and the sides of the Paria River canyon. Although no estimates are available for the quantity of discharge, Cooley and others (1969, p. 44) suggested that the Colorado River was the primary discharge area.

Since the filling of Lake Powell, water in the Navajo Sandstone that originally discharged to the Colorado River is now either going into storage, discharging to springs or streams near Lake Powell, discharging to the lake, or discharging to the Colorado River downstream from Glen Canyon Dam. The relative amounts of this pre-lake discharge that goes to these different areas can not be estimated. In the model area, water is still discharging to the Paria River and sides of the Paria River Canyon. Discharge to the Paria River in the model area, in October 1981 was calculated to be about 3 cubic feet per second (Blanchard, 1984, p. 34), and this is assumed to be representative also of prelake conditions.

Ground-water discharge to Lake Powell in response to seasonal fluctuations of the lake level has been noted by Jacoby and others (1977, p. 61). As lake levels decline after spring runoff, water stored in the Navajo Sandstone discharges to the lake. This discharge, which represents bank storage, is indicated by the decline of water levels in wells close to the shoreline. The distance from the shoreline where this movement can no longer be observed is unknown.

## SIMULATION OF GROUND-WATER FLOW

### General Description of the Model

The two-dimensional, finite-difference, digital-computer model (McDonald and Harbaugh, 1984) used in this study can be employed to simulate the response of a ground-water system to natural or manmade stresses. In order to model a ground-water system, it is necessary to use average conditions, to estimate conditions where sufficient data are not available, and to make simplifying assumptions. Thus, the model is a simplification of the natural system, and results from the model must be viewed with discretion.

In the finite-difference model, a study area is divided into rectangular blocks called cells or nodes in which the aquifer properties are assumed to be uniform. The node at the center of each block is designated by row and column numbers; for example, the node at row 15, column 10, is expressed as 15-10. Using boundary conditions, initial heads, aquifer properties, and sources and sinks, a matrix solver (SIP algorithm in this study) within the model code calculates the values of hydraulic head at the center of each block (block-centered nodes). With these head values, the rate, volume, and direction of ground-water movement through the system can be calculated.

The boundaries used in this study are: constant-head, constant-flux, no-flow, and head-dependent. The altitude of the water surface (head) assigned to a constant-head boundary cell will remain constant throughout the simulation, however, flux (flow) through the cell can change. The flow assigned to a constant-flux cell will remain constant throughout the simulation, however, the head may change. Head-dependent boundaries simulate flow from external stresses, such as flow from a river to the aquifer or flow from the aquifer to some external discharge area, such as a drain.

The head-dependent boundaries used in this study are the river, general-head and drain boundaries. The subroutines, which represent these boundaries, simulate flow through an interface between the aquifer and the external source or discharge area. This interface is described as a prism of porous material for which Darcy's Law applies. Therefore, the flow into or out of the aquifer through a prism of porous material is approximated by the product of the conductance of the prism and the head difference across the prism. The head difference is between the simulated head in the aquifer and an assigned head for the external source or discharge area. The equation for determining conductance is:

$$C=KA/L \tag{1}$$

Where C is conductance, K is hydraulic conductivity of the porous material, A is cross-sectional area perpendicular to the path of flow, and L is distance along the path of flow.

## Model Design and Construction

### Assumptions Used in the Model

Simplifying assumptions used in both steady-state and transient-state simulations are:

1. Ground water moves horizontally in a single-layered medium with no upward or downward movement, therefore a two-dimensional simulation was done.

2. Precipitation directly on the model area does not result in any recharge to the aquifer.

3. Water levels in the aquifer in the steady-state system are static. The water levels available for the modeled area prior to the existence of Lake Powell were not measured frequently enough to show evidence of a seasonal water-level fluctuation. However, other studies of the Navajo Sandstone in southern Utah (Blanchard, 1984 and Cordova, 1981) have shown that seasonal fluctuations of water levels are small.

4. Evapotranspiration from the aquifer is insignificant.

5. Well discharges have an insignificant effect on water levels in the aquifer.

6. The aquifer acts as an isotropic, porous medium. This assumption is not strictly true, because the Navajo Sandstone is greatly fractured throughout the area; and water moves more readily through these cracks than through the interstitial pores in the sandstone. Despite this, the model is still considered to be valid because it covers a large area (approximately 600 square miles), and on this scale the aquifer probably acts as an isotropic, porous medium. Also, over a long period of time, ground-water flow through fractured rocks probably is similar to flow through an isotropic, porous medium.

### Grid Network

The two-dimensional, block-centered grid used to model the aquifer is shown in figure 7. The grid is oriented on a northwest-southeast axis because this is the principal direction of ground-water movement, and it is perpendicular to the channel of the Colorado River in the area. The rectangular grid has 35 rows and 30 columns, with the cells ranging from 0.3 to 2.0 miles on a side. The number of active nodes is 801. The cells are small near the Colorado River and Lake Powell where steep water-level gradients are created by infiltration from the lake and where the most information is available. The cells are large at the north and west edges of the model where gradients are less steep and information is meager.

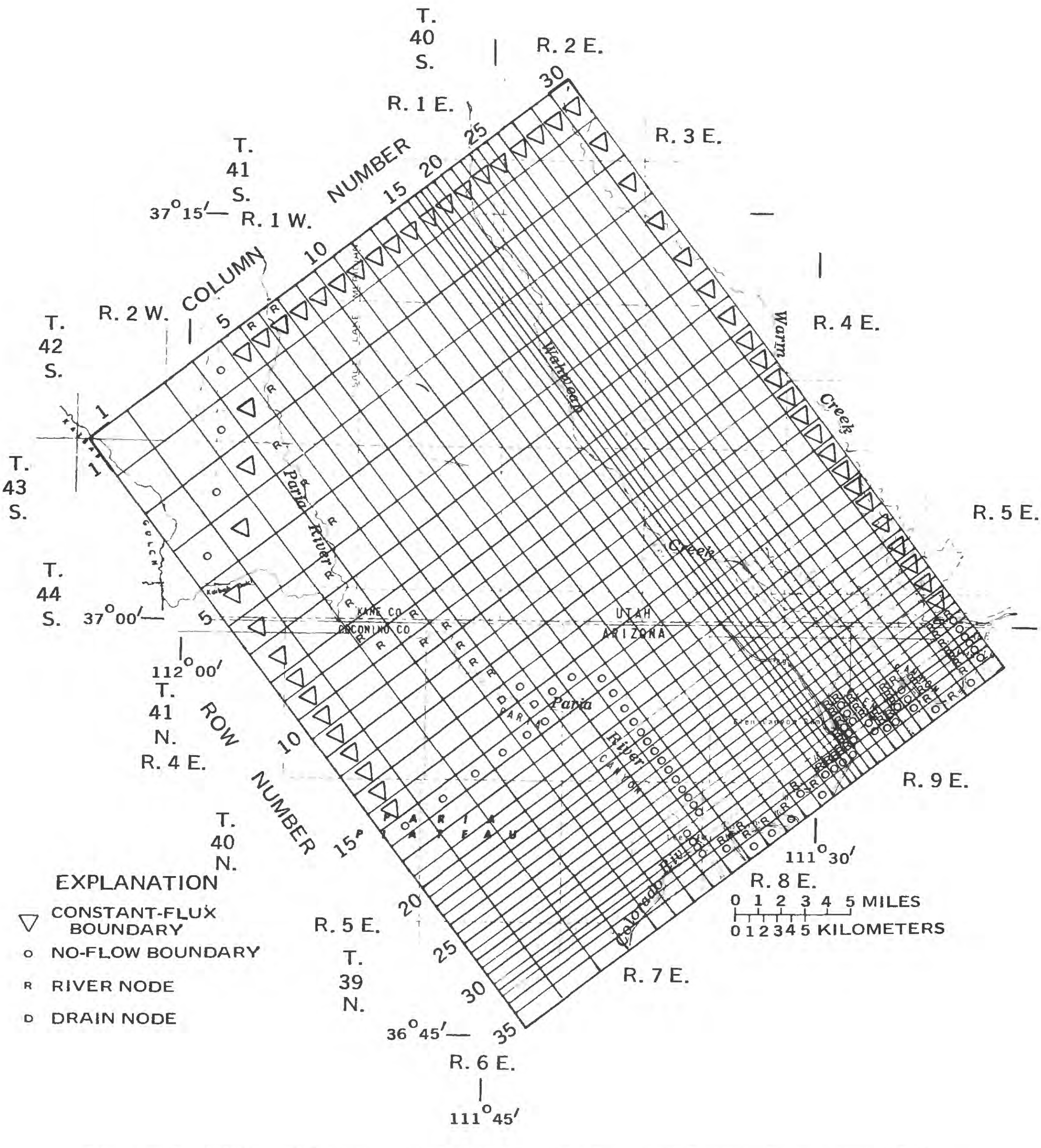



Figure 7.—Finite-difference grid and boundaries used in steady-state model.

## Boundary Conditions

The location of the boundaries used in the model is based on the distribution of water-level data and the conceptual ground-water flow pattern. Most of the data in the study area are in the Wahweap Bay area or the Paria Plateau (fig. 6). Some water-level data are available northeast of the model area near the town of Boulder, however, water in the Navajo Sandstone in the Boulder area moves toward the south and discharges to the Escalante River or eventually the Colorado River or Lake Powell upstream from the confluence with Warm Creek. Thus, these water levels are not relevant to the Wahweap Bay area.

The upper boundary of the system was simulated as either unconfined or confined. A no-flow boundary was used at the top of the Navajo in the confined areas, implying that there is no upward or downward movement through the overlying Carmel Formation. A no-flow boundary also was used at the base of the Navajo, implying no vertical movement through the underlying Kayenta Formation. The assumption of no downward vertical flow is considered justified for the model area, because the recharge areas, where downward vertical flows are likely, are outside the model area. Upward vertical flow from the Moenave Formation may occur at the Colorado River in pre-lake conditions, but, this component is not considered significant enough to affect the water levels in the Navajo used for steady-state calibration.

The steady-state boundaries used to simulate the system are shown in figure 7. Constant-flux boundaries were used on the northwest and southwest sides to simulate subsurface inflow from outside the modeled area. The Navajo Sandstone crops out at the East Kaibab monocline at the west corner of the model (fig. 3), and constant-flux nodes are used to simulate recharge on this outcrop. The northeast boundary is approximently parallel to the direction of ground-water flow and perpendicular to the channel of the Colorado River (fig. 6). However, the boundary was simulated as constant flux, with the assumption that the flow in or out of the boundary would be small and most flow would be parallel to the boundary.

South of row 15 on the Paria Plateau and at the southwest corner of the model, water levels are beneath the Navajo Sandstone and a no-flow boundary was used along row 16. South of row 15 and along column 9, a no-flow boundary was used at the outcrop of older rocks. A no-flow boundary also was used on the southeast boundary, becaused it was assumed that all water in the Navajo is discharged into the Colorado River before reaching the southeast boundary.

The transient-state boundaries used to simulate the system are shown in figure 8. The only change from the steady-state boundaries is the Colorado River above Glen Canyon Dam changes from a river and discharge boundary to a lake and recharge boundary. Lake Powell is defined as an internal part of the system, and the method of simulating the lake is described in the following section on Recharge. The lake is assummed to fully penetrate the system, thus the no-flow boundary on the southeast side of the model that was used for the steady-state simulation was not changed.

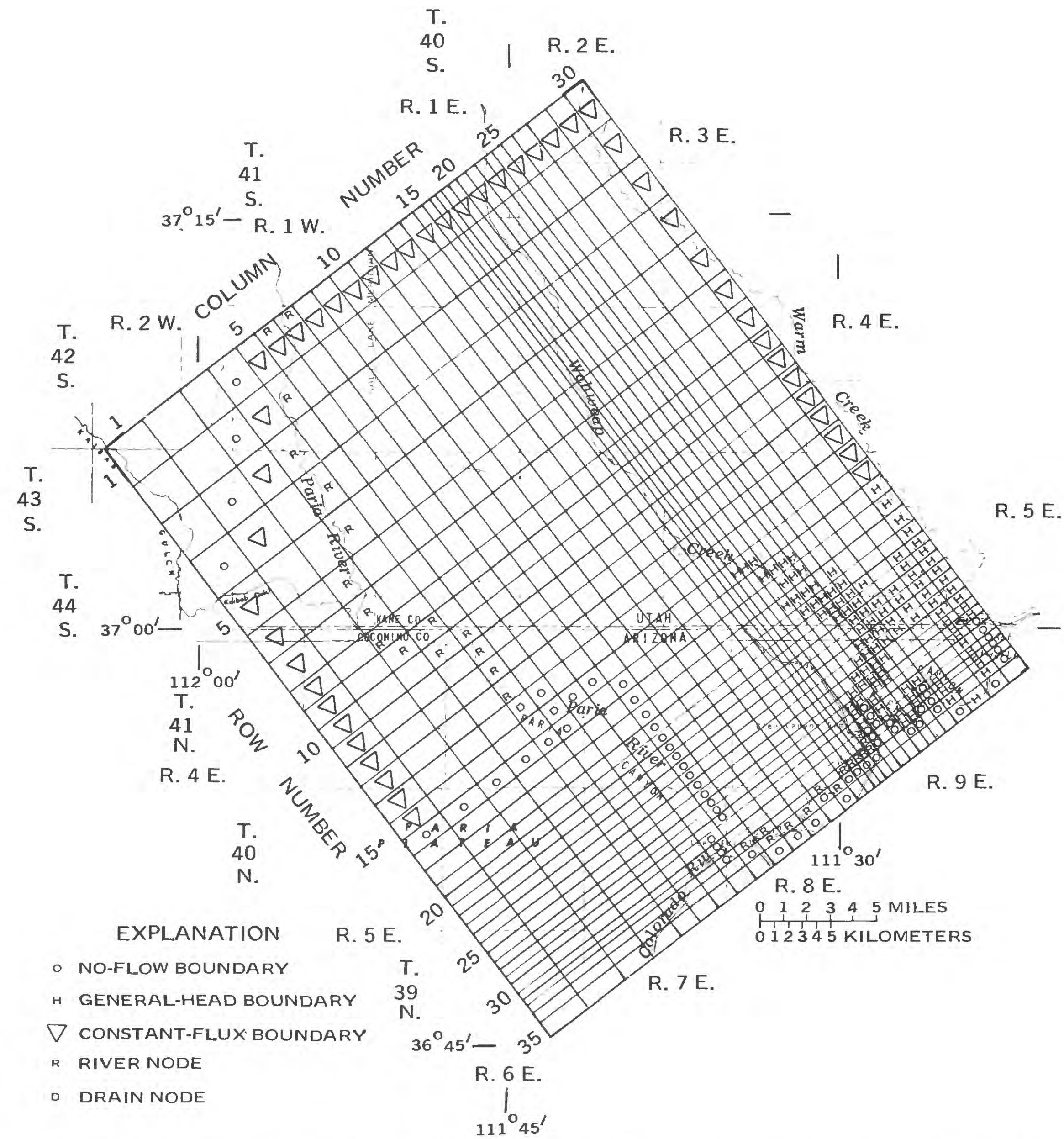



Figure 8.—Finite-difference grid and boundaries used in the transient-state model, 1983.

A major assumption during the transient-state simulation was that the constant-flux boundaries on the southwest, northwest, and northeast sides of the model were not affected by Lake Powell. This probably is a reasonable assumption for the southwest and northwest boundaries because they were more than 20 miles from the shoreline in 1983. The northeast boundary is approximately parallel to the direction of ground-water flow during steady-state conditions. The effects of Lake Powell on this steady-state flow pattern are uncertain, because no water-level data are available near this boundary. East of Wahweap Bay and north of Lake Powell, it is assumed that water-level changes caused by the filling of Lake Powell would be approximately the same at equal distances from the lake; thus, water-level gradients and quantity of flow across the northeast boundary should not change significantly. The small quantity of flow across the boundary that was determined from steady-state calibration was therefore kept at a constant rate for the transient-state simulations. This allows the heads on this boundary and flow to or from the lake to change in response to lake-level fluctuations. Sensitivity of the model to this boundary is presented in the section entitled "Sensitivity of model boundaries".

## Data Input

### Water Levels

The water levels that were used in the calibration of the model are listed in table 2. Water levels in five wells in the model area were measured during 1957-59. These water levels, some water levels measured during 1970-81 on the Paria Plateau just outside the southwest boundary of the modeled area, and the potentiometric map in figure 6 were used to estimate the steady-state water level at each node in the model. Initial water levels for the transient-state simulations were determined from the calibrated steady-state model.

### Recharge

A range of recharge of 5,000 to 15,000 acre-feet per year was simulated using constant-flux nodes. Recharge by seepage from the Paria River in the north part of the modeled area (rows 1-3) was simulated as a head-dependent river boundary. The amount of leakage was determined by the conductance of the streambed, the stage of the river, and the simulated head for the aquifer in the stream node.

Recharge from Lake Powell in the transient-state period was simulated by the general-head-boundary subroutine (GHB). As the lake rose throughout the transient-state period, the nodes that were inundated by the lake were changed to GHB nodes. The quantity of recharge entering the aquifer was determined by the conductance of the GHB node, the assigned head for the node (lake level), and the simulated aquifer head.

The transient-state period for the model was from March 1, 1963, to March 31, 1983. During this period, Lake Powell near Glen Canyon Dam rose about 530 feet. The 20-year period was divided into 22 stress periods, with the first two periods of 2 months length, the third of 9 months, and the remaining stress periods of 1 year. The number of time steps in each stress period ranged from 5 to 17. The first two stress periods are short because the lake rose 175 feet in 4 months. The rise was more gradual after that. A lake level, which was assigned to each stress period, was input as the head for all GHB nodes. The lake level used was a weighted average based on monthly levels observed during the stress period. For example, the lake level for stress period 5 (April 1, 1965, to March 31, 1966) was computed by adding 13 end-of-month water levels from March 31, 1965, to March 31, 1966, and dividing by 13.

The conductance of the GHB nodes simulating the lake was determined by equation (1), $C=KA/L$. Conductance in this simulation was defined as the vertical area of the cell, which is in contact with the lake, times a coefficient. Thus, each cell differs according to the area, and the coefficient was kept constant for all cells. The coefficient is equal to $K/L$ in equation (1); and for an initial estimate it was given a value of 1.0 based on values of 1.0 feet per day for K and 1.0 feet for L.

The area of each cell in contact with Lake Powell was determined by multiplying the height of the canyon inundated by water times the effective length of the shoreline in the cell. The height was the water level of the lake, for that stress period, minus the average low altitude of the cell. The average low altitude was the stream-channel altitude, or if there was no channel in the cell, it was the average low-altitude contour. The effective length of shoreline was calculated two different ways depending on whether water moving from the lake to the aquifer went through one or two sides of a cell. For cells with one side contributing water, the length of shoreline was computed as follows: the average shoreline contour for each stress period was drawn on topographic maps, and the length of the shoreline parallel to the cell side was measured with a straight edge. Thus, the sinuous path of a shoreline was averaged into an "effective" length that was more appropriate to the average conditions used in a finite-difference model. For a cell that had water movement through two sides, the length used was the arithmetic average of both shorelines.

## Hydraulic Properties of the Aquifer

The initial estimate of areal distribution of hydraulic conductivity was varied across the model according to the depth of burial of the Navajo Sandstone. In the west and south parts of the area, the Navajo crops out, and the hydraulic conductivity was assigned a value of 1.5 feet per day. The Navajo dips toward the north; and at the north corner of the modeled area, where the aquifer is approximately 3,000 feet below land surface, the hydraulic conductivity was assigned a value of 1.0 feet per day. The altitudes of the top and bottom of the Navajo were input, and the model computed transmissivity from saturated thickness times hydraulic conductivity. The initial values input for storage coefficient and specific yield were 0.0001 and 0.08.

## Discharge

Ground water discharges to the Paria River, Paria Canyon (where the bottom of the Navajo Sandstone crops out above the canyon bottom), the Colorado River, Lake Powell, and across the constant-flux boundary at the northeast part of the modeled area. The Paria River was modeled as a partially penetrating stream, which assumes that the stream intersects only part of the saturated aquifer. Thus, water can move through the aquifer beneath the stream without discharging into the stream. Therefore, the effects in the aquifer of a stress on one side of the stream can be observed on the other side of the stream. The stage of the Paria River was assumed to be 2 feet above the streambed, and conductance for the Paria River nodes was determined during calibration.

The Colorado River was modeled as a fully penetrating stream by placing a no-flow boundary on the south side, thus causing all ground water to discharge into the stream. The stage of the Colorado River was assumed to be 10 feet above the streambed. The initial estimates of conductance for the Colorado River nodes were calculated from the area of the river times an average conductivity for the Navajo Sandstone (1.5 feet per day) divided by the average thickness of river sediments (70 feet) (Phoenix, D. A., U.S. Geological Survey, written commun., 1962). Discharge at two nodes (14-5 and 15-6) at the contact of the Navajo Sandstone and the Kayenta Formation on the west side of the Paria Canyon was simulated using the drain subroutine. Conductance for the drain nodes was assigned a value of 1.0 square foot per second, and the head for each drain was determined from the steady-state water-level surface.

The discharge into Lake Powell in response to decreases in the lake level is simulated by the GHB. Seasonal fluctuations are not simulated because the stress periods are 1 year long. However, for 1967-68, 1972-73, 1976-79, and 1981-82 the average lake level used in the model decreased, and this caused some ground-water discharge into the lake.

## Calibration Procedure

The calibration of a ground-water model is a trial-and-error procedure wherein aquifer properties and boundary conditions are varied to achieve an acceptable match between measured (historical) water levels and simulated water levels. Comparisons also are made between simulated and measured streamflow (gains or losses), head gradient, known discharge areas, or any relatively well known aspect of the ground-water system. The aquifer properties and boundary conditions are varied within previously determined limits, which are based on knowledge of the geologic and hydrologic characteristics of the system. The limits for this model were given earlier in the section entitled "Ground-Water System." The measured water levels used in the calibration process are from wells that are in specific nodes of the model grid (table 2). Hereafter, in this report, measured water levels are referred to according to their node location.

## Steady-State Calibration

A steady-state condition exists when there are no changes in storage or head in the aquifer with time. This condition was approximated prior to March 1963, before the filling of Lake Powell. Water levels measured during 1957-59 in five wells within the model area were used for calibration (table 2). Water levels measured during 1970-81 at a few wells in the Paria Plateau just outside the modeled area also were used to estimate the altitude of the potentiometric surface at the southwest boundary of the model. It was assumed that these water levels were far enough from the lake to approximate the steady-state condition. The water-level altitude at node 15-1 was estimated to be about 4,900 feet by extrapolating a water-level contour from the Paria Plateau. A water level measured in 1971 at node 3-4 also was used for reference in the steady-state calibration, under the assumption that the water level at this node did not change much from 1963 to 1971.

The model was calibrated with a range of 5,000 to 15,000 acre-feet per year for recharge and 0.2 to 10.0 feet per day for hydraulic conductivity because of the small amount of available data. The range for recharge is considered to be more accurate than the range for hydraulic conductivity. Therefore, the steady-state model was calibrated for three recharge options, about equally spaced between 5,000 and 15,000 acre-feet per year. The objective was to determine the magnitude and distribution of hydraulic conductivity that provides an acceptable calibration to the measured water levels and conceptual flow pattern.

The following properties and boundary conditions of the Navajo Sandstone were varied as part of the steady-state calibration: hydraulic conductivity, conductance for the Paria River and Colorado River nodes, and the flows at the three constant-flux boundaries (southwest, northwest, and northeast sides). Initially, the calibration was made with three constant-head boundaries in order to match the measured water levels and to obtain a general idea of the distribution of head and hydraulic conductivity. These boundaries were then converted to constant flux, and the three recharge options were calibrated by adjusting the flow through the boundaries and the hydraulic conductivity of the aquifer. Conductance values for the Paria River were adjusted with the middle recharge option of about 10,000 acre-feet per year to match a seepage run on the river which had shown a gain of 3 cubic feet per second. Conductance values for the Colorado River were adjusted to match measured water levels in the Navajo Sandstone within a few miles of the river.

## Transient-State Calibration

The period used for transient-state calibration was March 1, 1963, to March 31, 1983. Monthly water-level measurements that began during June 1964 were available for several wells near the shore of Lake Powell. These wells are all within 1 square mile of each other, and the measurements all are incorporated in node 23-24. The nodes of four other wells used in the transient-state calibration (table 2) and their distance from the shoreline in 1983 are: node 6-10 (12 miles), node 19-16 (2 miles), node 22-19 (2 miles), and node 26-19 (1 mi). Two observation wells operated by the Bureau of Reclamation (well (A-41-8)23dcd-1 and well (A-41-8)23dac-1) are less than 1 mile west of Glen Canyon Dam. Water levels measured from 1967-83 in these wells were not used because the grid size chosen for this study is too large to simulate the large water-level gradient around the dam, which is reflected in water levels in these wells. Also, the transient-state simulation of the river just below the dam probably is not an accurate representation of a complex situation which might include seepage faces and heads at the river nodes that may change with time.

The following properties and boundary conditions of the Navajo Sandstone were varied during the transient-state calibration: hydraulic conductivity, storage coefficient, specific yield, and the conductance for the general-head boundary that simulates Lake Powell. Recharge was kept constant through the constant-flux boundaries on the southwest and northwest sides of the area. The flows through the northeast boundary that were determined from steady-state calibration were kept constant for transient-state simulations. The lake GHB was changed each stress period by putting in the average lake level as the head at each GHB node. Also, the value for area used in equation (1) was increased as the area of the aquifer in contact with the lake increased. The area was determined independently and was not adjusted as part of the calibration. The only term in equation (1) that was varied was K/L.

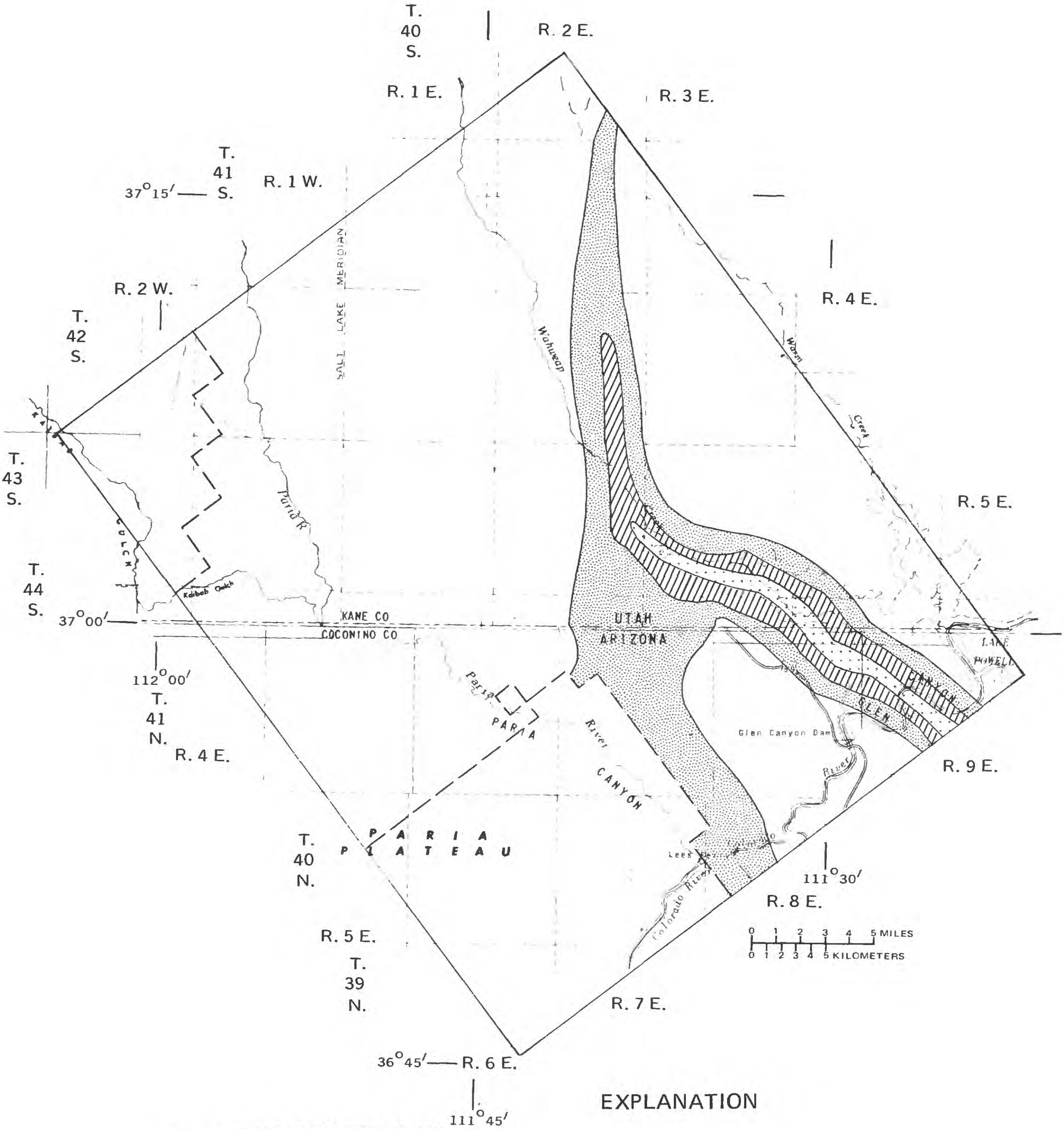
## SIMULATION RESULTS

No single set of aquifer properties and boundary-flow quantities can be presented that would constitute a reliable representation of the system because of the small quantity of available data and the consequent uncertainty of the estimates of steady-state recharge, hydraulic conductivity and storage properties of the Navajo Sandstone, and recharge from Lake Powell. Therefore, a range of values for subsurface recharge, hydraulic conductivity, specific yield, and conductance for the GHB are presented. The evaluation of the various simulations is given in the section entitled "Discussion".

### Results of Model Calibration

Three rates of recharge of 5,720, 10,440, and 14,820 acre-feet per year were simulated using a range of subsurface inflow of 5,000 to 15,000 acre-feet per year. These recharge rates corresponded to hydraulic conductivities of 0.25 to 1.12 feet per day, 0.5 to 2.25 feet per day, and 0.75 to 3.38 feet per day. The initial distribution of hydraulic conductivity (1.0 to 1.5 feet per day) provided simulated water levels that were comparable to the measured water levels. After comparing with transient-state water levels, however, it was necessary to put in a zone of larger conductivity along the Wahweap syncline (4.5 feet per day at the axis) for better simulation of the large rises shown in water levels measured a few miles from the lake. This distribution, shown in figure 9, was determined from simulation with constant-head boundaries.

Flow was computed through the constant-head boundaries using the new distribution of hydraulic conductivity (1.0 to 4.5 feet per day), and the boundaries were converted to constant flux for the remainding simulations. The flow through constant-flux nodes was then redistributed because the constant-head flow resulted in unreasonably large flows (17-53 percent of mean annual precipitation) at the outcrop area at the East Kaibab monocline. The redistributed inflow (12-39 percent of mean annual precipitation at the outcrop area) more accurately depicts recharge conditions. The excess flow from the outcrop area was spread evenly across the two adjacent boundaries.




## EXPLANATION

— — — NO-FLOW BOUNDARY

VALUES OF HYDRAULIC CONDUCTIVITY, IN FEET PER DAY, FOR THREE RECHARGE OPTIONS

| | 5,720 acre-feet per year | 10,440 acre-feet per year | 14,820 acre-feet per year |
|---|---|---|---|
| | 1.0 - 1.12 | 2.0 - 2.25 | 3.0 - 3.38 |
| | 0.75 - 1.0 | 1.5 - 2.0 | 2.25 - 3.0 |
| | 0.5 - 0.75 | 1.0 - 1.5 | 1.5 - 2.25 |
| | 0.25 - 0.5 | 0.5 - 1.0 | 0.75 - 1.5 |

Figure 9.—Distribution of hydraulic conductivity determined from calibration of the transient-state model.

This redistributed flow had no effect on the measured water levels used for comparisons. For the three options of recharge, the hydraulic conductivity was changed by multiplying a constant factor times the entire array, thus keeping constant the ratio for hydraulic conductivity of 1.0 to 4.5 that was determined by simulation with constant-head boundaries.

Nodes for the Paria River were calibrated to a gain of 3 cubic feet per second, which resulted in a conductance of 0.001 square foot per second for each node. Nodes for the Colorado River were calibrated to measured water levels in the Navajo Sandstone, and the resulting values for conductance were found to be about 70 percent of the initial values. These final conductances ranged from 0.07 to 0.34 square foot per second.

Simulated water levels for the three recharge rates are compared to measured water levels in table 3. All three rates resulted in a fairly close match to the measured water levels. During calibration, the water-level match attempted at node 6-10 was to a water level of 4,020 feet instead of the measured water level of 4,035 feet. The 4,020 feet water level is an adjustment for the one-half mile distance between the well and the center of the cell. The potentiometric surface for the recharge of 10,440 acre-feet per year is shown in figure 10. This surface is similar for all three recharge rates, and the water budgets for the three rates are shown in table 4.

Transient-state models often are used to confirm or substantiate the values for recharge and hydraulic conductivity that are used in the steady-state model. In this case, the uncertainty of the steady-state representation and the small amount of data precluded a traditional analysis. Instead, the analysis of the transient-state system was made with a range of input data. Using reasonable values of input data, it was found that measured water levels at node 23-24 (0.3 mile from the 1983 shoreline of Lake Powell) could not be matched concurrently with the measured water levels in nodes more than 1 mile from the lake. If a reasonable match was achieved for node 23-24, the simulated water levels for 1981 at nodes 22-19 and 26-19 were about 100 feet too low. Using the three recharge options, therefore, the specific yield and the conductance value for the lake (GHB) were varied to match two situations: matching water levels at node 23-24 and matching water-level changes at nodes 6-10, 19-16, and 26-19. A range of values of specific yield and conductance (GHB) thus is presented. The actual values probably fall somewhere within this range.

The criteria used to define a reasonable match are somewhat arbitrary; but they include, in part, the proximity of a well to the center of the cell, the reliability of measured water levels, and the magnitude of water-level changes. Measured water-level changes were used for the nodes more than 1 mile from the lake. For node 6-10, the measured and simulated changes had to be within 20 percent of each other. For nodes 19-16 and 26-19, measured and simulated changes had to be within 10 percent of each other. The changes used are from steady-state conditions to 1981 for node 6-10, from 1972 to 1981 for node 19-16, and from steady-state conditions to 1983 for node 26-19 (table 2). The criterion used for a match with the water levels at node 23-24 is that the difference between measured and simulated water levels (residual) in each stress period must be less than 15 feet.

Table 3.--Simulated and measured water levels for the steady-state model

Node: Row-column
Simulated subsurface recharge: Simulations were made with three recharge rates.

| | Measured | | Simulated | |
|---|---|---|---|---|
| Node | Date | Water-level altitude (feet) | Subsurface recharge (acre-feet per year) | Water-level altitude (feet) |
| 3-4 | August 1971[1] | 4,300 | 5,720 | 4,300 |
| | | | 10,440 | 4,294 |
| | | | 14,820 | 4,292 |
| 6-10 | September 1957 | 4,035 | 5,720 | 4,020 |
| | | | 10,440 | 4,025 |
| | | | 14,820 | 4,024 |
| 23-24 | January 1959 | 3,313 | 5,720 | 3,309 |
| | | | 10,440 | 3,312 |
| | | | 14,820 | 3,313 |
| 26-19 | October 1957 | 3,240 | 5,720 | 3,247 |
| | | | 10,440 | 3,250 |
| | | | 14,820 | 3,252 |
| 30-17 | July 1957 | 3,182 | 5,720 | 3,180 |
| | | | 10,440 | 3,181 |
| | | | 14,820 | 3,182 |
| 30-18 | January 1958 | 3,182 | 5,720 | 3,173 |
| | | | 10,440 | 3,174 |
| | | | 14,820 | 3,176 |

[1] The water level for this node is not in the steady-state period-prior to March 1963. This node, however, is about 20 miles from the 1971 lake shoreline, and the measured water level probably is close to that during steady-state conditions.

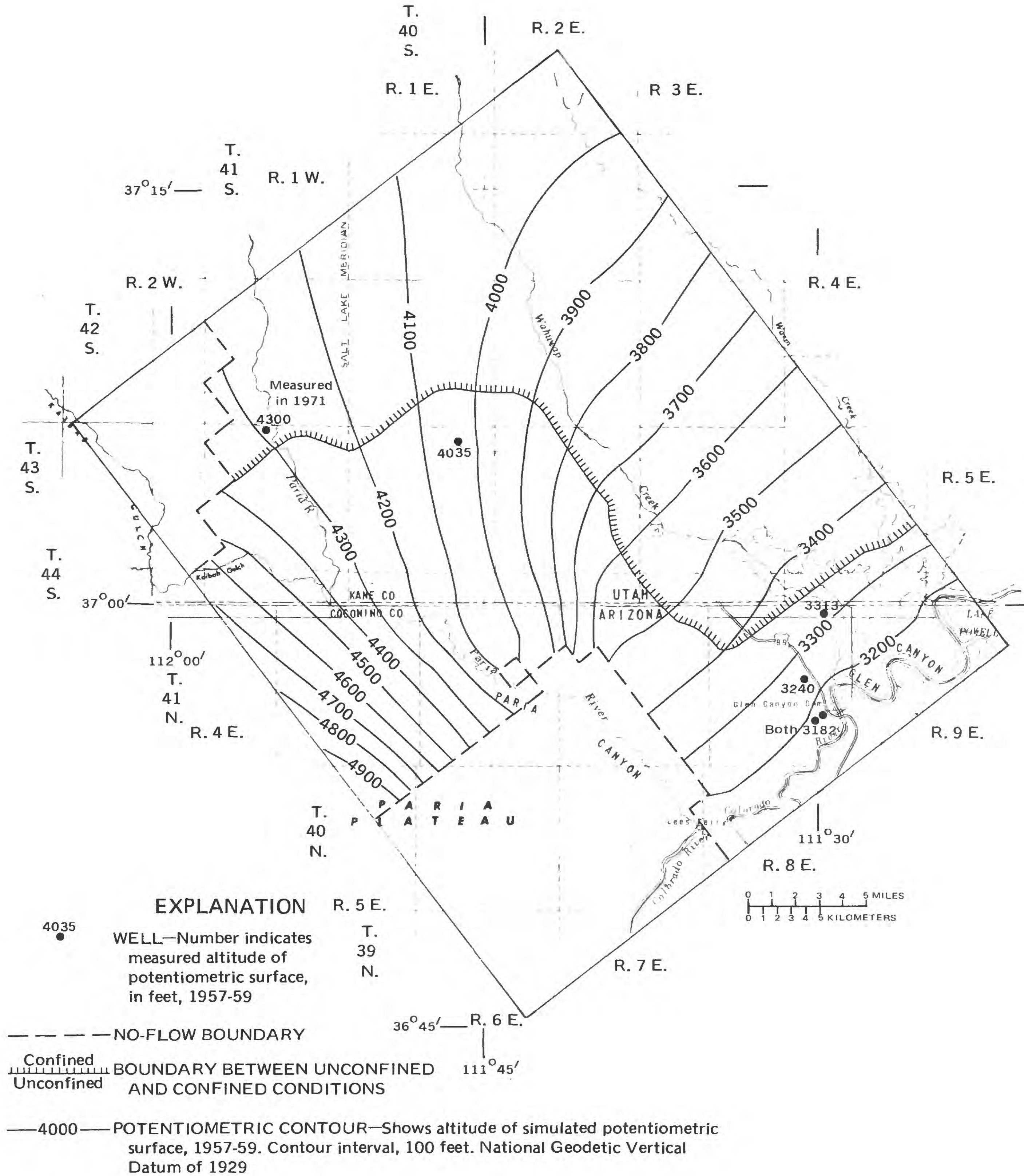



Figure 10.—Potentiometric surface for steady-state simulation using recharge of 10,440 acre-feet per year.

Table 4.--Simulated water budget for steady-state conditions for three recharge options.

[Values for budget components, in acre-feet per year, are not intended to imply accuracy to the precision shown. See figure 7 for location of model boundaries]

| Budget component | Recharge option (acre-feet per year) | | |
|---|---|---|---|
| | 5,720 | 10,440 | 14,820 |
| Inflow | | | |
| Subsurface recharge | | | |
| Southwest boundary (rows 5-15, col. 1) | 3,040 | 5,710 | 7,950 |
| Outcrop area boundary (nodes 2-4, 3-3, 4-2) | 690 | 1,380 | 2,240 |
| Northwest boundary (row 1, columns 5-30) | 1,600 | 2,610 | 3,530 |
| Northeast boundary (rows 18-27, col. 30) | 390 | 740 | 1,100 |
| From Paria River (rows 1-4) | 10 | 10 | 10 |
| Total inflow | 5,730 | 10,450 | 14,830 |
| Outflow | | | |
| Subsurface discharge | | | |
| Northeast boundary (rows 2-17, col. 30) | 350 | 780 | 1,220 |
| To Paria Canyon (nodes 14-5, 15-6) | 20 | 250 | 480 |
| To Paria River | 1,740 | 2,190 | 2,330 |
| To Colorado River | 3,620 | 7,230 | 10,800 |
| Total outflow | 5,730 | 10,450 | 14,830 |

Using the above criteria, specific yield was varied between 0.02 and 0.15, and the conductance coefficient (GHB) for the lake (see page 23) was varied between 0.1 and 1.0 until a match was achieved. Table 5 shows the specific yield and conductance coefficient (GHB) needed to meet the match criteria for the three recharge options and the two sets of measured water levels (node 23-24, 0.3 mile from shoreline, and nodes 1 to 12 miles from the shoreline). Calibration did not result in a decrease in the conceptual range of specific yield of 0.02 to 0.15. When water levels at node 23-24 are matched, the simulated rises at node 6-10 range from 0 to 7 percent of the measured levels; simulated rises at node 19-16 range from 97 to 100 percent of the measured levels; and simulated rises at node 26-19 range from 71 to 74 percent of the measured levels. When measured rises at nodes 6-10, 19-16, and 26-19 are matched, the maximum difference between measured and simulated water levels at node 23-24 is almost plus 50 feet for all the specific yield and conductance (GHB) options. The model was not sensitive to storage coefficient in the range of 0.001 to 0.000001. When storage coefficient was varied from 0.001 to 0.0001, the maximum head change at any node, during 1983, was 11 feet. When storage coefficient was varied from 0.0001 to 0.000001, the maximum head change was 1 foot. Consequently storage coefficient was set at 0.0001 for all options used here.

The simulated change in storage from steady-state conditions to March 1983 is shown in table 5. The change in storage is given in acre-feet per mile; and it can be converted to acre-feet for the entire model area by multiplying by 35, which is the approximate number of miles of lake shoreline included in the model area. The increase in discharge to the 9-mile reach of the Colorado River below Glen Canyon Dam from steady-state conditions to 1983 was calculated for the nine calibration options, and it ranged from 140 to 170 percent of the discharge during steady-state conditions.

## Simulated Effects of Lake Powell

A match can be made for all calibration nodes with the recharge option of 10,440 acre-feet per year, specific yield equal to 0.08, and conductance coefficient (GHB) equal to 0.1. With this set of input data, the maximum residual for node 23-24 is plus 37 feet, the simulated water-level rise at node 6-10 is 20 percent of the measured rise, the simulated rise at node 19-16 is 105 percent of the measured rise, and the simulated rise at node 26-19 is 86 percent of the measured rise. The water-level changes from March 1963 to March 1965 for this set of input data are shown in figure 11. The water-level changes have moved up the canyons of Wahweap and Warm Creeks.

The water-level changes after 20 years of recharge from Lake Powell are shown in figure 12, and the simulated potentiometric surface for March 1983 is shown in figure 13. Comparison of this surface with the steady-state surface in figure 10, shows a reduced gradient north of the lake in 1983 and a steep gradient at Glen Canyon Dam. The northeast trending part of the boundary between unconfined and confined conditions (shown in figure 10 near row 20 of the model grid) moved about 2 miles south in the simulated potentiometric surface shown in figure 13. This simulated change from unconfined to confined conditions covers about 10 square miles.

Table 5.--Results of transient-state calibration and simulated changes in storage

Calibration option: The transient-state model was calibrated for two options: (1) matching water-level changes at nodes 6-10, 19-16, and 26-19, and (2) matching water levels at node 23-24.
Conductance coefficient (GHB): Equivalent to K/L in equation (1), C=KA/L.
Increase in storage: For 1971-83, the model included about 35 miles of lake shoreline.

| Recharge (acre-feet per year) | Calibration option | Conductance coefficient (GHB) | Specific yield | Increase in storage (1963-83) (acre-feet per mile) |
|---|---|---|---|---|
| 5,720 | nodes 6-10, 19-16, 26-19 | 0.1 | 0.02 | 7,000 |
| | | 1.0 | .02 | 7,000 |
| | node 23-24 | .1 | .10 | 28,000 |
| | | 1.0 | .11 | 30,000 |
| 10,440 | nodes 6-10, 19-16, 26-19 | .1 | .03 | 11,000 |
| | | 1.0 | .03 | 11,000 |
| | node 23-24 | .1 | [1]>.15 | -- |
| | | 1.0 | >.15 | -- |
| | match for all calibration nodes[2] | .1 | .08 | 25,000 |
| 14,820 | nodes 6-10 19-16, 26-19 | .1 | .05 | 18,000 |
| | | 1.0 | .05 | 18,000 |
| | node 23-24 | .1 | >.15 | -- |
| | | 1.0 | >.15 | -- |

1 To match the water levels here, it is necessary to have the specific yield exceed 0.15, which is considered to be the largest reasonable value.
2 See section on "Simulated effects of Lake Powell".

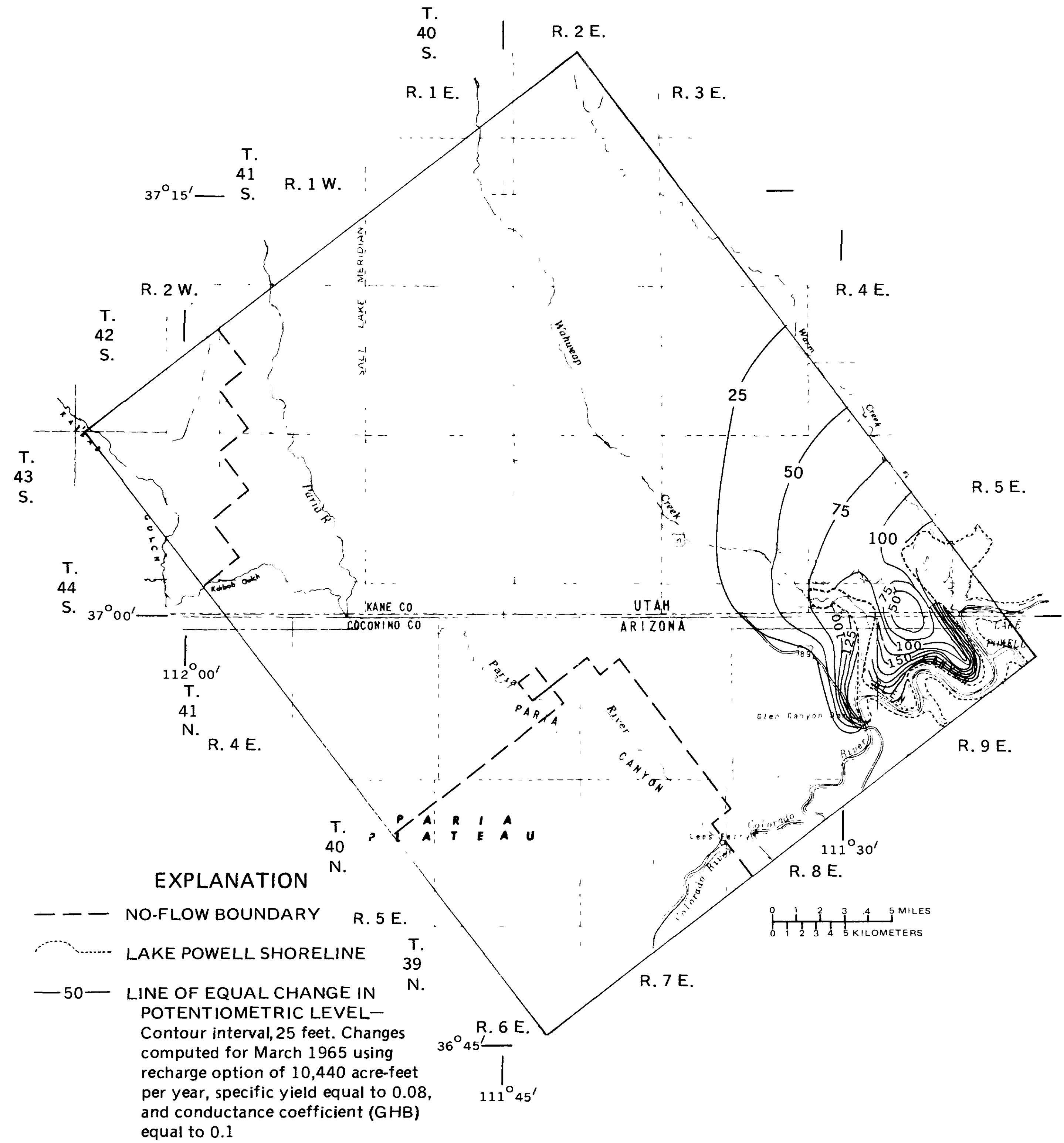



Figure 11.—Simulated changes in the potentiometric surface caused by 2 years of stress imposed on the system from Lake Powell.

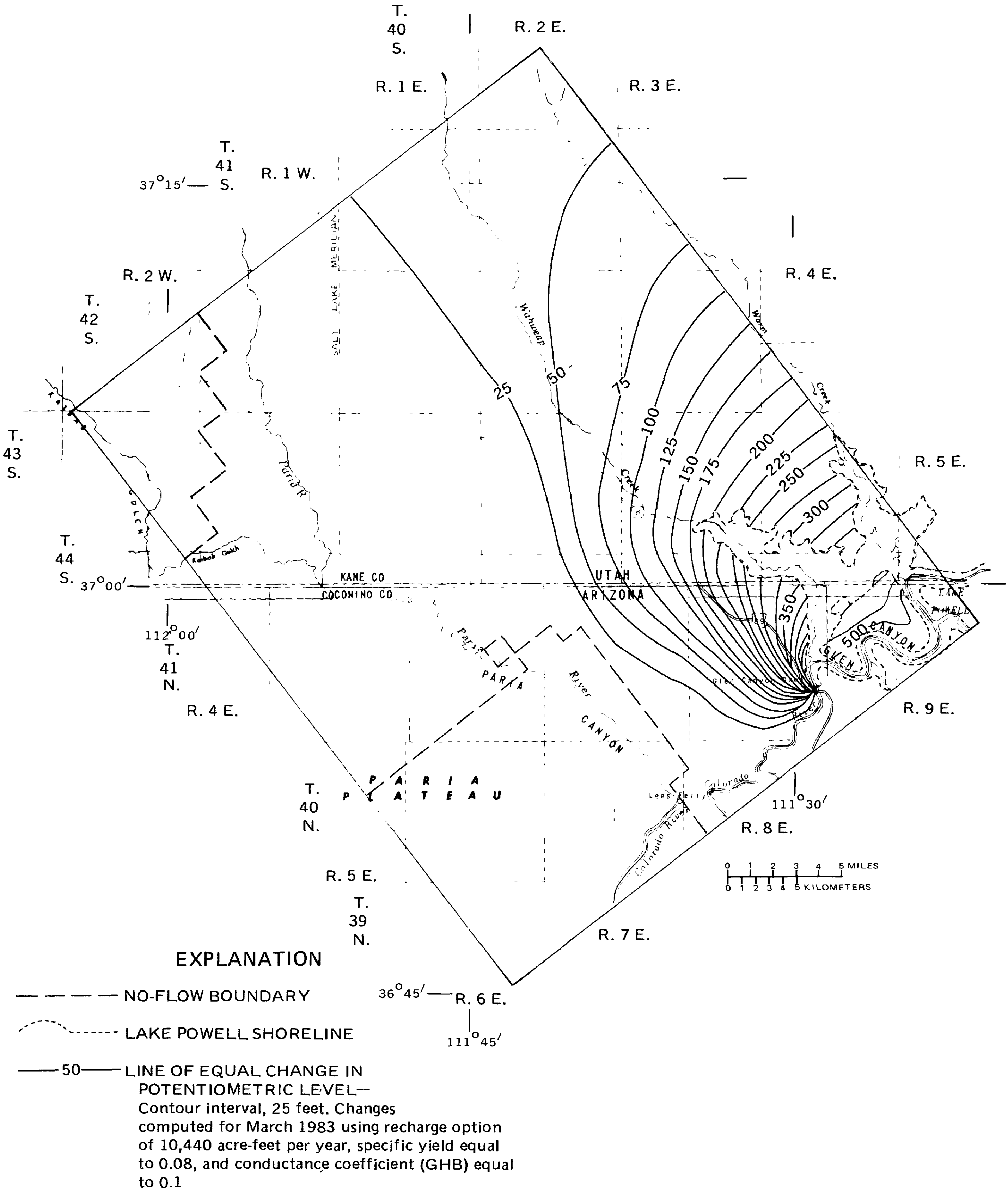
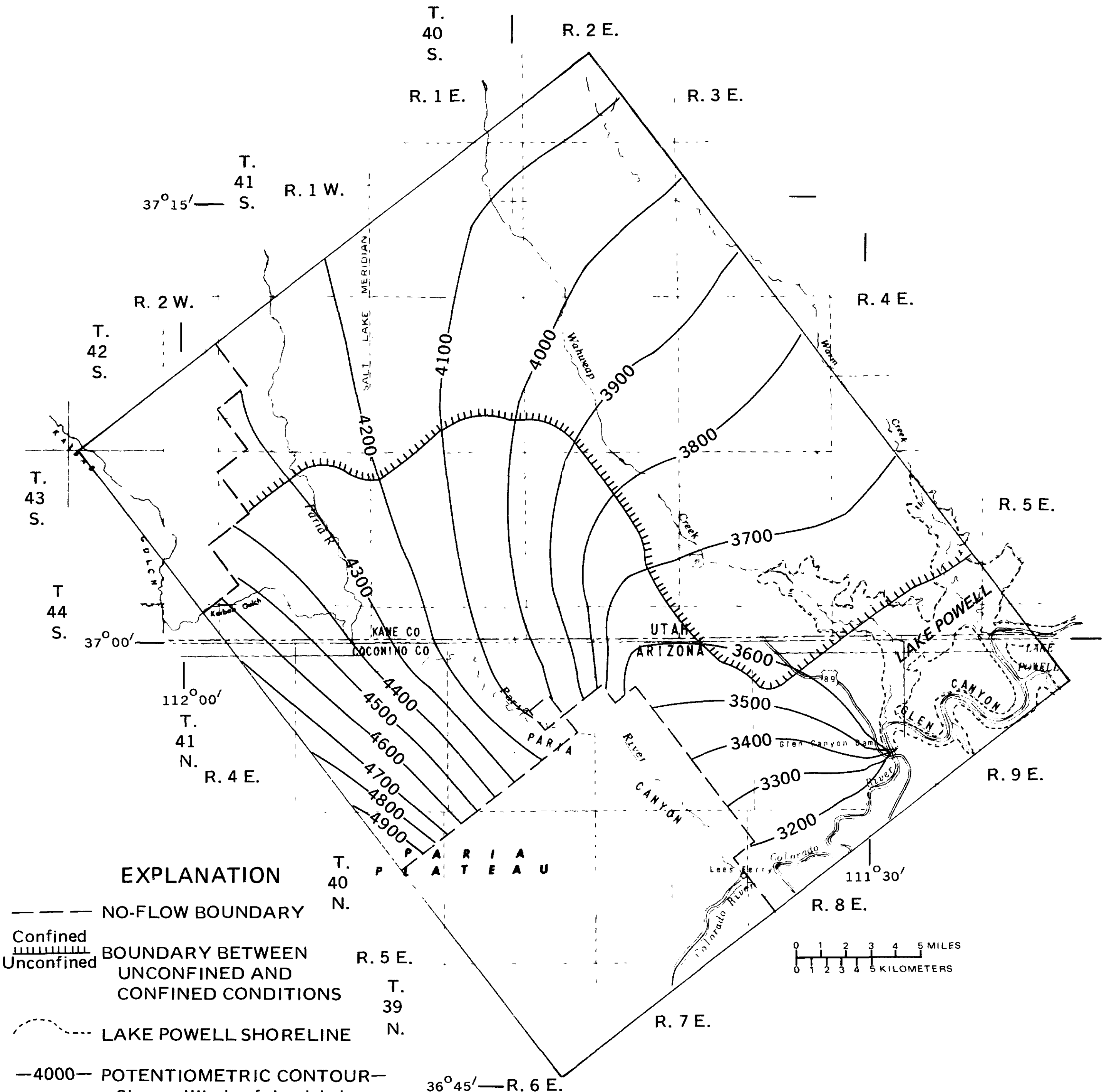



Figure 12.—Simulated changes in the potentiometric surface caused by 20 years of stress imposed on the system from Lake Powell.




EXPLANATION

— — — NO-FLOW BOUNDARY

Confined / Unconfined — BOUNDARY BETWEEN UNCONFINED AND CONFINED CONDITIONS

LAKE POWELL SHORELINE

—4000— POTENTIOMETRIC CONTOUR—Shows altitude of simulated potentiometric surface, March 1983. Contour interval, 100 feet. National Geodetic Vertical Datum of 1929. Surface was generated using recharge option of 10,440 acre-feet per year, specific yield equal to 0.08, and conductance coefficient (GHB) equal to 0.1

Figure 13.—Simulated potentiometric surface, March 1983.

A simulation was made to determine how long the system would take to reach equilibrium at a constant lake level. A major assumption was that the conductance between the lake and the aquifer did not change during the simulation period. Also, this projection of water levels is for an estimate of changes in the regional ground-water system, and it is not meant to simulate conditions accurately near Glen Canyon Dam. The 1983 potentiometric surface was used as the starting head surface, and the model was run to simulate 1,500 years, with 100 time steps, at a constant lake level of 3,680 feet (the approximate level during 1981-83). It was assumed that equilibrium was reached when the instantaneous change in storage for one time step dropped below 1.0 cubic foot per second. Increases in storage continued for hundreds of years after this point, but the total volume was small.

Specific yield is the most important aquifer property or boundary condition for determining how long the system will take to reach equilibrium. Using constant flux recharge equal to 5,720 acre-feet per year, specific yield equal to 0.02, and conductance coefficient (GHB) equal to 1.0, storage and aquifer head changes became negligible after 80 years. Using constant flux recharge equal to 14,820 acre-feet per year, specific yield equal to 0.15, and conductance coefficient (GHB) equal to 0.1, storage and aquifer head changes became negligible after about 700 years. Changing the conductance coefficient (GHB) from 0.1 to 1.0 and subsurface recharge from 5,720 to 14,820 acre-feet per year makes little difference in the projected head changes. Using specific yield equal to 0.08, the system reaches equilibrium after about 400 years; and at that time the increase in storage for the model area of 600 square miles is 840,000 acre-feet. From 1963-83, the simulated increase in storage is 880,000 acre-feet. Thus, when the system comes close to equilibrium, the present (1983) bank storage will have doubled. Of that total storage after 400 years, 36 percent is projected to occur in 50 years and 57 percent in 100 years.

## DISCUSSION

In this section, the various simulations are evaluated and the differences between the prototype ground-water system (the conceptualization of the system, which is independent of the results of the model calibration) and the simulated systems are discussed. The sensitivity of the model to changes in aquifer properties and boundary conditions also are discussed.

### Evaluation of Simulation Results

The steady-state system was represented with three options of subsurface recharge: 5,720, 10,440, and 14,820 acre-feet per year. All three options resulted in a similar potentiometric surface and used reasonable values of hydraulic conductivity (a range of 0.25 to 3.38 feet per day). The measured water levels in the model area and near the southwest boundary were closely matched and the largest residual was 15 feet. The simulated discharge to the Paria River closely matched the measured gain of 3 cubic feet per second (2,170 acre-feet per year) (table 4). The distribution of subsurface recharge along the boundaries fit the conceptual idea of the system; that is, most of the recharge comes from the Paria Plateau and outcrop area along the East Kaibab monocline (southwest and west boundaries). The model was run with most of the recharge through the northwest boundary, but the system could not be simulated accurately (measured heads in the modeled area could not be matched). This does not confirm the conceptual distribution of recharge, but it does give some confidence in it.

Any one of the recharge options could reasonably simulate the actual steady-state system; thus, there are several possible solutions to that system. The measured water levels could have been matched with larger values of recharge and hydraulic conductivity, however, the recharge option of 14,820 acre-feet per year was considered to be a reasonable upper limit. The steady-state simulation, therefore, shows that hydraulic conductivity could range from 0.25 to 3.5 feet per day and larger values over the entire model area are unlikely.

The results of the transient-state simulations show that a close match could not be developed between the available data and water levels generated by the present model configuration. Therefore, unique estimates of aquifer storage characteristics and boundary-flow quantities can not be made. Additional studies are needed to define the transient-state system more accurately.

During the transient-state simulations, it was found that a variable distribution of hydraulic conductivity (fig. 9) was necessary to simulate the large rises of measured water levels in the Navajo Sandstone a few miles from the lake. The relative change in hydraulic conductivity of the aquifer for the new distribution is 1.0 to 4.5 feet per day. Inital values for hydraulic conductivity of 4.5 feet per day were placed at the axis of the Wahweap syncline, and values of about 2.0 feet per day were placed along the Echo monocline. The reasoning for using a larger conductivity for the Wahweap syncline and Echo monocline, is that they probably are areas of extensive fracturing, and this would result in an increase in secondary permeability. The increased permeability in these areas was mentioned in an earlier study by the U. S. Bureau of Reclamation (Calder, L., written commun., 1965).

The Navajo Sandstone crops out near the southern part of the Wahweap syncline in the model area, and the initial value used for hydraulic conductivity in this area was 1.5. Hence, the new distribution has a range of hydraulic conductivity of 1.5 to 4.5 across the syncline, which gives a ratio of three. Hood and Danielson (1979, p. 36) found a ratio of three between secondary and primary permeability in their study of the Navajo Sandstone near Caineville, Utah, about 80 miles northeast of the area of this study. The two specific capacity values (12.5 and 3.5 gallons per minute per foot) for wells near Wahweap Bay differed by about a ratio of three. The well with a specific capacity of 12.5 gallons per minute per foot is in the axis of the Wahweap syncline, and the other well is to the west of the axis and in a less folded area. A greater ratio of the secondary to primary permeability might have improved the model fit, but without evidence to substantiate a greater ratio, three was set as the upper limit.

The simulated increase in storage to the Navajo Sandstone for the 20-year period simulated by the model is shown in table 5. Using a range of specific yield of 0.02 to 0.11, the increase in storage from 1963-83 ranged from 7,000 to 30,000 acre-feet per mile of shoreline. The recharge option of 10,440 acre-feet per year, with a specific yield equal to 0.08, results in an increase in storage of about 25,000 acre-feet per mile, and this is probably the most reasonable single value. After 10 years of recharge, the option for a specific yield of 0.08 results in an increase in storage of 19,000 acre-feet per mile. This matches an estimate of 19,000 acre-feet per mile for 10 years based on work by R. E. Glover (U. S. Bureau of Reclamation, written commun., 1949) and Jacoby and others (1977).

A range of 0.02 to 0.15 for specific yield is shown in table 5, which was the initial, conceptual range selected for the calibration process. A more precise estimate can not be made, given the data available for this study. In these simulations, the specific yield was kept at a uniform value for the entire model area. It might vary across the area, and it could have been changed to improve the model fit; but there were no data to support this, and the initial assumption of uniform specific yield was kept.

## Sensitivity of Model Boundaries

A discussion of the simulation of the recharge from Lake Powell by the general-head-boundary subroutine (GHB) follows to show the reader how it works in the model and to show that the head changes in the Navajo Sandstone caused by this boundary are reasonable. By varying the conductance coefficient (K/L) of the GHB (see page 23) over a wide range, the relation of this boundary to water levels in the Navajo can be studied. The recharge option of 10,440 acre-feet per year, with specific yield equal to 0.08, was used for the following sensitivity tests. Figure 14 shows the relation between conductance coefficient and water level at node 23-22 (1 mile from the shoreline) 3 years after the begining of the filling of Lake Powell. This relation is similar for all the nodes within a few miles of the lake and within a few years of the begining of filling. With the coefficient in the range of 1 to 100, water levels at all nodes remain the same or change a maximum of 2 feet. As the coefficient is decreased below 1.0, the water levels are more sensitive.

In order to examine the relation of water levels in the Navajo Sandstone with time and distance from Lake Powell using the GHB, several nodes along row 23 were selected at distances of about 1, 2, 3, 4, and 5 miles from the lake shoreline. The conductance coefficient of the GHB was set at 1.0 and 0.01, and the resulting changes in water levels are shown in figure 15. Row 23 was selected because it is perpendicular to the lake, and it is far from the northeast boundary, which is the most uncertain of all boundaries. A few other nodes were examined north of row 23, but they showed similar results to those observed for the nodes in row 23. The year with the largest head difference in figure 15 is the year when the lake has the most influence at a particular distance from the lake. Thus, the second year has the greatest effect 1 mile from the lake and the fourth and fifth years have the greatest effect 3 miles from the lake. These relationships support the idea that the response of the aquifer to the filling of Lake Powell can be visualized as a front of water moving slowly through the sandstone.

The most uncertain boundary in this model is the northeast side, which was modeled as constant flux during the steady-state and transient-state simulations. The quantity of flux was determined during the steady-state calibration, and this resulted in a flow pattern mostly parallel to the boundary. In rows 2-15, about 7 percent of the total discharge from the model flowed out the northwest boundary. In rows 16-27, about 7 percent of the total subsurface inflow to the model area was through the northeast boundary.

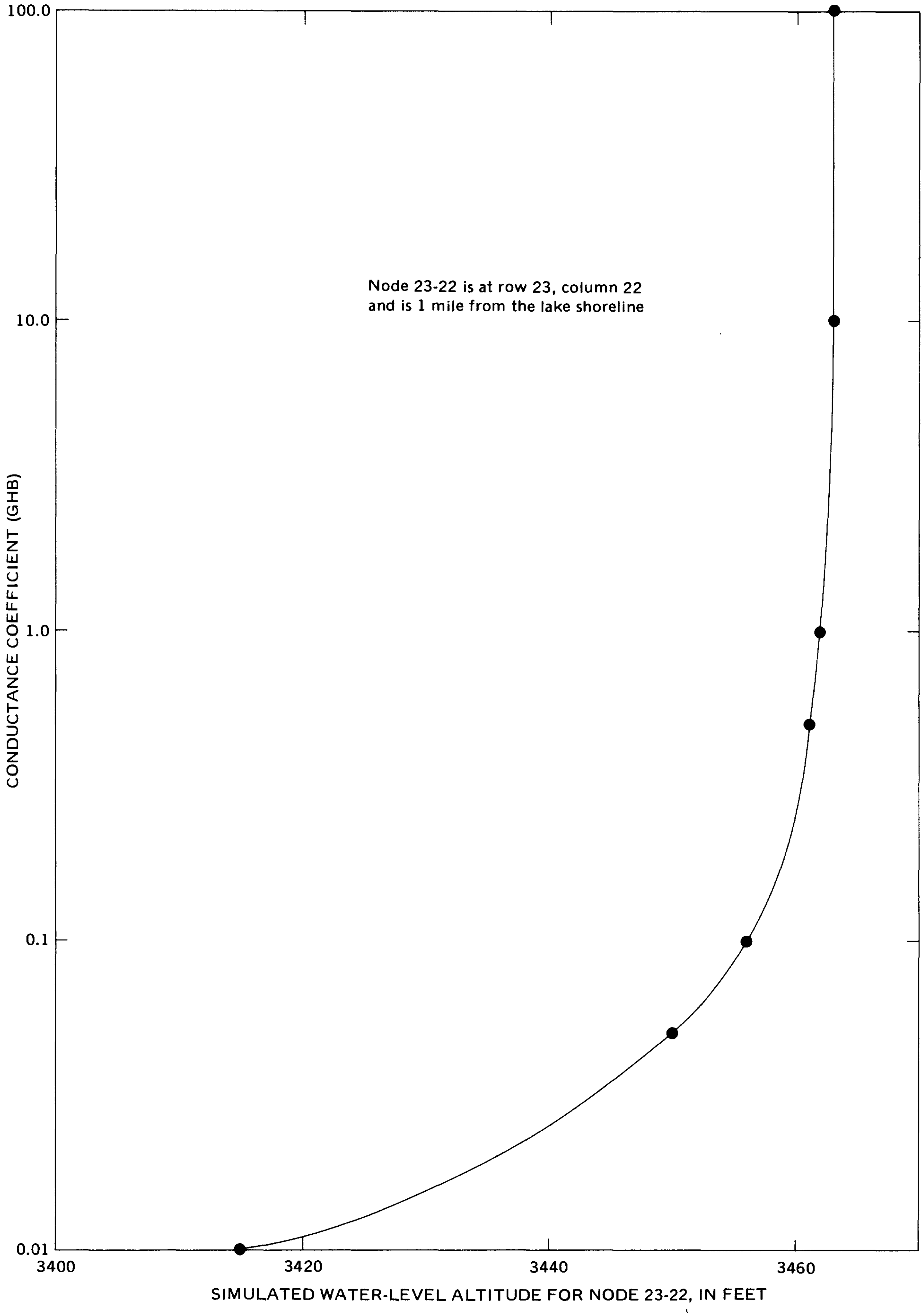



Figure 14.—Relation of conductance coefficient (GHB) to water-level altitude 3 years after the start of filling of Lake Powell for node 23-22.

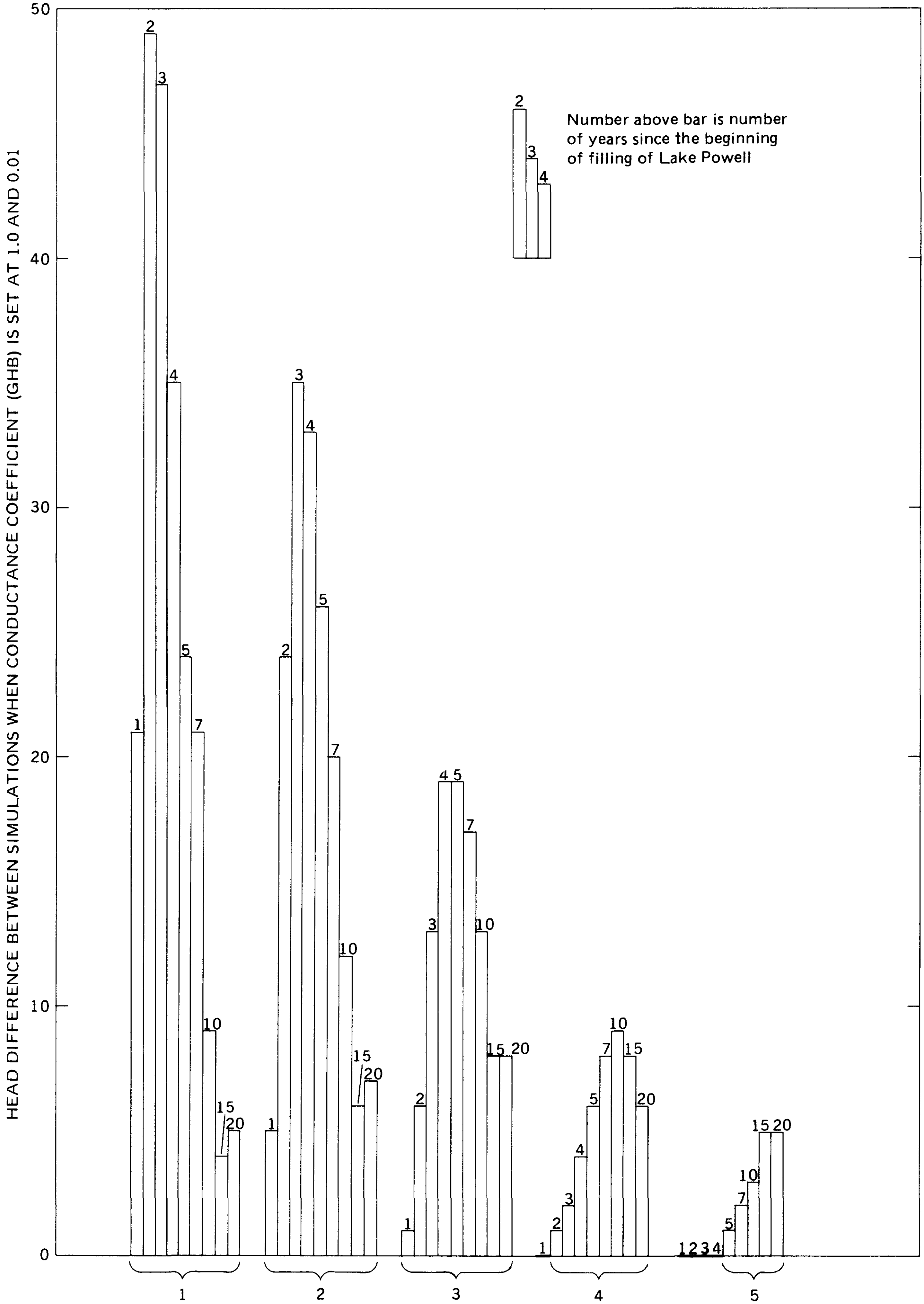



Figure 15.—Sensitivity relations for general-head-boundary condition.

The simulation of the northeast boundary is probably reasonable for the steady-state system; however, the changes along the boundary caused by Lake Powell are uncertain. The assumption made here is that the fluxes across the boundary did not change. The altitude at which ground water discharges at the southern end of the boundary and several miles to the east and west is raised over 200 feet (former Colorado River level versus Lake Powell level in 1983) so the gradient parallel to the boundary is decreased, but the quantity and direction of flow across most of the boundary probably do not change significantly. Comparison of the simulated potentiometric surfaces for steady-state conditions (fig. 10) and for 1983 (fig. 13) shows that the simulated gradient toward the lake (parallel to the northeast boundary) decreased from about 33 feet per mile to 22 feet per mile. The gradient under steady-state conditions across the northeast boundary ranged from 0 to 10 feet per mile, and it decreased by 1 to 3 feet per mile by the end of the transient-state simulation (1983).

The northeast boundary was modeled as one of no flow, with recharge equal to 10,440 acre-feet per year, specific yield equal to 0.08, and the conductance coefficient (GHB) equal to 0.1. Simulated water-level rises at nodes used for calibration were the same using the no-flow or the constant-flux boundary, except at node 6-10 where the simulated rise was 5 feet versus 3 feet for the constant-flux boundary. Therefore, this boundary seems to be reasonable for the purposes of this study.

## ADDITIONAL STUDIES

A more accurate model of the interaction of Lake Powell and water in the Navajo Sandstone in the Wahweap Bay area could be developed if a longer period of water-level measurements in existing wells were available as well as measurements in additional wells. The analysis of the simulation of the lake-aquifer boundary showed that the most important area is within 5 miles of the lake shoreline. The existing observation wells installed by the U. S. Bureau of Reclamation north of the lake are less than 1 mile from the shoreline, and they provide useful information on the near-shoreline response of water levels in the Navajo Sandstone to lake fluctuations. The best location for additional observation wells, would be between 1 and 5 miles from the shoreline. Water-level measurements in such wells need to be made monthly. Additional observation wells are needed from 5 to 30 miles from the lake shoreline to define the regional characteristics of the system. Such wells would only need to be measured once or twice a year.

Additional aquifer tests throughout the area would provide a more complete understanding of the system. Comparison of water levels in the Entrada Sandstone, Carmel Formation, Navajo Sandstone, and the Kayenta and Moenave Formations would provide information on interformational leakage. An investigation of the natural recharge to the system from the Paria Plateau and areas to the north would help to provide a better definition of the steady-state system.

## SUMMARY

A two-dimensional, finite-difference, digital computer model was used to simulate ground-water flow in the Navajo Sandstone near Wahweap Bay, Lake Powell, Utah and Arizona. The filling of Lake Powell started in March 1963; and from 1963-83 the lake rose almost 550 feet, and water levels in the Navajo Sandstone in a well 1 mile from the lake rose 395 feet. The steady-state system (prior to 1963) and the transient-state system (1963-83) were simulated; however, the model could not be calibrated to a single set of input data because of the small quantity of available water-level data and a lack of independent estimates of recharge to the aquifer and its hydraulic properties. Therefore, a range of input data was used for various representations of the system.

A steady-state model simulated subsurface recharge options of 5,720, 10,440, and 14,820 acre-feet per year, which resulted in a range of hydraulic conductivity of 0.25 to 3.38 feet per day. Transient-state simulations did not result in a decrease in the conceptual range of specific yield of 0.02 to 0.15. Transient-state water levels were not greatly affected by storage coefficient in the range of 0.001 to 0.000001, and a value of 0.0001 was used in all simulations. The results of the transient-state simulations indicate that permeability of the aquifer increases in the Wahweap syncline and Echo monocline. The change in aquifer storage from 1963-83 was estimated to range from 7,000 to 30,000 acre-feet per mile of lake shoreline using the three recharge options and a range of specific yield of 0.02 to 0.11. The model was run to simulate 1,500 years at a constant lake level; and the system was estimated to reach equilibrium between 80 and 700 years using a range of specific yield of 0.02 to 0.15 and a range of subsurface recharge of 5,720 to 14,820 acre-feet per year. Additional field data are needed to develop a more accurate model of the interaction of water in the Navajo Sandstone and in Lake Powell.

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