Research Technical Completion Report

APPLICATION OF A NUMERICAL GROUND-WATER FLOW MODEL TO THE MUD LAKE AREA IN SOUTHEASTERN IDAHO

by

G.S. Johnson
C.E. Brockway
S.P. Luttrell

Kimberly Research and Extension Center

Idaho Water Resources Research Institute
University of Idaho
Moscow, Idaho 83843

December, 1984

TECHNICAL COMPLETION REPORT

APPLICATION OF A NUMERICAL GROUND-WATER FLOW MODEL TO THE MUD LAKE AREA IN SOUTHEASTERN IDAHO

by
G. S. Johnson, C. E. Brockway, and S. P. Luttrell

University of Idaho

Research supported by the U.S. Geological Survey
Department of the Interior
Contract No. 14-08-0001-A-0016

A contribution of the
Regional Aquifer-System Analysis Program
this study under Contract No. 14-08-0001-A-0016

Idaho Water Resources Research Institute
University of Idaho
Moscow, Idaho

December, 1984

CONTENTS

ILLUSTRATIONS

TABLES

ABSTRACT

The regional Snake Plain aquifer is the primary source of ground water throughout the eastern Snake River Plain including the Mud Lake area. Ground-water flow from mountain valleys to the north of the Mud Lake area annually supply about 247,000 acre-feet of recharge to the regional aquifer. Ground water exits the study area in a southwesterly direction as part of the Snake River Plain regional ground-water flow system. A local, overlying ground-water system in the alluvium east of Mud Lake annually leaks about 113,000 acre-feet of recharge to the regional aquifer. About 143,000 acres of agricultural land present within the study area are irrigated from local and regional ground-water sources and from surface water.

A numerical ground-water flow model was calibrated to determine variations in transmissivity and storage coefficient. Calibration was based on hydrologic data collected from April, 1980 through March, 1981; included were several determinations of the water table based on depth to water measurements. A band of steep hydraulic gradients transects the study area and coincides with a band of relatively low transmissivities, less than 4.0×10^4 ft^2/day. Transmissivities in the remainder of the area generally exceed 1.0×10^6 ft^2/day. Refined calibration of the storage coefficient was not possible in some areas due to little change in water levels during the calibration period.

A 20-year simulation of current irrigation practices under normal hydrologic conditions predicts a general decline in ground-water levels. Simulation of a hypothetical increase in water use for irrigation predicts an additional decline of as much as 80 feet. Simulation of

186,000 acre-feet of artificial recharge biennially results in a ground-water mound reaching a maximum height of 40 feet. Such a rise could cause temporary flooding of depressions in the vicinity of the recharge.

INTRODUCTION

Purpose

The demand placed on the water resources of southern Idaho by agriculture, industry, and increasing population continues to expand. With diminishing availability of surface-water supplies, future growth must rely heavily on ground-water sources, particularly the Snake Plain aquifer. Improved hydrogeologic knowledge about the aquifer is needed to fully develop the ground-water resource with minimal adverse effects.

The purpose of this investigation is to develop a better understanding of the ground-water flow system in the Mud Lake area and determine its relationship to the rest of the Snake Plain aquifer by use of a numerical ground-water flow model. This study was a cooperative effort by the University of Idaho and the U. S. Geological Survey as part of the Geological Survey's RASA (Regional Aquifer Systems Analysis) study of the Snake River Plain.

Location and Description

The study area consists of about 870 mi^2 in the easternmost part of the Snake River Plain in southeastern Idaho (fig. 1). It is bounded by mountains to the north, land irrigated from the Snake River and tributaries to the east, and by a combination of rangeland and irrigated agricultural land on the south and west. The study area is located in parts of Clark, Jefferson, and Fremont counties and is centered at approximately 44° 00′ latitude, 112° 15′ longitude. Land surface altitudes vary from about 4800 to 5400 feet above mean sea level.

The climate is semi-arid with a mean annual precipitation ranging

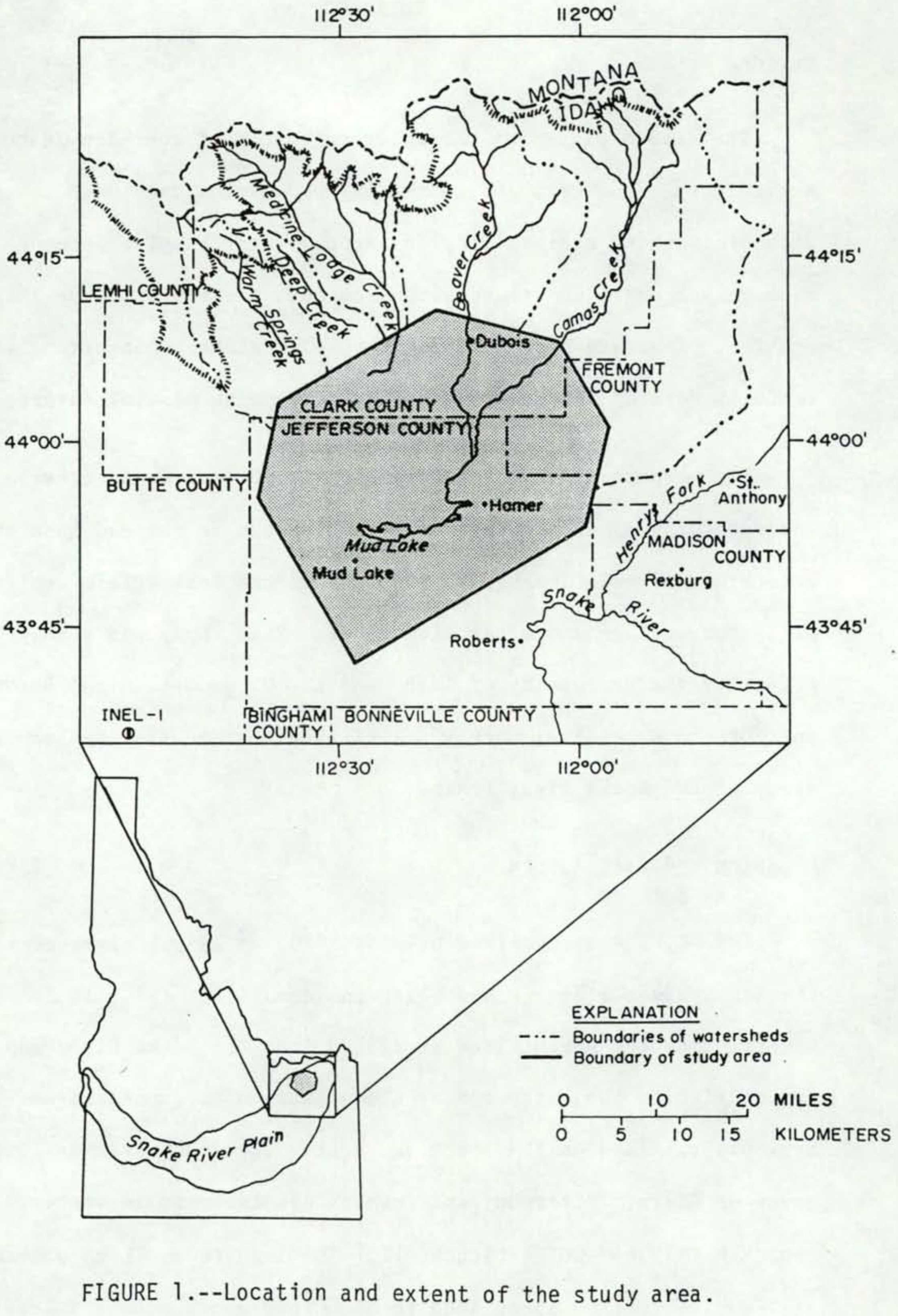

FIGURE 1.--Location and extent of the study area.

from 10 to 15 inches, which occurs mainly in spring. Winter temperatures may remain below freezing for several weeks at a time with occasional periods of thawing. Surface water enters the area from Medicine Lodge, Beaver, and Camas creeks which flow out of the mountains to the north. No surface water leaves the study area. Topography varies from flat lakebeds to highly irregular basalt outcroppings.

Approximately 25 percent of the study area is devoted to irrigated agriculture; the remainder is mostly rangeland. Principal crops are alfalfa, grain, and potatoes. Slightly over half of the agricultural lands are sprinkler irrigated supplied by on-site ground-water pumping. The remainder is primarily surface irrigated with surface water or ground water conveyed by canal systems. Mud Lake is used cooperatively by several canal companies as an irrigation reservoir supplied by a combination of surface water and ground-water. Several square miles of marshland just north of Mud Lake are maintained for wildlife.

Previous Investigations

Previous ground-water investigations specifically of the Mud Lake area have been made by Stearns, Bryan, and Crandall (1939), Ralston and Chapman (1969), and Crosthwaite (1973). Other studies which have included the Mud Lake area are Stearns, Crandall, and Steward (1938), Skibitzke and da Costa (1962), Mundorff, Crosthwaite, and Kilburn (1964), and Norvitch, Thomas, and Madison (1969).

Methods of Investigation

The hydrogeology and ground-water flow characteristics of the Mud Lake area, cooperatively investigated with this study, are reported by

Luttrell (1982).

The study area was divided into a 1-mile square grid with a model node at the center of each grid cell. Data for model calibration were collected for the period of April 1, 1980 through March 31, 1981 and organized into 24 equal timesteps (15.21 days).

Stream flows and irrigation diversions were determined from field measurements and U. S. Geological Survey publications. Land use information was gathered from maps, aerial photos, and published reports. Meteorological data from the U. S. National Oceanic and Atmospheric Administration were used for determination of aquifer recharge and discharge.

Depth to water was measured in a network of 224 wells at four different times during the data collection period. Selected measurements were used in a computer program which numerically interpolates head values at each grid location and plots water-table contours.

Hydrologic data were input to a computer program to determine net aquifer recharge and discharge (surface flux) in every cell for each timestep. Surface flux and the aquifer head at each node were input for model calibration of transmissivity and storage coefficient. The model uses a two dimensional finite difference method described by deSonneville (1972), deSonneville (1974), and Johnson and Brockway (1983).

The calibrated model was used to simulate long-term effects of current irrigation practices with normal hydrologic conditions. The model was also applied to simulate aquifer response to hypothetical situations of increased irrigation pumping and application of artificial recharge.

Acknowledgements

The investigation was a cooperative effort by the University of Idaho, through the Water Resources Research Institute, with funding from the U. S. Geological Survey. The Idaho Department of Water Resources provided data, equipment, and funding. Thanks are extended to the residents of the area for their cooperation and use of their wells for depth to ground-water measurements. The authors are also grateful to Mr. Don Shenton, watermaster in the Mud Lake area, and several individuals who assisted in data collection.

HYDROGEOLOGY

Geology

The eastern Snake River Plain is a large graben or downwarped structural basin. It has been filled to its present level with 2,000 to 10,000 feet of rhyolite deposits, thin basaltic lava flows, and interbedded sediments (Barraclough, Lewis and Jensen, 1981, p.4). Surface geology of the Mud Lake area is illustrated in figure 2.

Rhyolitic rocks underlie the basalt in much of the eastern plain, usually at depths of less than 3000 feet. Although their hydraulic conductivity is highly variable, rhyolitic rocks seldom contain large quantities of ground water. (Luttrell, 1982, p. 15).

Basaltic rocks of the Snake River Group are the predominant rock type in the study area. Being fractured and in places highly vesicular, the Snake River basalts have a high horizontal hydraulic conductivity. As a result they are the major source of ground water over most of the plain (Luttrell, 1982, p. 17). Sediments in the Mud Lake area include fluvial deposits of gravel, sand, silt, and clay. Coarse deposits occur primarily as broad coalescing alluvial fans which extend southward and southeastward from the bordering mountains. Fine-grained lacustrine sediments accumulated near Mud Lake in a closed depression in the basalt. The lacustrine deposits interfinger with alluvial fan deposits both of which interfinger with basalt flows. (Luttrell, 1982, p. 22).

Detailed descriptions of the geology are given by Stearns, Crandall and Steward (1938), Stearns, Bryan, and Crandall (1939), Mundorff, Crosthwaite, and Kilburn (1964), Crosthwaite (1973), Robertson, Schoen, and Barraclough (1974), and Luttrell (1982).

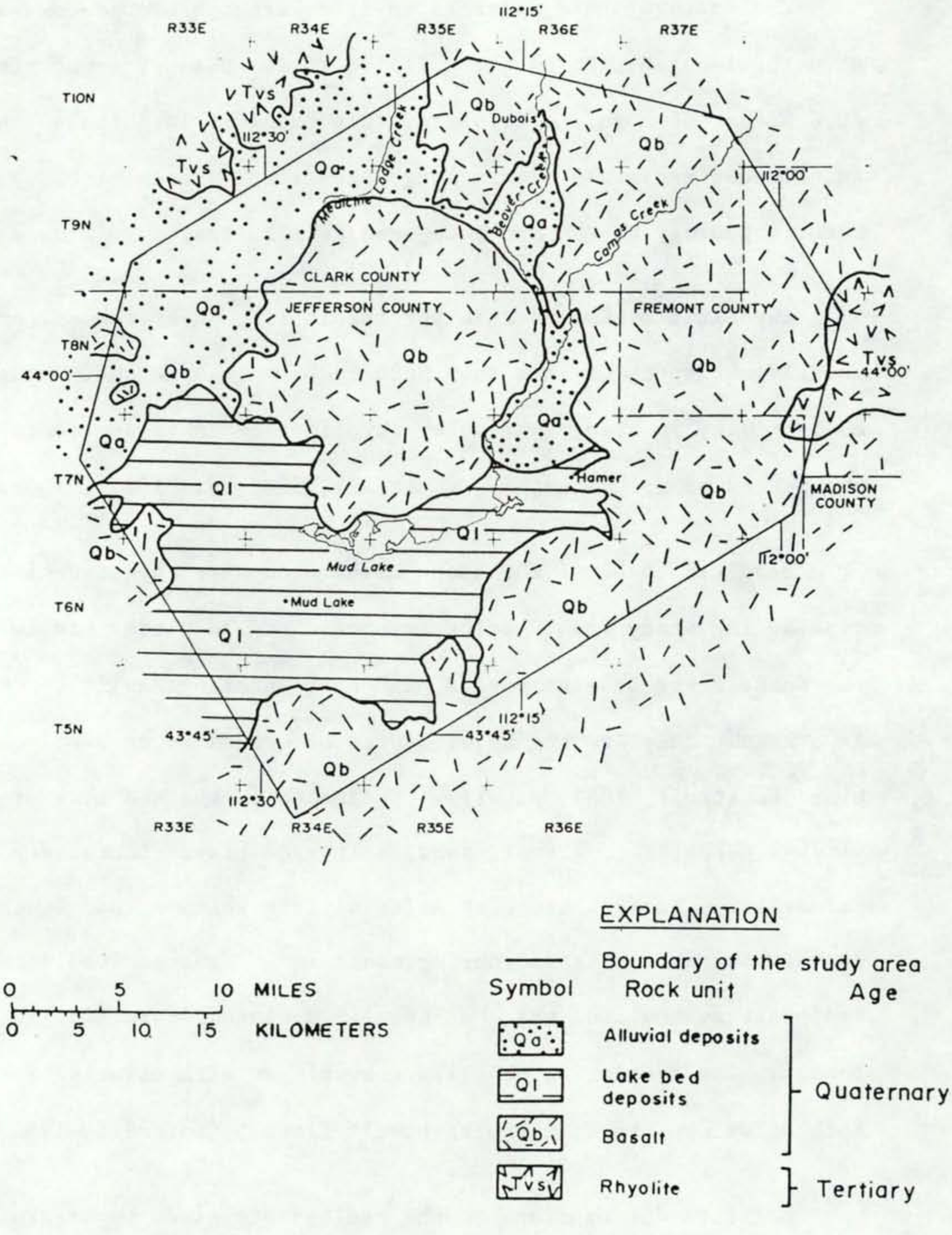

FIGURE 2.--Generalized geologic map (after Luttrell, 1982).

The Shallow Ground-Water System

Ground water in the alluvium overlies the Snake Plain aquifer in a 96 mi^2 area shown in figure 3. The water table in the local system ranges from 6 to 75 feet below land surface. The texture of the saturated alluvium grades from layers of gravel, sand, and clay in the north, to a predominance of silt and clay in the south. Saturated thickness is highly variable but in most areas is less than 200 feet. The extent and geohydrologic properties of the saturated alluvium were determined from drillers′ logs and depth to water measurements.

The local aquifer is recharged by percolation from the land surface and by underflow along the eastern edge, and is discharged by irrigation pumping and leakage to the regional aquifer. Net recharge (recharge minus pumping withdrawals) from land surface for the period April 1, 1980 through March 31, 1981 was estimated to be 113,000 acre-feet. Underflow, based on analysis of hydraulic gradients and estimates of transmissivity, was estimated to be insignificant relative to surface percolation. Hydraulic gradients were determined from depth to water measurements and from water levels reported by drillers and by Ralston and Chapman (1969). Transmissivity estimates were based on typical values for materials reported on drillers′ logs.

Ground water in the alluvium discharges by downward leakage to the Snake Plain aquifer throughout the saturated region, except along the eastern edge where heads in the regional aquifer are similar to those of the shallow system. Discharge also occurs by pumping from numerous domestic wells and from several irrigation wells, mostly in Clark

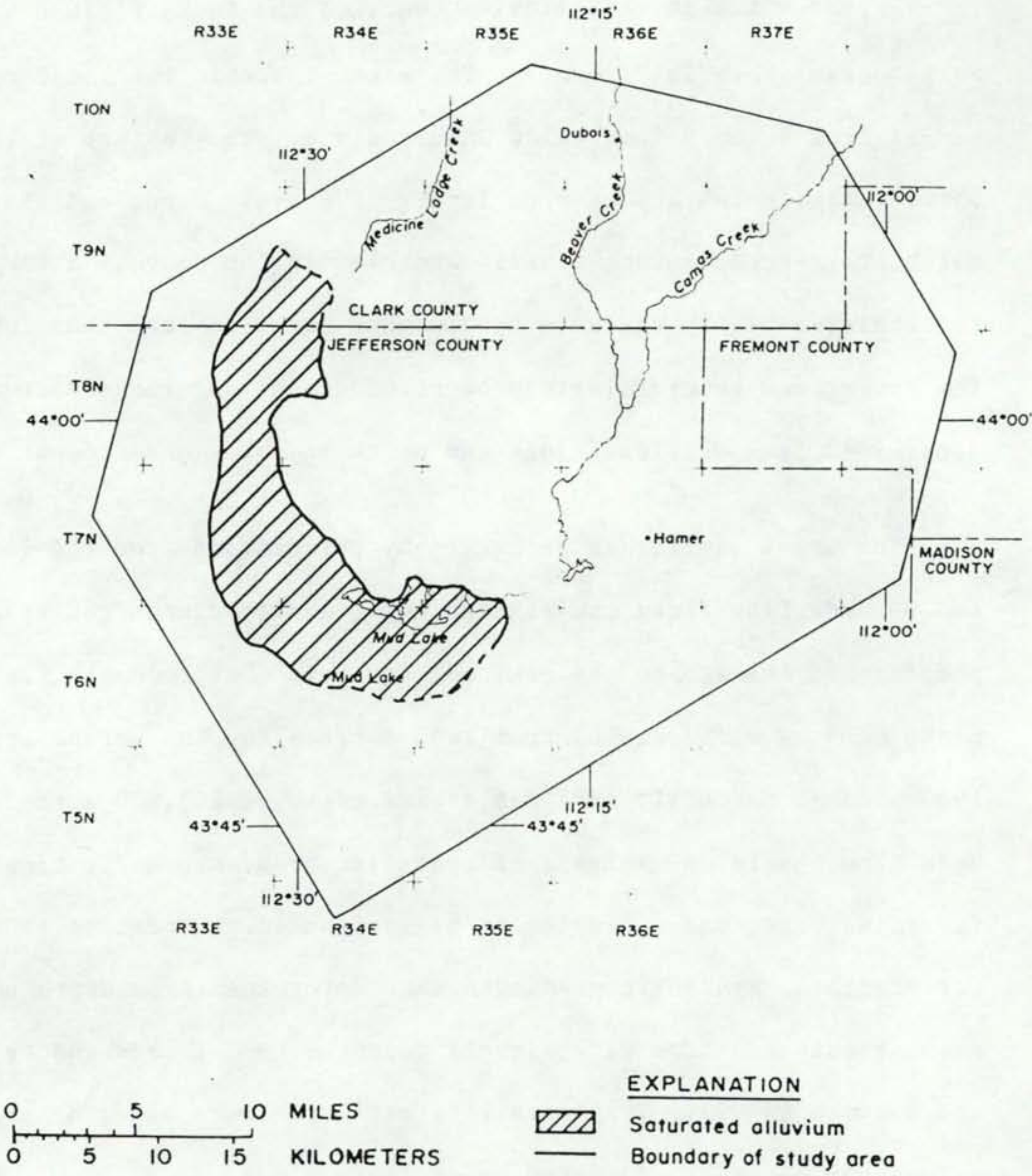

FIGURE 3.--Extent of the saturated alluvium.

County. Leakage, estimated as the difference between recharge and pumpage withdrawls, is about 113,000 acre-feet annually. Water levels in the alluvium could not be mapped due to erratic variations of head with respect to depth and areal distribution.

Hydraulic properties of the alluvium are highly variable. Coarse materials in the northern part of the study area, township 9N, yield quantities suitable for several irrigation wells. A few miles south, aquifer yields are sufficient only for domestic wells. Further south, near Mud Lake, yields are usually insufficient for even domestic wells.

The Regional Ground-Water System

The Snake Plain aquifer underlies much of the 10,800 mi^2 eastern Snake River Plain including the Mud Lake area. It consists of basalts and interbedded sediments of the Snake River Group. Recharge to the regional aquifer is primarily from mountain runoff on the northern and eastern sides of the Plain and from percolation of irrigation water. Regional ground-water flow is southwesterly. Major discharge from the aquifer is to the Snake River in the vicinity of American Falls reservoir and in the Twin Falls to Hagerman reach. Discharge to the Snake River in these reaches was estimated by L. C. Kjelstrom (U. S. Geological Survey, written communication, 1983) to be 6.3 million acre-feet in 1980.

The regional aquifer underlying the Mud Lake study area is recharged by underflow across the northern, northwestern, and northeastern boundaries. Recharge also results from creek and canal seepage, percolation from irrigation and precipitation, and leakage from the shallow ground-water system.

The regional aquifer discharges from the study area by underflow across the southern and western study boundaries. Discharge also occurs by irrigation pumping and by direct evapotranspiration where the water table is shallow.

The bottom of the regional aquifer in the Mud Lake area consists of rhyolitic rocks near the northern mountains and basaltic rocks of low hydraulic conductivity in the rest of the area. Decreasing hydraulic conductivity of the basalts underlying the Snake River Plain with increasing depth was proposed by Mundorff, Crosthwaite, and Kilburn (1964, p. 27). This concept is supported by Robertson, Schoen, and Barraclough (1974, p. 12) who suggest that the upper 500 feet of the aquifer are the most permeable. Aquifer tests in a test hole drilled to about 10,000 feet on the Idaho National Engineering Laboratory site (INEL-1, see fig. 1) indicate that transmissivity in the upper 1200 feet may be 10^6 to 10^8 ft^2/day; below 1500 feet the transmissivity may be 10^3 ft^2/day or less (J. T. Barraclough, written communication, 1981).

Sediment interbeds and massive basalts may be locally confining and cause changes in head with depth throughout much of the Mud Lake area. Piezometers to depths of 1000 feet indicate that heads vary from 140 feet above the water table to 190 feet below, depending upon location and depth. No consistent relationship of head to depth is apparent from the few piezometers showing multiple water levels. The water-table map and model results presented in this report are representative of heads in the upper part of the regional aquifer. The regional aquifer is unconfined except in the area shown in figure 4 where it is confined beneath thick alluvial deposits. The extent of the confined area was

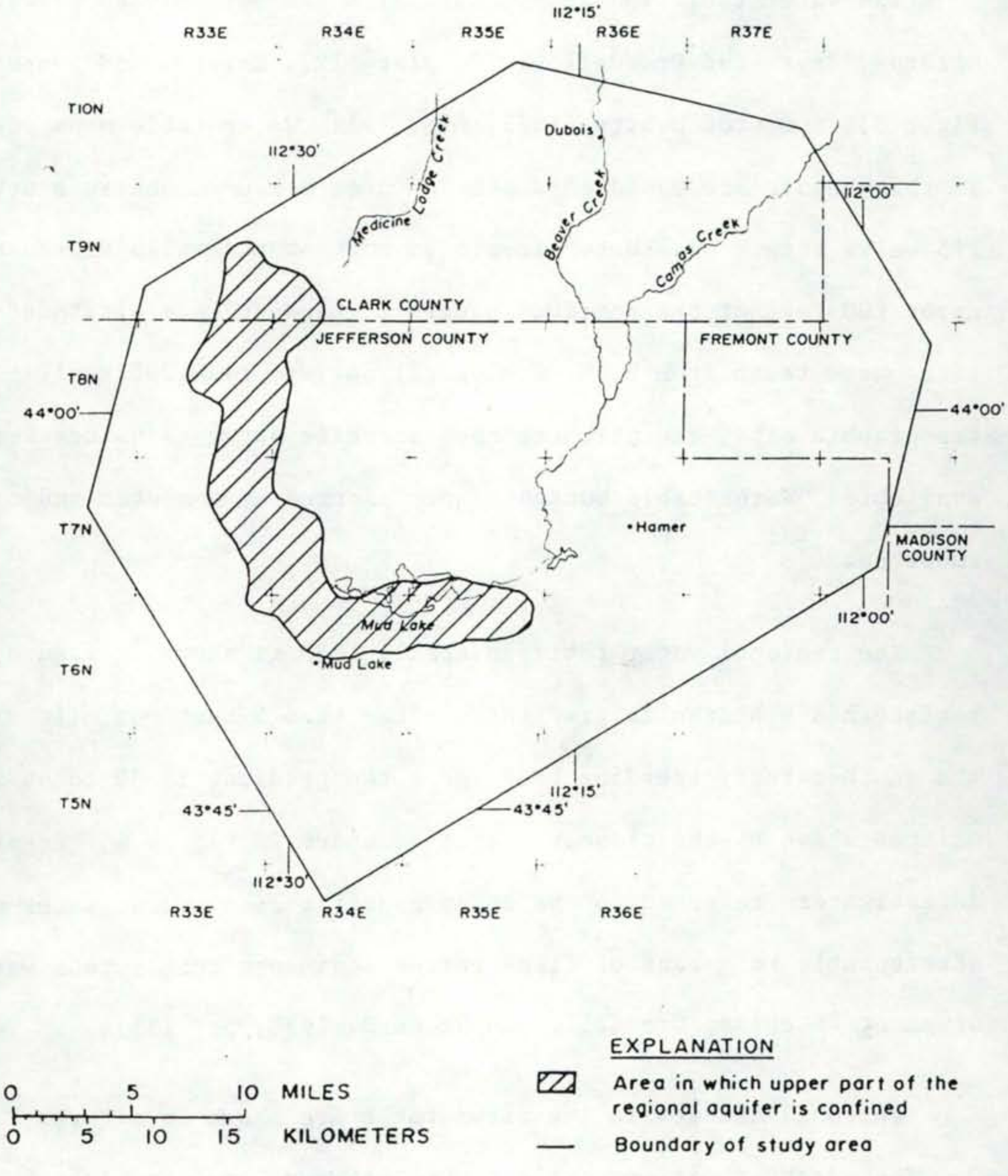

FIGURE 4.--Extent of the area in which the upper part of the regional aquifer is confined.

determined by comparison of aquifer head to the bottom of the fine grained sediment deposits.

The water table in the Mud Lake area has been mapped previously by Stearns, Bryan and Crandall (1939, plate 12), Ralston and Chapman (1969, fig. 3), and Crosthwaite (1973, fig. 3). Water-table maps presented in this report are based on depth to water measurements in a network of 115 wells (fig. 5). Water levels in most of the wells represent the upper 100 feet of the regional aquifer. Land-surface altitudes at well sites were taken from U. S. Geological Survey, 1:24,000 scale topographic maps, except where more accurate surveyed values were available. Water-table contours were plotted by computer and manually smoothed.

The regional water table in April, 1980 is shown in figure 6. The surface has a hydraulic gradient of less than 5 feet per mile except for the southeasterly trending band where the gradient is 30 to 60 feet per mile as shown by the closely spaced contours in figure 6. Previous investigators referred to the steep gradient as a ground-water cascade attributable to a band of fine-grained sediments that impede water movement (Stearns, Crandall, and Steward, 1938, p. 111).

Seasonal changes in the water table are shown in figures 7 through 9. The greatest differences occurred between April and August, 1980, when the water table dropped nearly 60 feet in an area just northwest of Mud Lake. The area of decline is within the band of steep hydraulic gradients where part of the aquifer is confined. No major withdrawals occur in that area, however, substantial ground-water pumping by canal companies does occur just up-gradient of the head drop. Well fields are

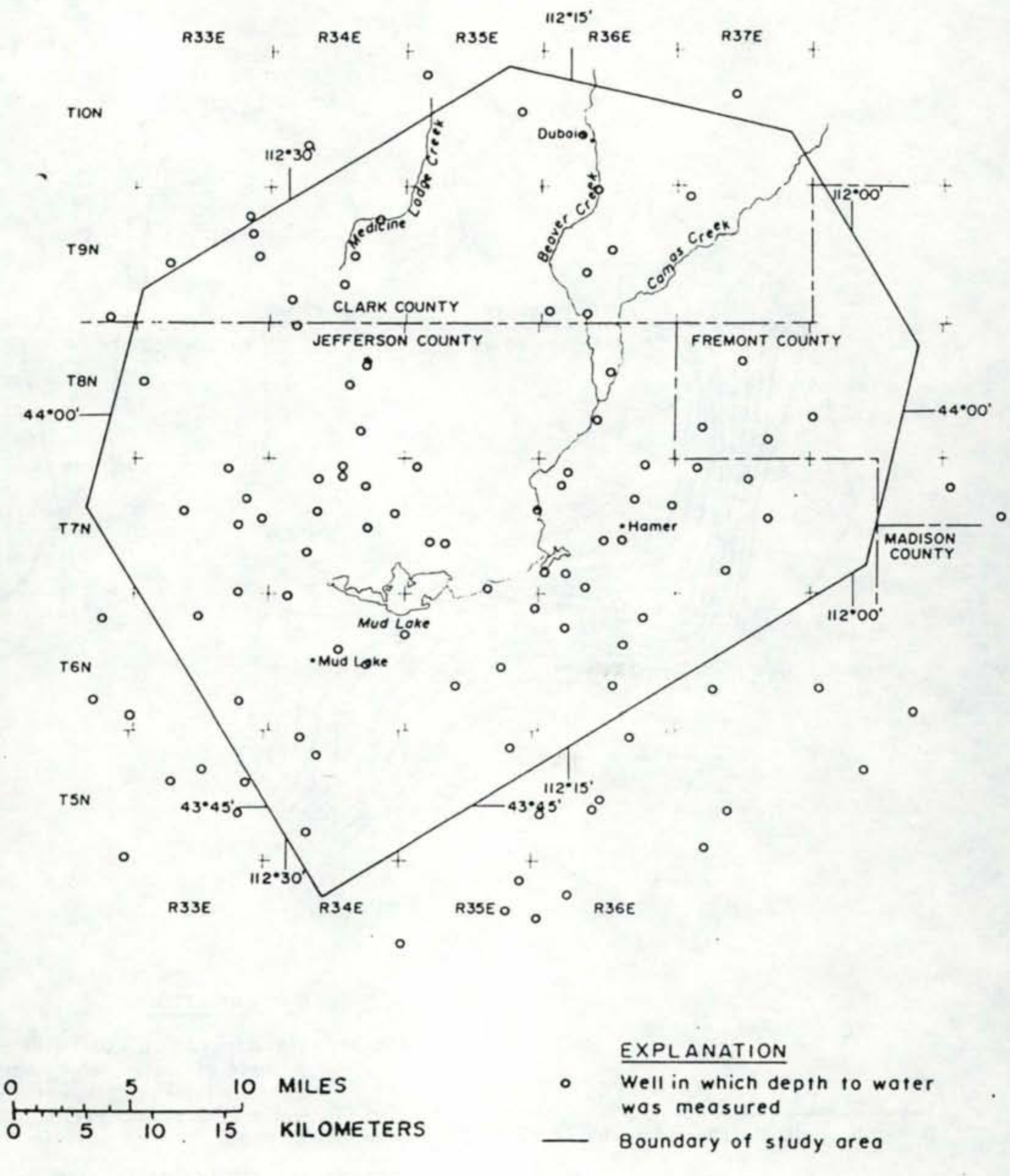

FIGURE 5.--Locations of wells used to define the water table.

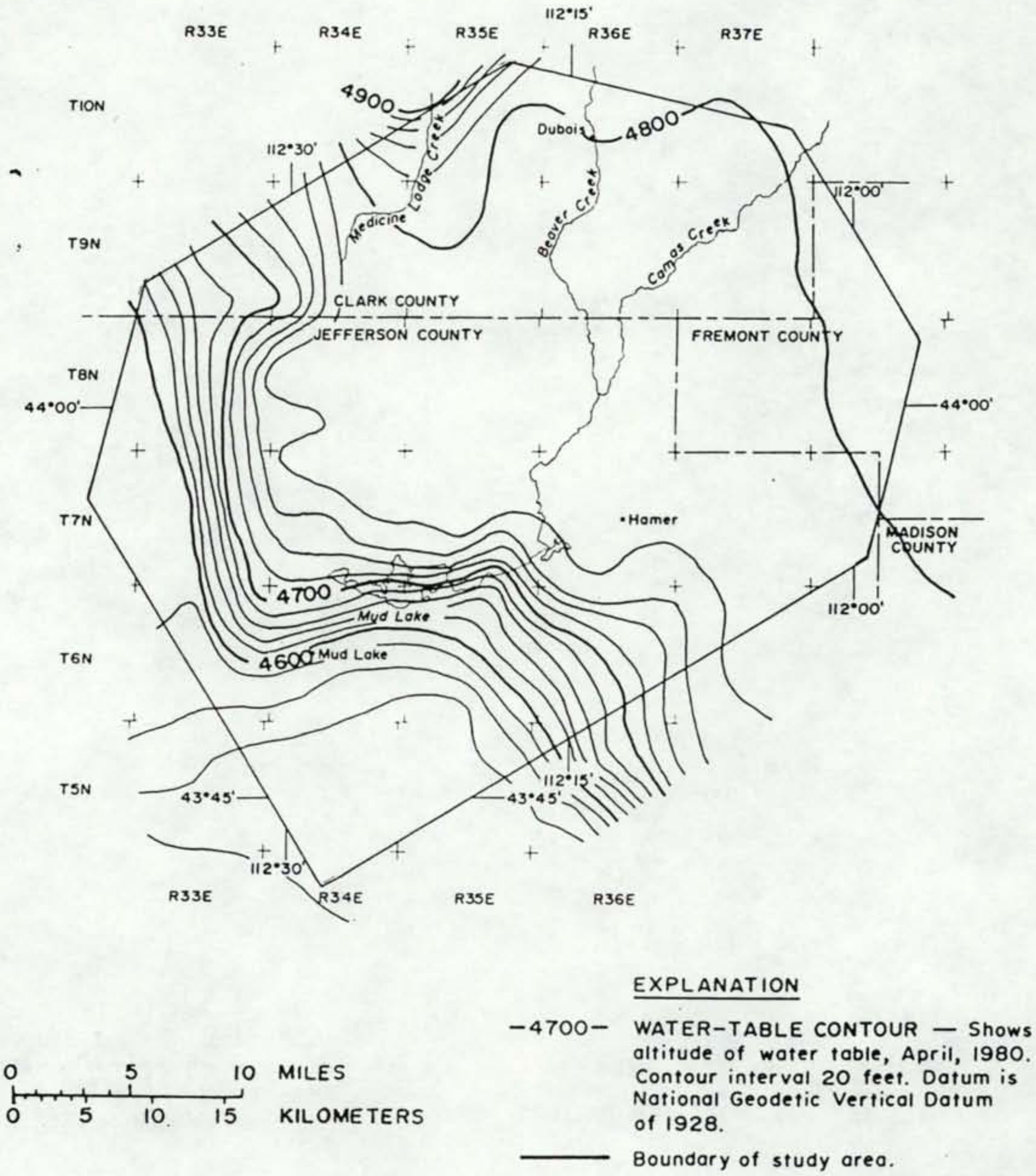

FIGURE 6.--Water-table configuration, April, 1980.

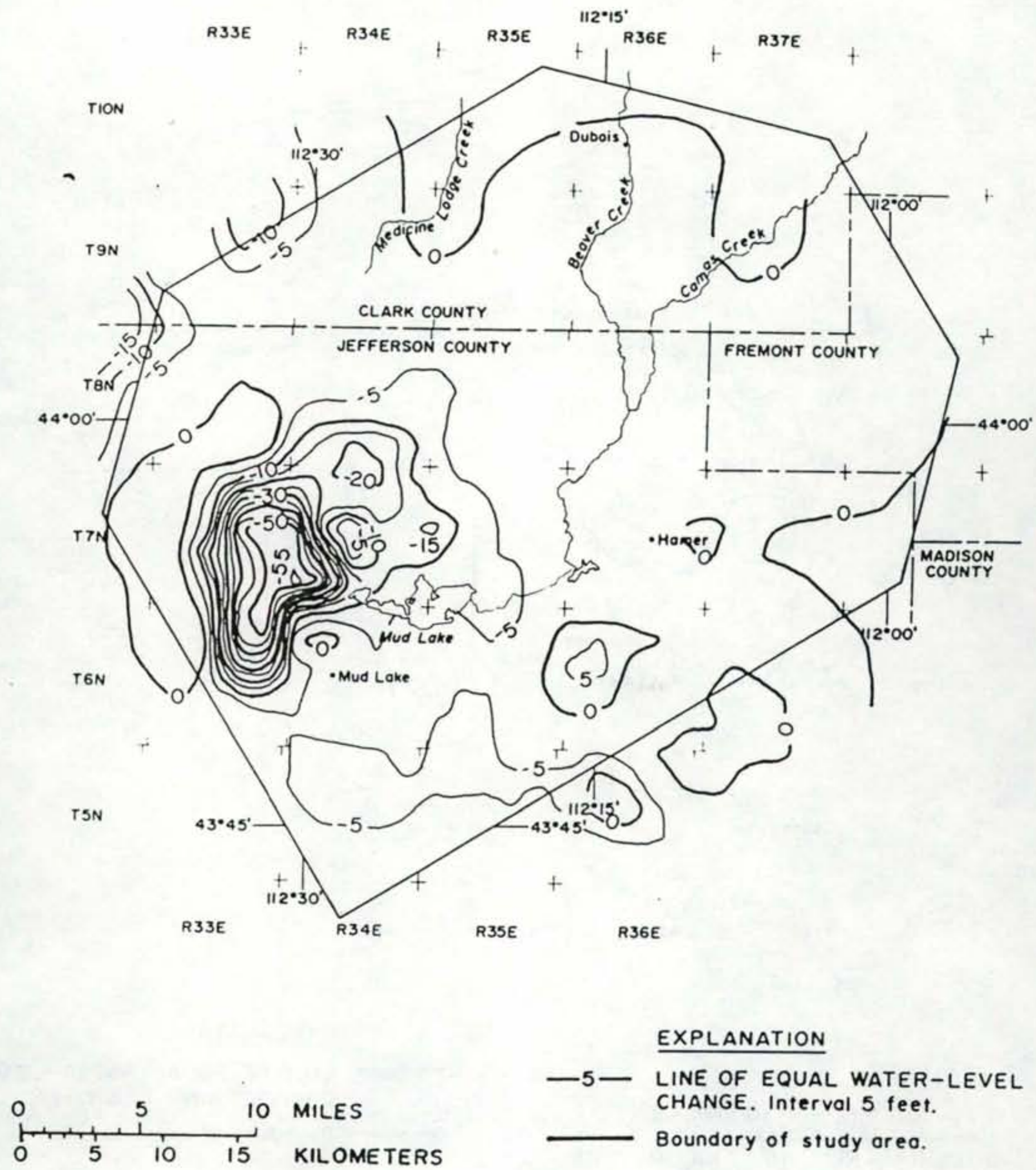

FIGURE 7.--Changes in the water table, April to August, 1980.

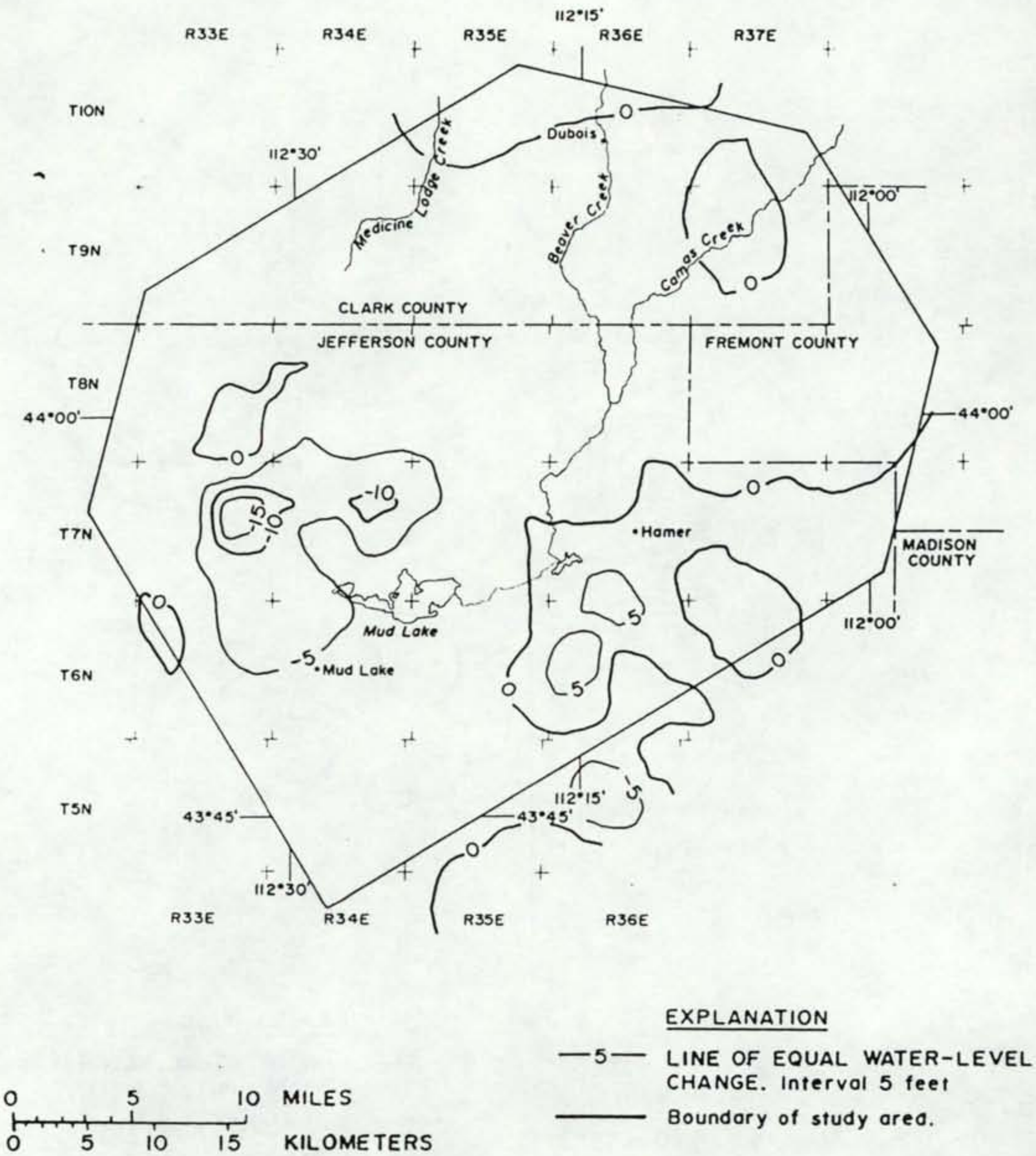

FIGURE 8.--Changes in the water table, April to November, 1980.

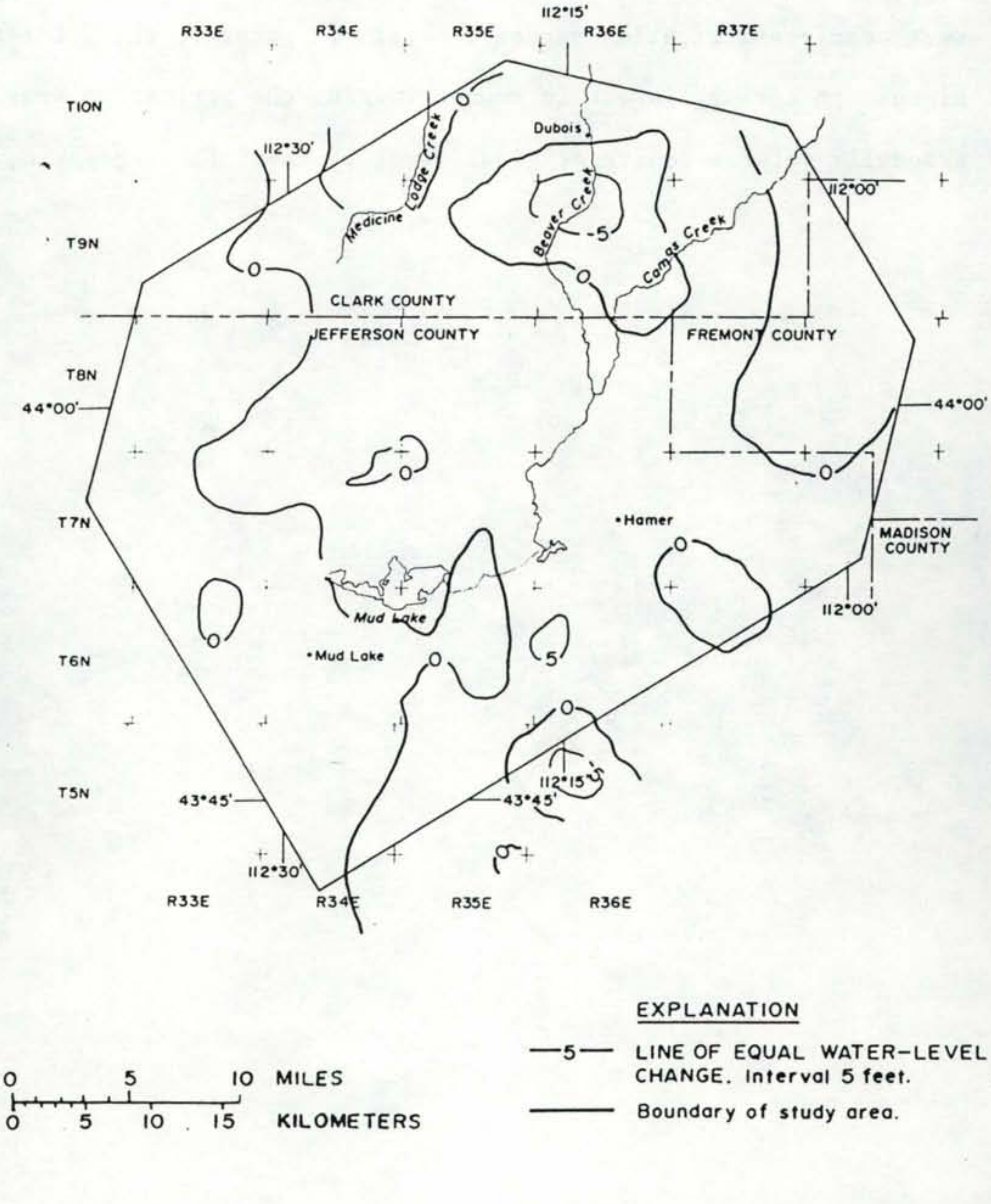

FIGURE 9.--Changes in water table, April, 1980 to April, 1981.

located at the origins of canal systems shown in figure 13. By November, 1980 water levels recovered to within 20 feet of pre-irrigation season levels, and by the following spring water levels were near pre-irrigation season levels. In general, the water table is highest in spring, lowest in summer (during the irrigation season), and gradually returns to its original level by the following spring.

SIMULATION MODEL AND RECHARGE PROGRAM

The selected ground-water flow model is a two-dimensional, finite-difference routine employing an alternating direction technique (Johnson and Brockway, 1983). The model was originally developed by deSonneville (1972) and is described in additional detail in deSonneville (1974). A data management routine titled "Recharge Program" was used to determine surface flux to the aquifer on a cell by cell basis and subsequently served as input to the model.

The model simulates flow under confined, unconfined, or a combination of the two conditions. The simulated aquifer may be homogeneous or heterogeneous and isotropic or anisotropic. Boundaries may be irregular in shape and may be either fixed head or fixed flux. Leakage from one overlying or underlying constant head aquifer may be calculated within the program. The grid interval must be constant and equal in the X and Y directions. The program also includes a routine for automatic calibration of hydraulic conductivity, storage coefficient, leaky aquifer heads, and vertical conductivity between the aquifer and leaky layer. Although calibration adjustments are made to hydraulic conductivity, proportional changes are effected in transmissivity. This report will consider only transmissivity to eliminate potential errors introduced in estimates of saturated thickness. Calibration is based on a comparison of simulated to measured head values.

The model has been applied several times previously to parts of the Snake River Plain. Ground-water flow in the Rigby-Ririe area and the entire eastern Snake River Plain was modelled by deSonneville (1974).

Newton (1978) updated the eastern Snake River Plain model. The same model was applied to the Silver Creek aquifer system in Blaine County, Idaho (Brockway, 1978), to the Henry's Fork area (Wytzes, 1980), and to the Boise Valley aquifer (Lindgren, 1982).

Hydrologic data input to the "Recharge Program" were used to determine surface flux to the ground-water system for each cell and timestep. The program accounts for flux from canal seepage, irrigation percolation, precipitation, evapotranspiration, changes in soil moisture, well withdrawals, river gains and losses, and boundary underflow. All factors included in the "Recharge Program" must be independent of the aquifer head.

MODEL APPLICATION TO THE MUD LAKE AREA

Grid System

The finite-difference grid has the same orientation as the grid used by the U. S. Geological Survey for their Snake River Plain regional ground-water flow model. Grid origin is located at 43° 36′ 27" latitude, 112° 30′ 57″ longitude, corresponding to row 12, column 37 of the Survey's regional grid. The grid is rotated 31.4 degrees counterclockwise to a Universal Transverse Mercator Projection grid with a non-standard central meridian of 113.5°. The grid interval is 1-mile as shown in figure 10. Node points are at the center of each cell (block centered) and represent the average properties and total surface flux within that cell.

Model Boundaries

Model boundaries were assigned a constant flux in all areas where underflow could be reasonably estimated (fig. 10). The remaining boundaries were designated as fixed head.

Water yield and underflow from watersheds north of the study area (shown in figure 1) are listed in table 1. Underflow was calculated as the difference between streamflow and basin water yield. The method of determining water yield depended on the data available for each watershed.

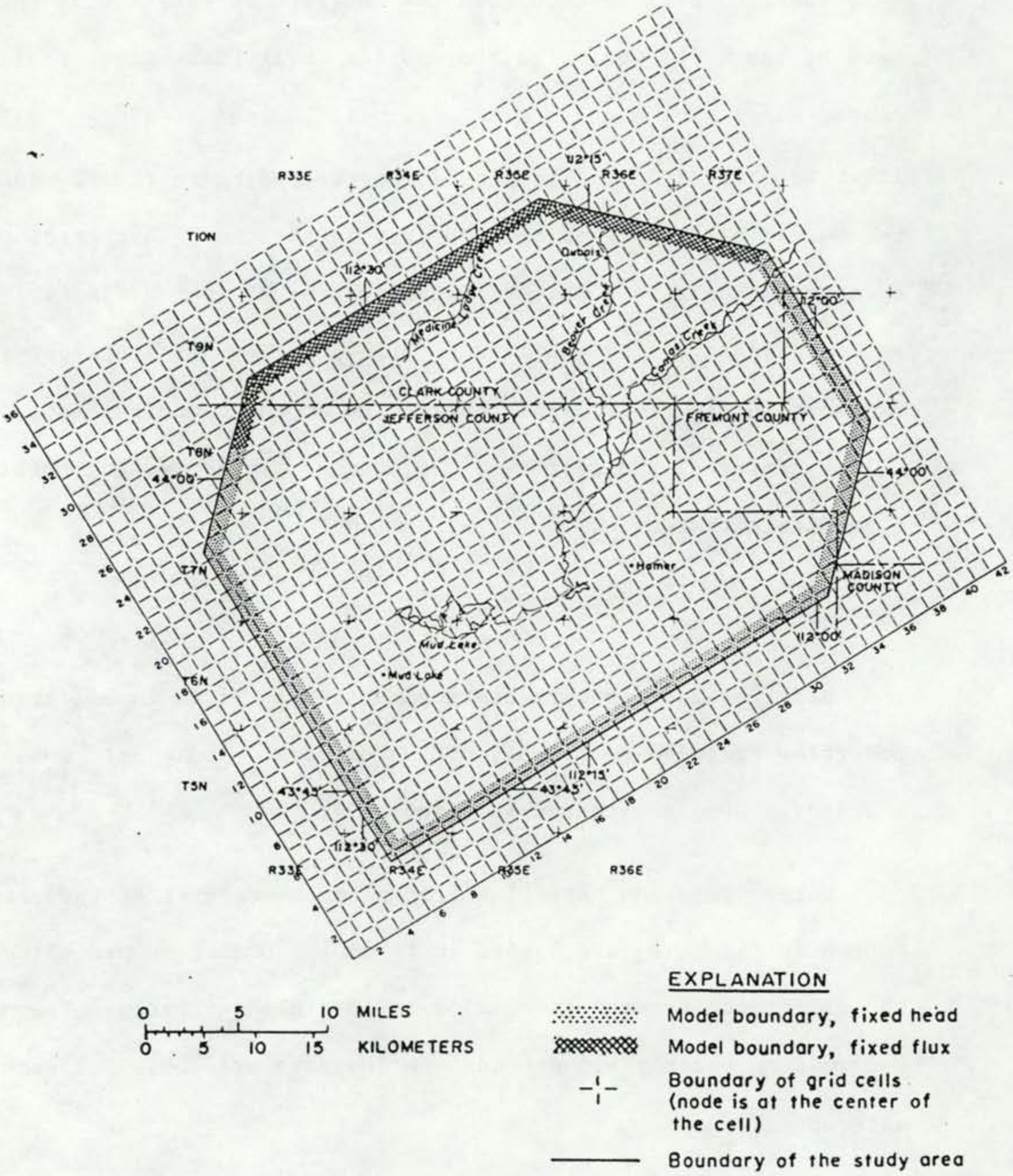

FIGURE 10.--Model grid and boundaries.

TABLE 1. Water yield and underflow from watersheds north of the study area for the period April 1, 1980 through March 31, 1981.

Watershed	Water yield (acre-feet)	Stream flow (acre-feet)	Underflow (acre-feet)
Warm Springs Creek	25,600	0	25,600
Deep Creek	3,300	0	3,300
Medicine Lodge Creek	58,600	50,200	8,400
Beaver Creek	92,300	33,100	59,200
Camas Creek	215,700	65,000	150,700

The water yield from Warm Springs Creek and Deep Creek watersheds were calculated using the following equation:

$$Q = .000903 \, (A^{0.90}) \, (P^{1.83}) \, (F^{0.29}) \quad (cfs) \quad (1)$$

where

A is the watershed area (sq.mi.),

P is the annual precipitation (in.),

F is the percentage of the watershed forested.

The equation was developed by the U. S. Geological Survey and is based on known water yields in southern Idaho (L. C. Kjelstrom, oral communication, 1981). Values of variables A and F were determined from U. S. Geological Survey 1:250,000 scale topographic maps. Precipitation was determined from a long-term isohyetal map adjusted to the calibration year.

Underflow from Medicine Lodge Creek watershed was determined from estimates of creek seepage within the small alluvial valley upstream from the study boundary. Underflow was small relative to creek flows due to the small size of the alluvial fan (Stearns, Bryan, and Crandall, 1939, plate 3).

Water yields from Beaver and Camas Creeks were determined from precipitation data and estimates of evapotranspiration. Evapotranspiration from forested areas was estimated to be 17 inches annually as reported by Whitehead (1978, p. 30). Annual precipitation in forested areas in excess of 17 inches was considered to contribute to water yield. Water yield from rangeland was determined by subtracting the available water holding capacity of the root zone from winter precipitation.

Minor seasonal variations in the ground-water gradient near fixed-flux boundaries indicate that underflow remained relatively constant throughout the year. Annual underflow was therefore distributed evenly among timesteps and among boundary nodes within each watershed.

Fixed-head boundaries establish the altitude of the water table at each boundary node for each time interval. Each node of the fixed-head boundary was assigned a unique initial head. Boundaries were divided into 31 reaches, each having a series of head changes between timesteps. The head at a given node in a given timestep is determined from the initial head of the node and the head changes for that reach in which the node is located.

Leakage from the Saturated Alluvium

Leakage from an overlying saturated zone is calculated within the model as a product of the vertical hydraulic conductivity and the head difference between the modeled aquifer and the constant head leaky layer.

A series of constant head values representing each node in the saturated alluvium, outlined in figure 3, was determined from depth to water measurements, drillers′ logs, and water levels reported by Ralston and Chapman (1969). To determine leakage in the model it was necessary to estimate the difference between the water level in the alluvium and that in the underlying regional aquifer. Uncertainty of water levels in the alluvium prevented mapping the water-table, but in most places uncertainty was small relative to the head difference between aquifers. Errors in model calculated leakage were therefore expected to be small. The water table in the alluvium is not constant and in most areas head variations are small relative to head differences between the shallow and regional ground-water systems.

Vertical hydraulic conductivity between the saturated alluvium and the regional aquifer was estimated for each node based on lithology as reported in drillers′ logs.

Surface Recharge and Discharge

The "Recharge Program" was used to determine flux to the regional aquifer and to the alluvium overlying the regional aquifer. The relative magnitudes of annual (April, 1980 through March, 1981) surface recharge and discharge components are shown in figure 11. Annual distribution of flux is illustrated in figure 12. Surface recharge above the saturated alluvium, outlined in figure 3, does not directly affect the regional aquifer, and is therefore not included in figures 11 and 12.

Seepage losses from Medicine Lodge, Beaver, and Camas creeks within the study area (fig. 1) totaled about 89,000 acre-feet during the calibration year. Seepage was determined for reaches between gaging stations and study area boundaries. Seepage was determined as the difference between inflow to a reach and the sum of the outflow and irrigation diversions within that reach.

Irrigation pumping accounts for most of the ground-water discharge (excluding underflow) in the study area. Approximately 42 percent of the 143,000 acres irrigated is serviced by canal companies. All of the companies depend primarily upon ground water for their supply although some supplement with surface-water diversions. Ground-water pumping by canal companies is concentrated in the area north of Mud Lake where water levels are close to land surface. Withdrawals by canal companies during the calibration year totaled about 136,000 acre-feet. In addition, areas supplied by local ground-water pumping (fig. 13) withdrew sufficient water to meet crop consumptive use. Irrigated areas were determined from "Water Related Land Use" maps published by the Idaho Department of Water Resources (1978) and from U. S. Geological Survey U2 infrared aerial photos. Canal company withdrawals were determined from flow measurements, watermaster records, and pump power consumption records.

Evapotranspiration and precipitation were determined from data

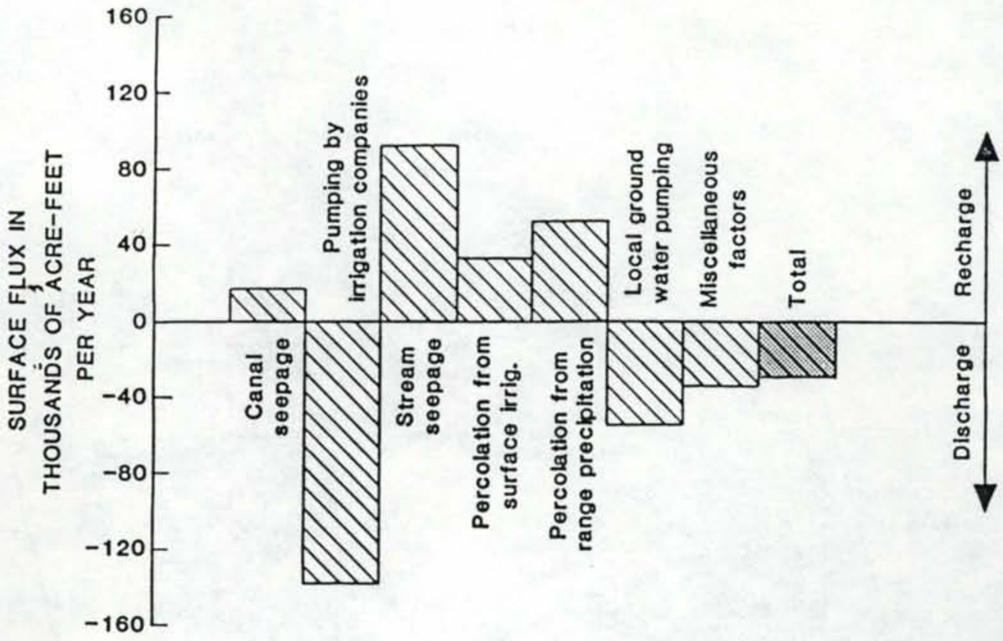

FIGURE 11.--Relative magnitudes of surface recharge and discharge components.

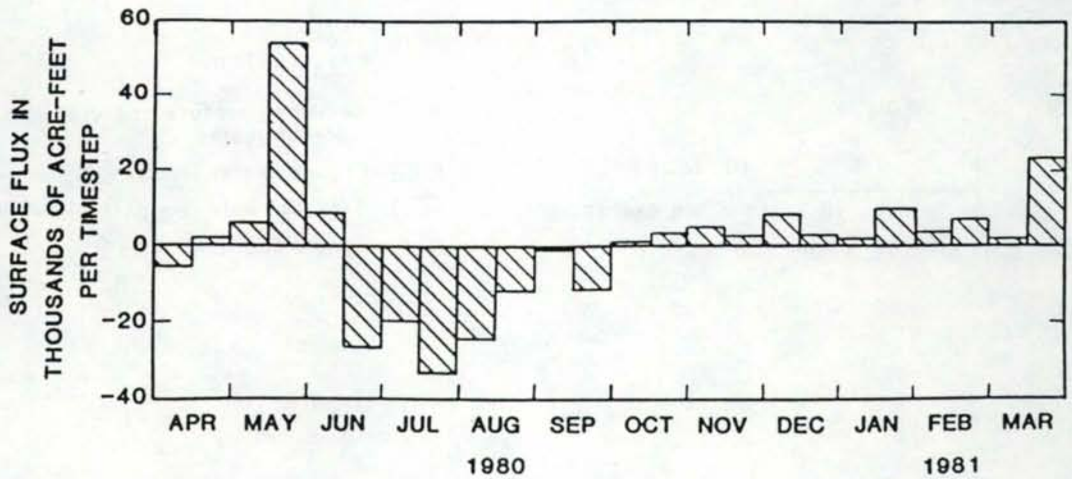

FIGURE 12.--Time distribution of surface recharge and discharge.

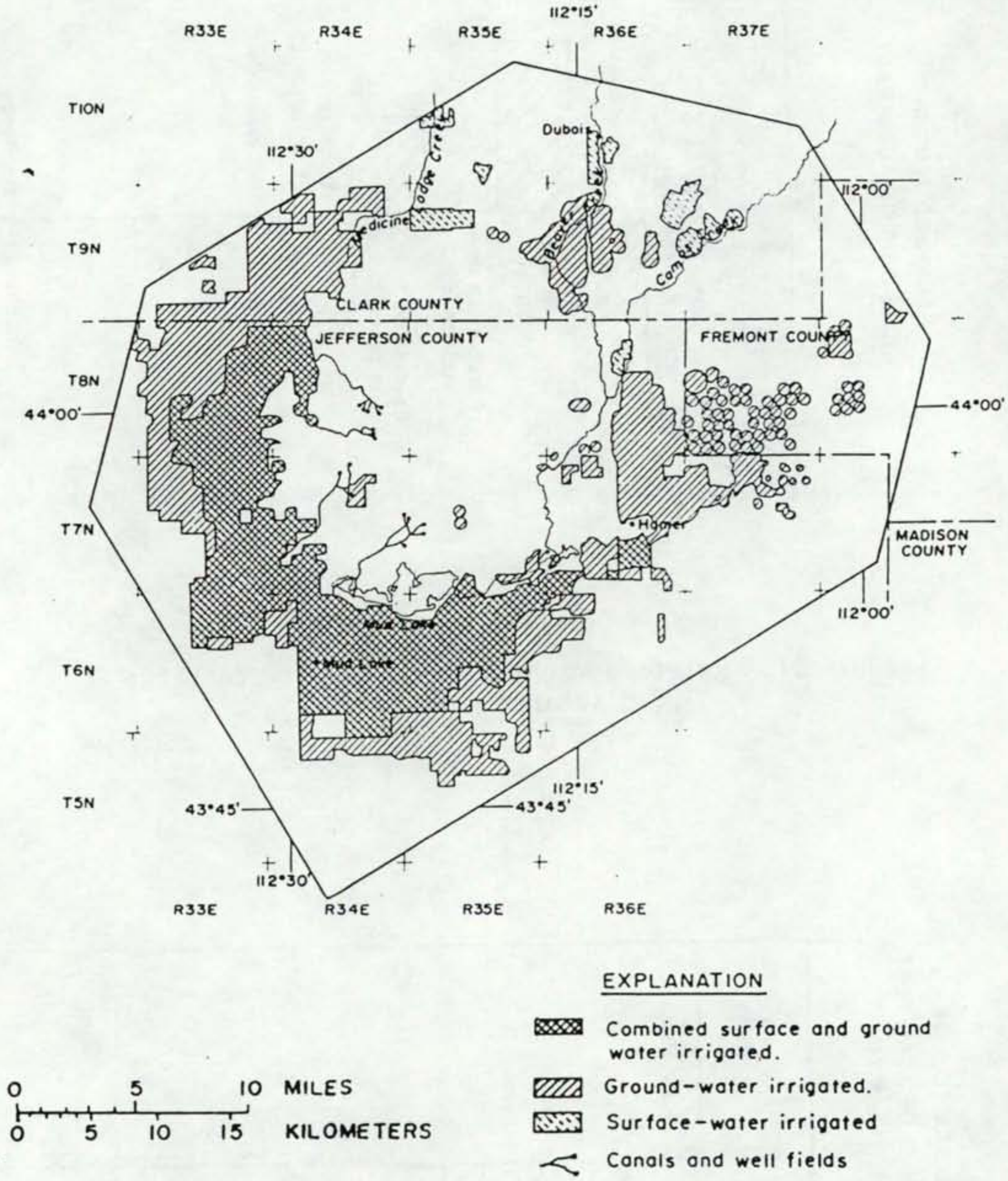

FIGURE 13.--Distribution of irrigated lands in 1980.

collected at weather stations in and around the study area (fig. 14). Evapotranspiration was determined using a modified version of the Blaney-Criddle method (Doorenbos and Pruitt, 1977). The method requires data on mean daily temperature, solar radiation, relative humidity, and wind conditions. These data were selectively taken from weather stations at Dubois, Hamer, Terreton, Rexburg, and Idaho Falls, based on availability and on the degree to which the data represent the conditions of the study area. Also shown in figure 14 are weather stations at Monida and Lakeview used in calculation of water yield in tributary valleys. Crop coefficients given by Wright (1981) were used to determine evapotranspiration for individual crops. Precipitation in the study area was determined from data at Dubois, Hamer, and Terreton stations.

Canal seepage totaled 16,800 acre-feet to the regional aquifer and 10,000 acre-feet to the shallow ground-water system during the calibration year. The total is equivalent to 16 percent of canal inflow. Seepage volumes were calculated as the product of seepage rates and canal areas. Rates were determined from flow measurements, soil properties, and from previous investigations by Worstell (1976) and Netz (1980). Data on soil properties were obtained from the Jefferson County Soil Survey (Jorgenson, 1979). Canal areas were determined from U. S. Department of Agriculture, Agricultural Stabilization and Conservation Service 1:15,840 scale aerial photos and from U. S. Geological Survey 1:24,000 scale topographic maps.

Seepage losses from Mud Lake were estimated to be 16,400 acre-feet during the calibration year. Lake losses recharged only the shallow ground-water system in the alluvium. Seepage from Mud Lake and from the

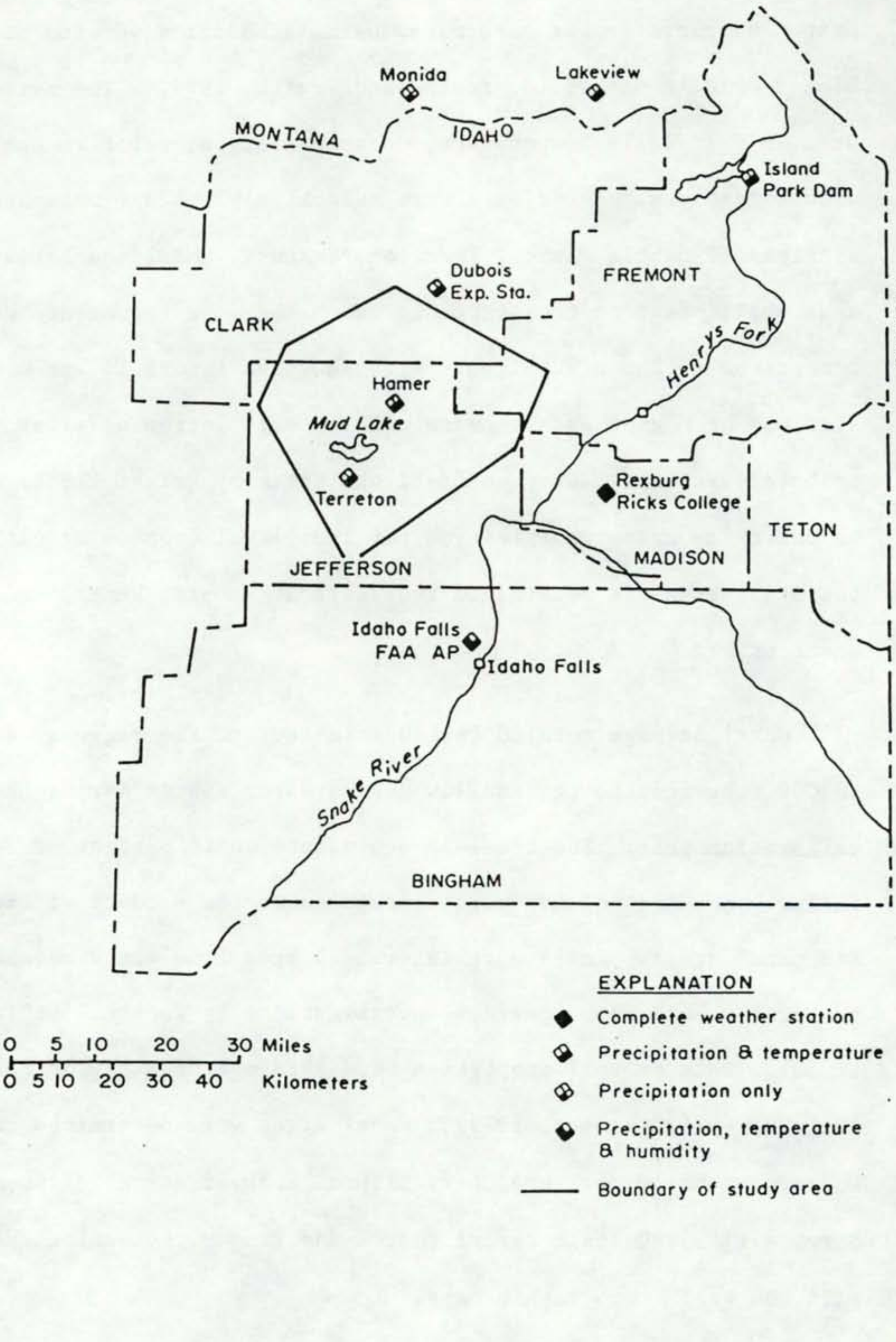

FIGURE 14.--Location of weather stations.

Camas Wildlife Refuge are based on a water balance determined from watermaster records and diversion data collected by the Refuge staff.

Steady-State Calibration

The model was run in a steady-state mode to eliminate effects of storage while refining estimates of transmissivity. A steady-state simulation requires the following assumptions: 1) water levels at the beginning and end of the period are identical, 2) changes in water level within the simulation period do not significantly affect transmissivity, and 3) surface flux and leakage are constant throughout the period.

The first condition was met by selecting a simulation period of 1-year, April 1, 1980 through March 31, 1981. The maximum measured head difference between the beginning and end of the period was 7.7 feet with most differences being less than 1 foot. Aquifer thickness is sufficiently large in much of the area so that water-table fluctuations of this magnitude do not substantially alter transmissivity. Surface flux however, is highly variable in different timesteps and locations. The assumption of constant flux with time may therefore cause distortion of the simulated water table.

Simulations in the steady-state mode failed to produce an adequate calibration of transmissivity. Repeated adjustments of transmissivity did not result in simulated heads which closely matched measured head values. The procedure was abandoned, and original estimates of transmissivity were used in the transient simulation.

Transient Simulation

Transient conditions were simulated in 15.2 day timesteps over the 1-year calibration period. The steady-state assumptions that the water table, recharge, and discharge remain constant in time are not required in a transient simulation. Transient simulations may therefore provide a better means of calibration than possible in a steady-state simulation. Transient simulations, however, require calibration of storage coefficient in addition to parameters calibrated in steady-state.

Aquifer parameters were adjusted to minimize the sum of squares of the differences between the simulated water table and the water table defined using depth to water measurements. The sum of squares was calculated from differences determined at every node for timesteps 9, 15, and 24, corresponding to the approximate dates of water level measurements in August, November, and April. The best calibration is considered to be that which produces the minimum sum of squares at each timestep.

The simulated water table was observed to be generally more sensitive to changes in transmissivity than to changes in storage coefficient. Therefore priority was placed on adjustment of transmissivity. Sensitivity analysis of each parameter was not performed. Values of surface flux, underflow, and leakage were based on hydrologic data collected during the calibration period and therefore were less subject to adjustment.

Transmissivity was first adjusted manually to reduce positive or negative deviations occurring consistently over a large area. The automatic calibration routine within the model (deSonneville, 1974) was then employed to further refine the distribution. The automatic routine, however, resulted in values of transmissivity which were highly irregular from node to node. These values were therefore initially used only as a guide for making further manual changes. A series of transmissivity adjustments were made until no further reduction in the sum of squares was obtained. The automatic calibration routine was then applied to the resulting transmissivity values in a 10-year transient simulation.

The automatic calibration routine produced fewer irregularities in transmissivity and a lower sum of squares when the 1-year simulation was cycled over a period of 10 years. The calculated water table at the end of each year was used as the initial surface for the succeeding year. Calibration was based on heads calculated during the tenth year.

The transmissivity distribution resulting from the final run of the automatic calibration was manually adjusted to improve continuity of values between nodes. A small increase in deviations between simulated and measured heads resulted from the smoothing.

Automatic calibration of the storage coefficient also resulted in erratic values and failed to greatly improve the simulation. Results were therefore used as a guide for manual adjustment of nodal storage coefficients.

Seasonal variations in the water table were most difficult to simulate in the area immediately northwest of Mud Lake. None of the combinations of values of storage coefficients and transmissivities simulated the 56-foot decline in the water table that occurred in the summer of 1980 (fig. 7). In addition to the numerous combinations of transmissivities and storage coefficients that were evaluated, saturated thickness and leakage were also adjusted. No significant improvement in the simulation was obtained.

Calibrated values of transmissivity and storage coefficient resulted in a 1-year simulation which more closely approximated the water table during fall and spring than during the summer. The accuracy of the simulation during each season is shown statistically in table 2. The "Sum of Squares" is the sum of the squares of the absolute value of the difference between simulated and measured (interpolated from measured) head values at each node. The "Mean Error" is the arithmetic average of the absolute value of the differences at each node. The "Standard Deviation" represents the standard deviation of the difference between the mean error and the error at each individual node.

TABLE 2. Simulation statistics resulting from calibrated parameter values.

Timestep	Corresponding date	Sum of squares (ft^2)	Mean error (ft)	Standard deviation (ft)
9	August 15	4.975×10^4	4.1	6.2
15	November 15	1.199×10^4	2.6	2.6
24	April 1	1.514×10^4	2.8	3.0
Total		7.688×10^4	---	---
Average		---	3.2	---

The areal distribution of the differences between the final simulated and measured water tables are shown in figures 15, 16, and 17; for August, November, and April (timesteps 9,15, and 24). The greatest differences occur in August in an area immediately northwest of Mud Lake as previously discussed. Poor simulation of this area is the primary reason for the large mean error and sum of squares in timestep #9.

Calibration Results

Transmissivities calculated from model calibration are illustrated as lines of equal transmissivity in figure 18. Transmissivity contours show a general relation to the water table illustrated in figure 6. The band of steep hydraulic gradient results from a corresponding band of low transmissivity ($2x10^3$ to $5x10^4$ ft^2/day). On either side of this band transmissivities generally exceed $1x10^6$ ft^2/day. The low transmissivity near Medicine Lodge creek (less than $5x10^4$ ft^2/day) reflects thinness of the aquifer and a relatively high percentage of sediments as reported in drillers′ logs.

The accuracy of calculated transmissivities is reflected by the head differences shown in figures 15, 16, and 17. Areas where large differences between simulated and measured (reference) water levels appear suggest less accurate calibration was achieved in these areas. Large differences occuring at one or more of the calibration timesteps indicate that the model is not representative of the real system in these areas.

Accuracy of the calibration is also a function of the accuracy of water levels at each node which were interpolated between measured wells. Simulated water levels were compared to interpolated values at

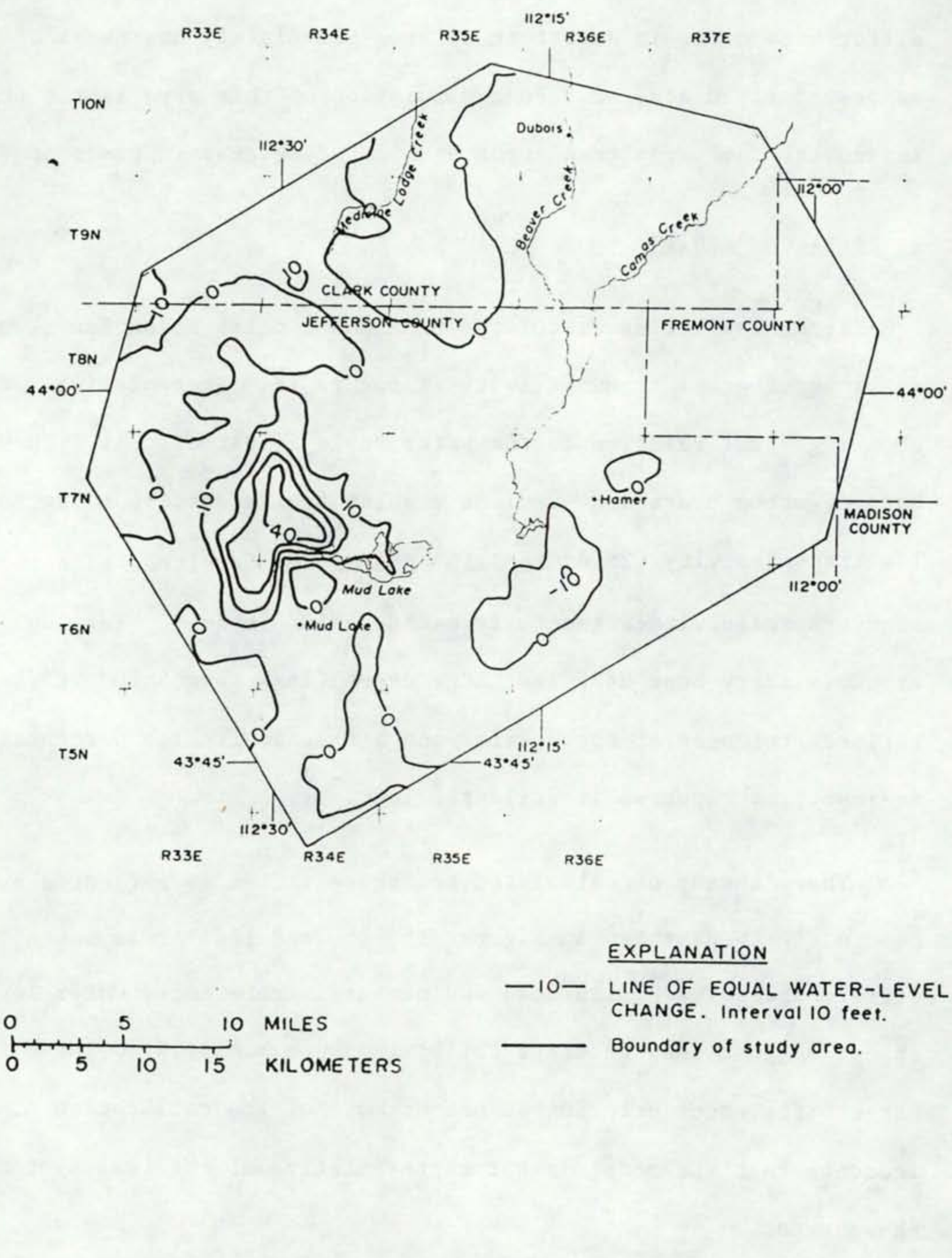

FIGURE 15.--Differences between measured and simulated water tables, August, 1980.

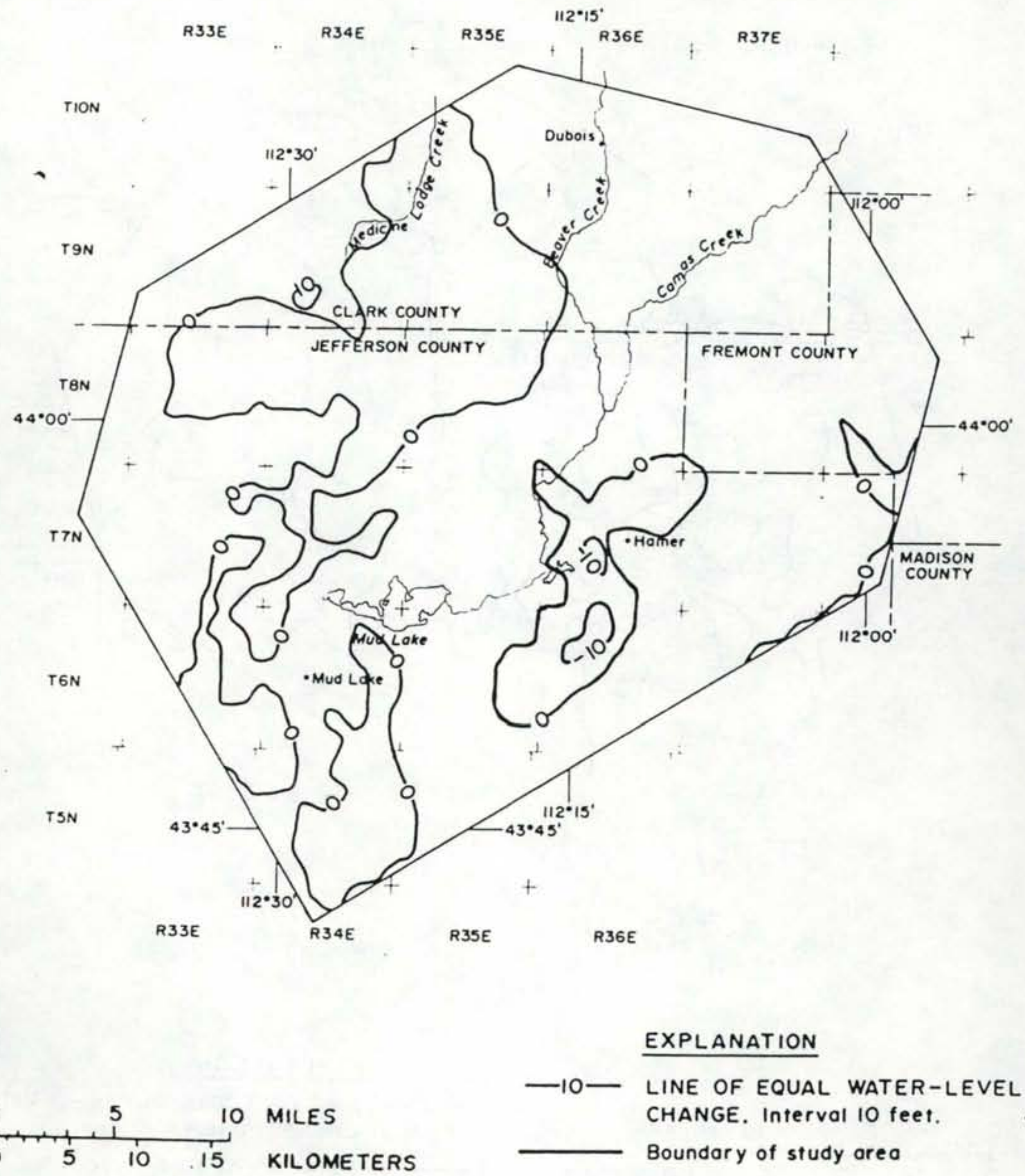

FIGURE 16.--Differences between measured and simulated water tables, November, 1980.

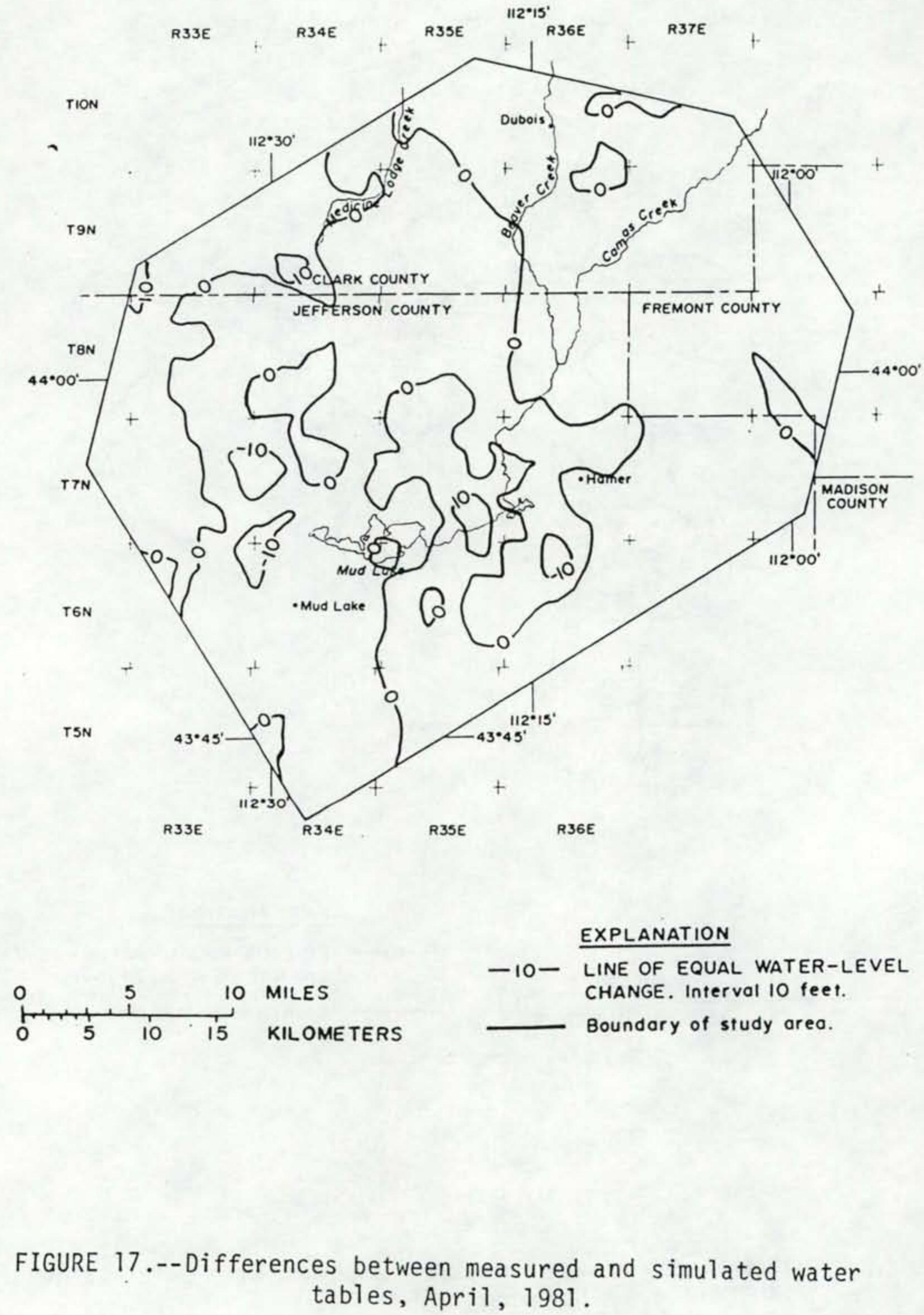

FIGURE 17.--Differences between measured and simulated water tables, April, 1981.

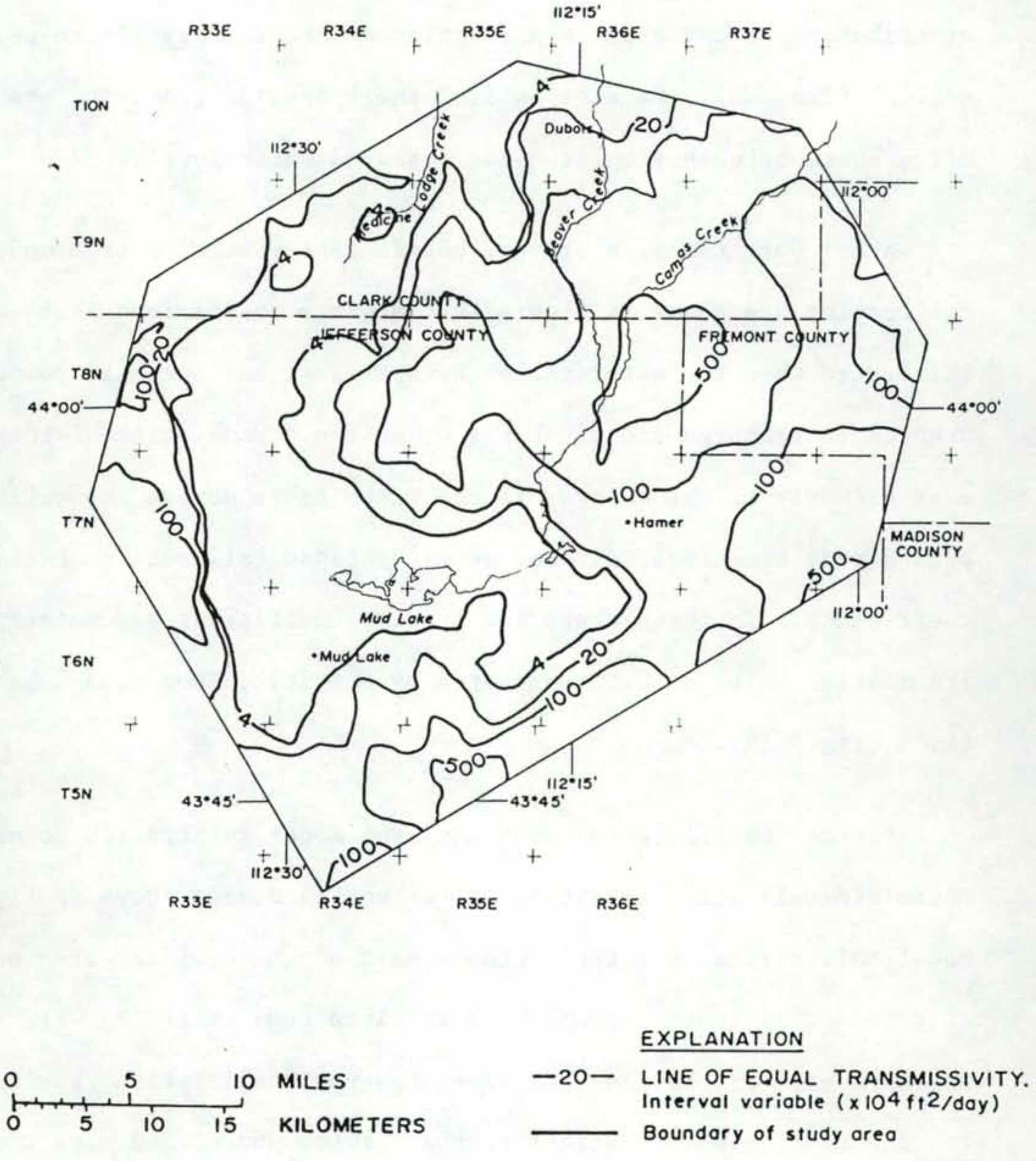

FIGURE 18.--Distribution of transmissivity resulting from model calibration.

each node, not to measured water levels. The interpolated water levels must therefore be representative of the real system to produce a valid model calibration. Accuracy of calibration of the transmissivity distribution in any area is a function of the density of the well network (fig. 5), the steepness of the hydraulic gradient, and the differences between simulated and measured water levels.

Areal variations in storage coefficient resulting from model calibration are shown in figure 19. Storage coefficient is best calibrated when the water table shows a large and definite response to changes in recharge and discharge with time. Some parts of the study area had only slight changes in the water table during the calibration period and, therefore, did not permit refined calibration of the storage coefficient. In these areas the storage coefficient was maintained at its initial value of 0.22 suggested by Norvitch, Thomas, and Madison (1969, fig. 12).

Storage coefficients resulting from model calibration do not correlate well with the extent of the confined area shown in figure 4. Model calibration over the southern half of the confined area produced storage coefficients typical of a confined aquifer (10^{-4}). In the northern half of the confined area, however, calibration resulted in a storage coefficient identical to that of the unconfined area to the east (10^{-1}). It is not known whether the different values of storage coefficient represent variations in aquifer conditions, type of aquifer, or calibration error.

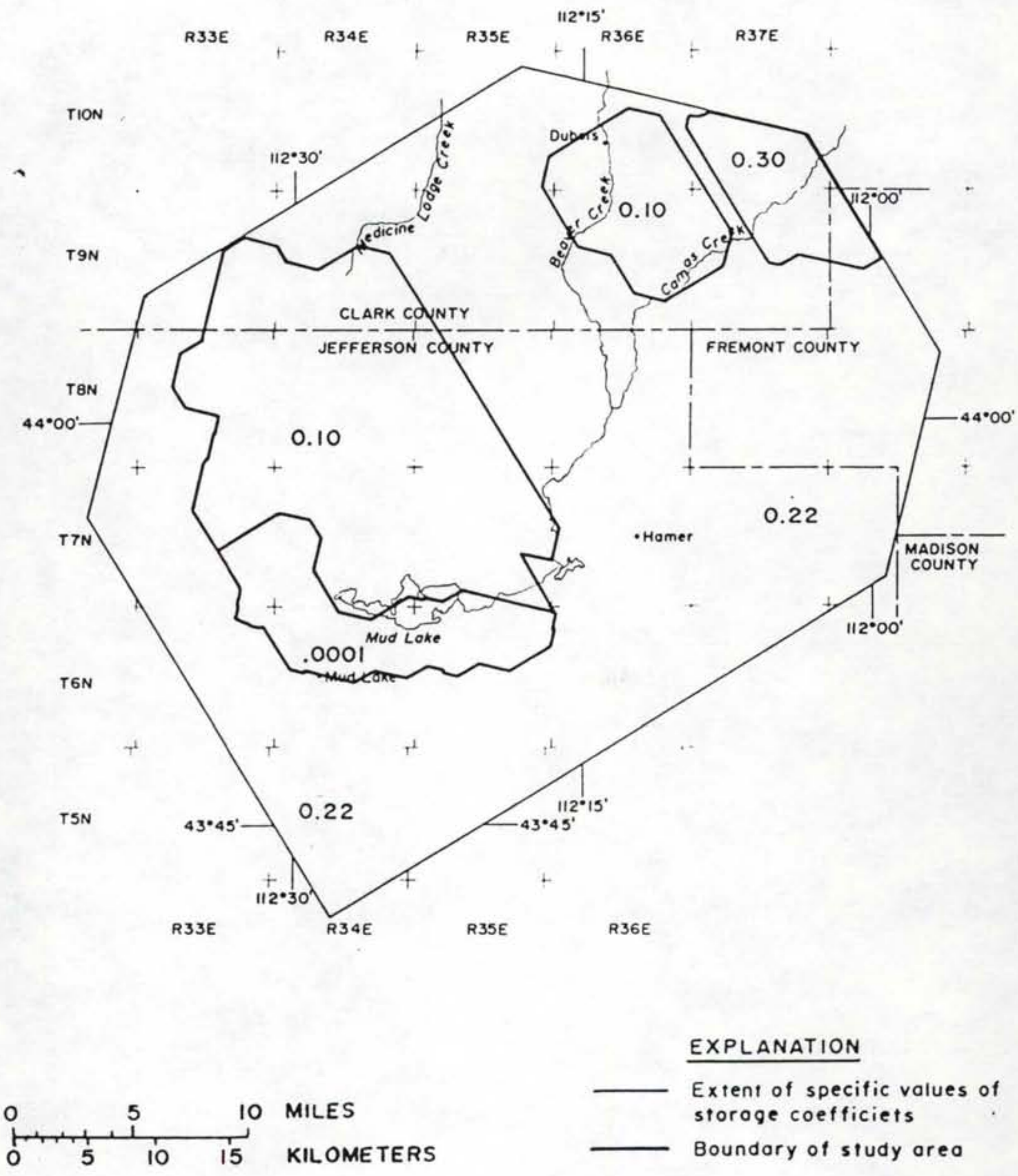

FIGURE 19.--Distribution of storage coefficient resulting from model calibration.

SIMULATION OF HYPOTHETICAL CHANGES IN WATER USE

The calibrated model was used to simulate long-term aquifer response under 3 sets of conditions: 1) continuation of current irrigation practices under normal climatic conditions (Normal), 2) a hypothetical increase in irrigated area (Case #1), and 3) hypothetical application of artificial recharge (Case #2). The details of each simulation are described in the following sections. Validity of the model calibration is implied by the continuity of historic water-level trends and water levels resulting from long-term simulation of current irrigation practices under normal hydrologic conditions. The long-term simulation of normal conditions also provides a basis for evaluation of the effects of the two hypothetical simulations. All simulations were conducted for periods of 20 years to observe long-term effects.

Normal Simulation

Current irrigation practices combined with long-term average meteorologic and streamflow conditions were used to determine surface flux for what is referred to as the normal simulation. Surface flux used in model calibration was determined specifically for the 1-year calibration period and does not represent long-term average conditions. In general, the average year is drier than the calibration year. During a normal year ground-water withdrawals exceed recharage from surface-water sources by 192,000 acre-feet, excluding recharge from leakage and boundary underflow. During the calibration year withdrawals exceeded recharge by only 30,000 acre-feet. The difference of 162,000 acre-feet represents approximately 45 percent of the combined annual recharge from underflow and leakage. The remaining 55 percent discharges across the study boundaries to the southwest. Boundary conditions and leakage from the ground-water system in the alluvium remained the same as during the calibration simulation.

Hydrographs at 8 nodes, located in figure 20, are presented in figures 21 and 22. Simulated hydrographs that represent water levels under normal hydrologic conditions are labeled "Normal". Historical hydrographs included on 6 of the plots, are for observation wells identified on each graph. The wells are located within the cell surrounding the given node. No historical hydrographs were available at nodes (23,7) and (29,11). Historical water levels were adjusted to represent the nearest node point. Magnitude of the adjustment was dependent upon hydraulic gradient and the distance between the well and node point.

Simulated water-level trends are generally consistent with historical trends. Historical hydrographs indicate a general water-level decline since the mid-seventies. Simulated normal hydrographs, beginning in 1980, continue the downward trend, sometimes at an accelerated rate. Normal water levels seem to stabilize or begin a gradual climb within about 10 years after the start of the simulation. An exception is noted at node (8,9) where water levels continue to decline at the nearly constant rate of 0.3 feet per year. This decline occurs despite the stabilizing influence of a fixed head boundary only 4 miles away. Seasonal fluctuations of simulated water levels are also consistent with the historical parts except at node (8,9). The amplitude of simulated fluctuations at node (8,9) are about 50 percent of the historical amplitude. The simulated normal hydrograph at node (23,7) indicates a nearly constant increase in water level of 0.35 feet

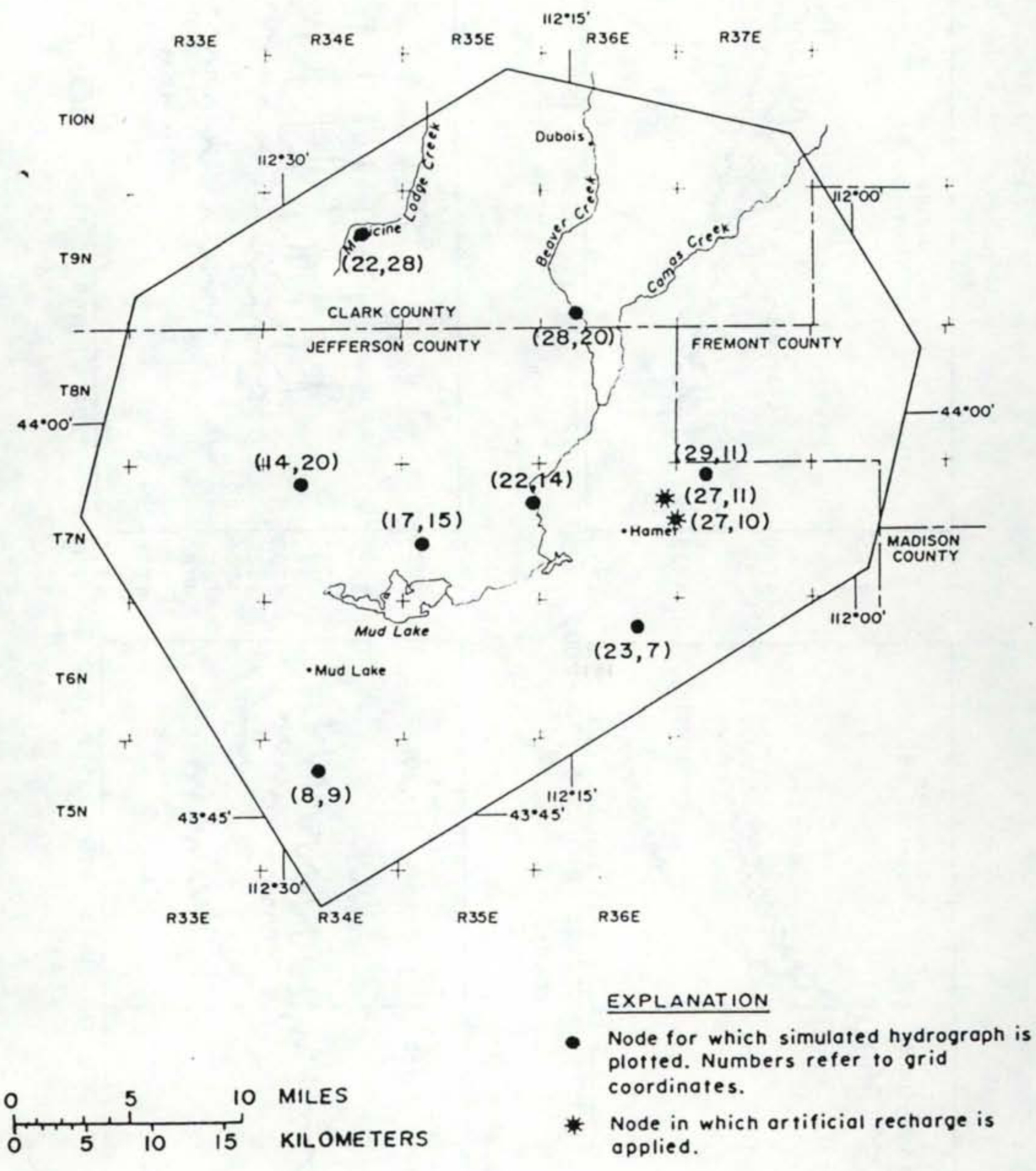

FIGURE 20.--Location of nodes for which hydrographs were plotted, and to which hypothetical recharge was applied.

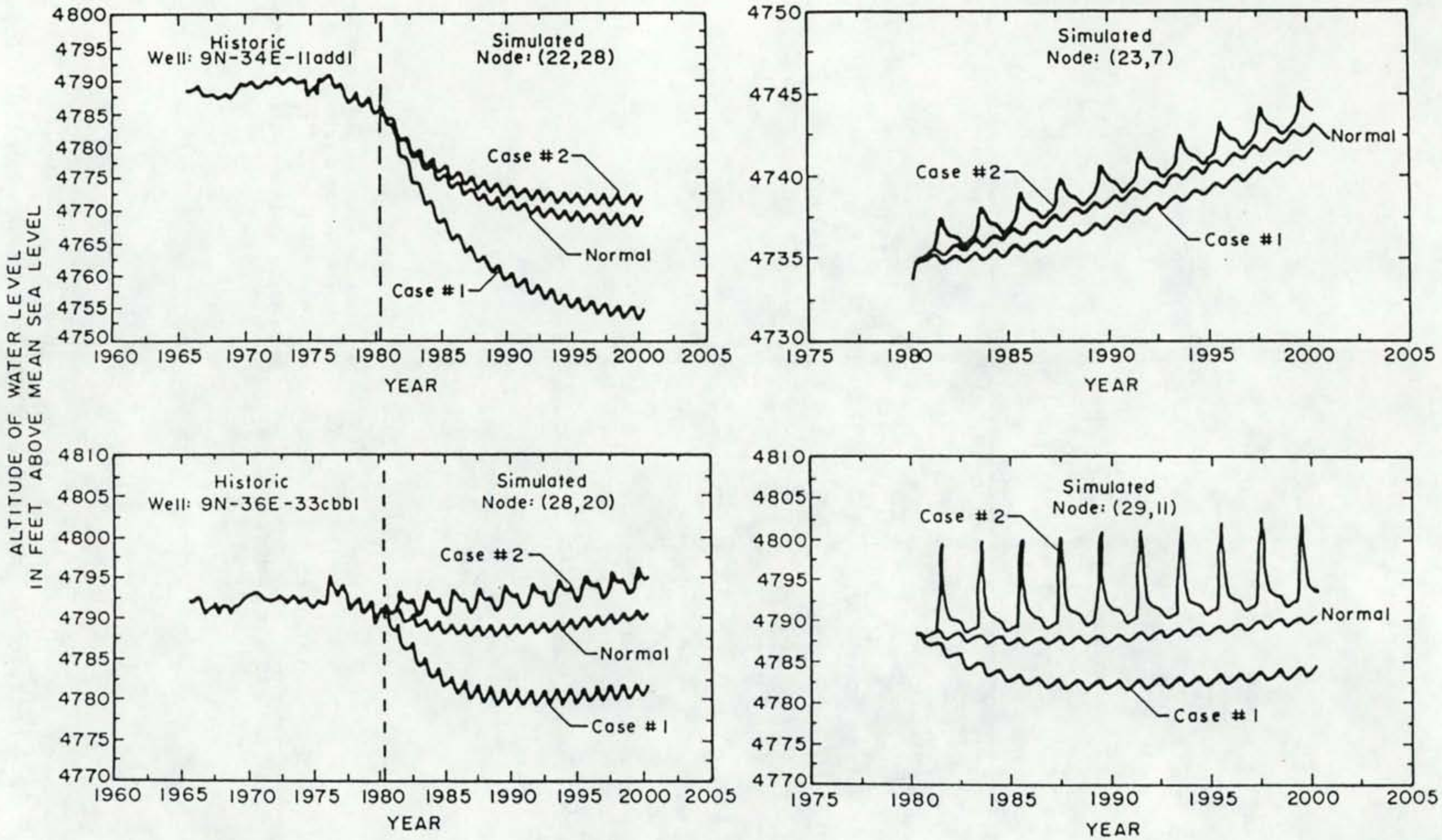

FIGURE 21.--Hydrographs of historical and simulated water levels at individual nodes.

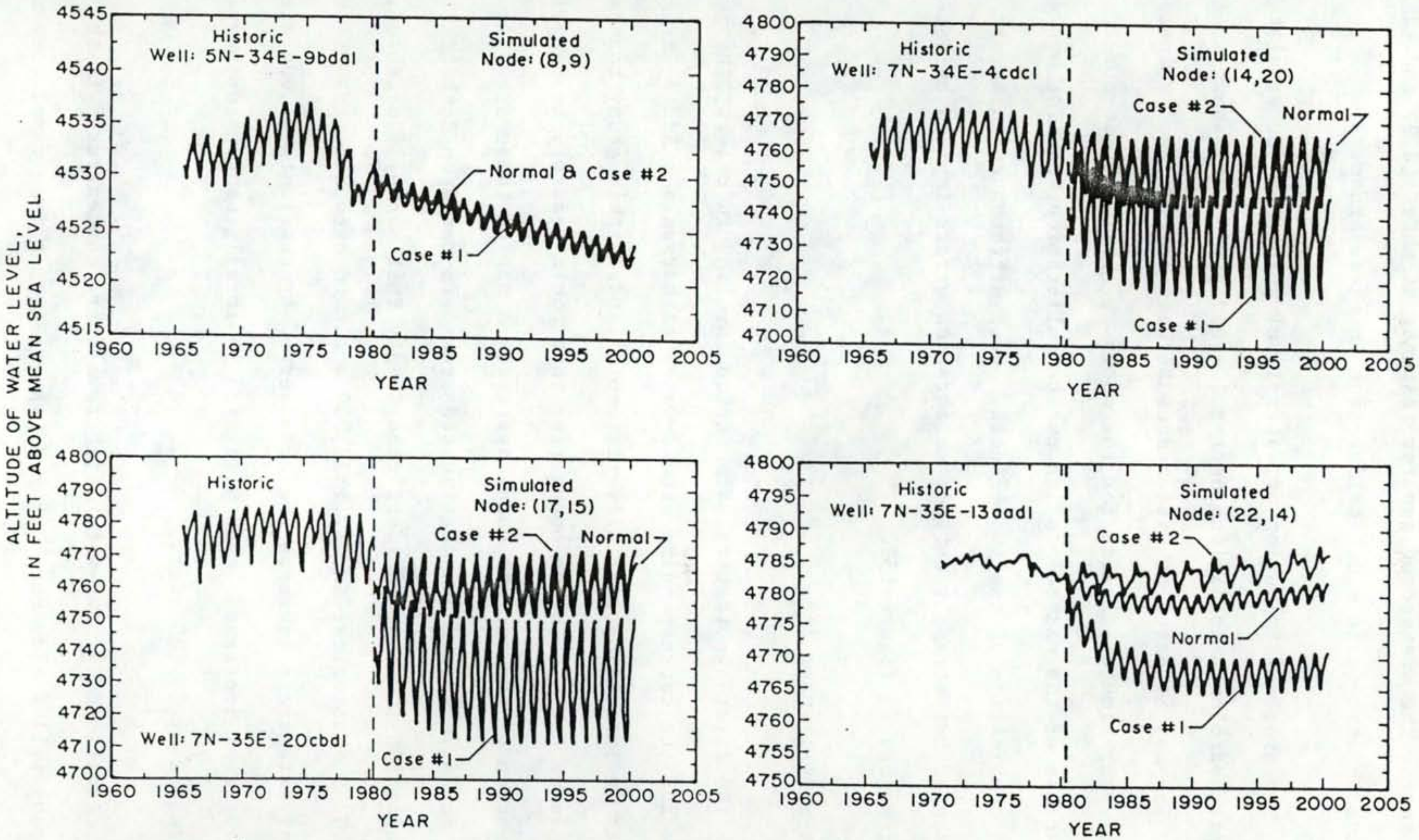

FIGURE 22.--Hydrographs of historical and simulated water levels at individual nodes.

per year. The unexpected aquifer response at nodes (8,9) and (23,7) may be due to inadequate model calibration in these areas.

The areal distribution of water-level changes under simulated normal hydrologic conditions over a 20-year period are shown in figure 23. The map is based on differences between initial head values and final year simulated heads for the month of April. It is not known if the cause of the changes is inaccurate calibration, aquifer overdraft, or a combination of the two effects. Head differences in August are not presented but strongly reflect calibration errors implied by water-level differences in figure 15.

Hypothetical Case #1

The first hypothetical simulation involves increasing the irrigated area and the corresponding ground-water withdrawals. Added irrigated areas include all potentially irrigable lands within study boundaries, and lands not within the boundaries that could feasibly be supplied by extension of the current canal systems. A map of these areas is presented in figure 24. Designated areas are based on Haskett (1979, drawing 886-100-56), the Jefferson County Soil Survey (Jorgenson, 1979), and U. S. Geological Survey 1:24,000 scale topographic maps of the area. Canal extensions increase the area served by canals to 61,000 acres, a 130 percent increase. Area irrigated by local ground-water pumping is increased by 28,600 acres.

Additional ground-water withdrawals are concentrated in the canal company well fields north of Mud Lake. Pumping from canal company well fields was not increased in proportion to area due to consideration of canal seepage and current use of surface water supplies. Pumping in

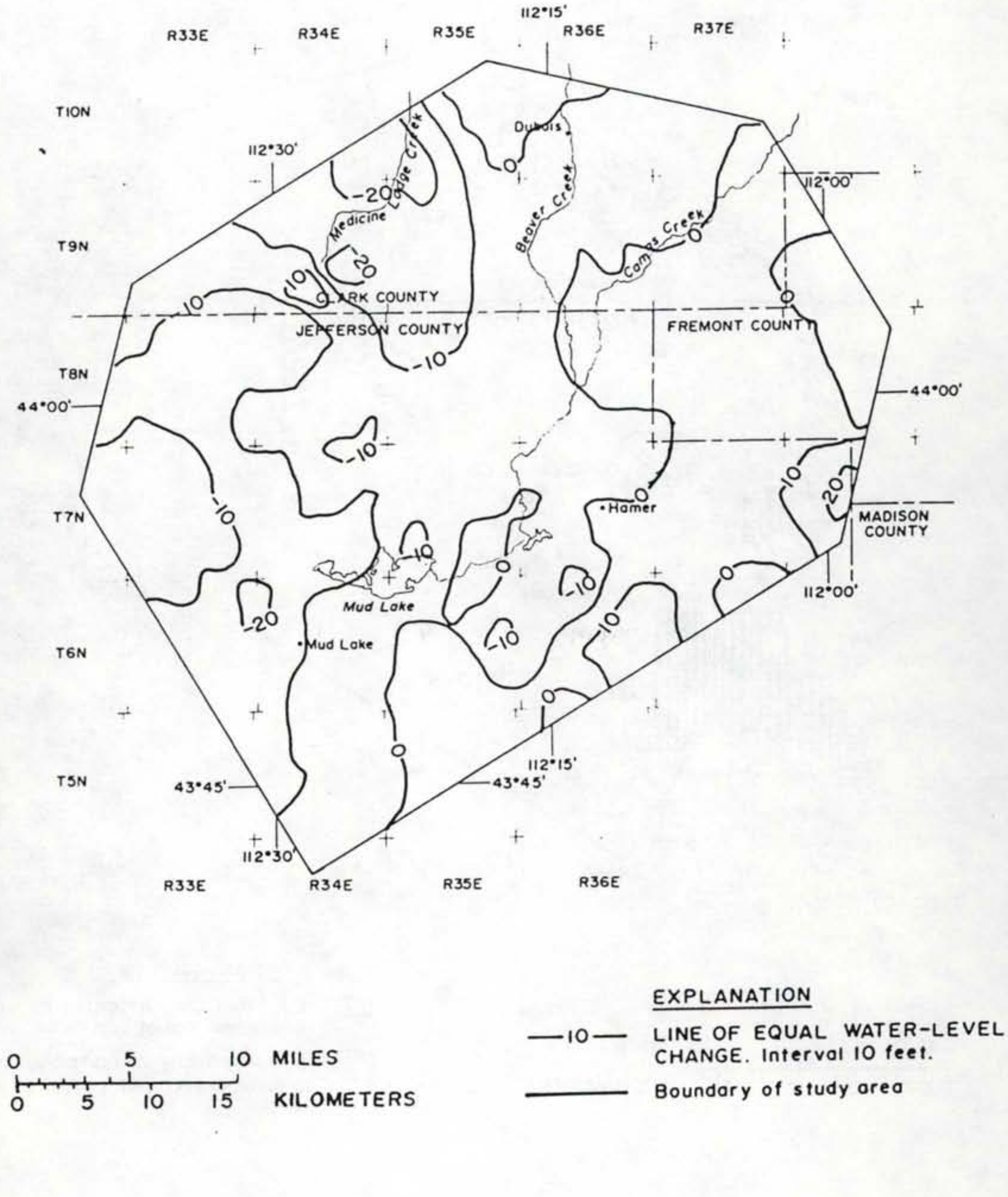

FIGURE 23.--Changes in the April water levels resulting from a 20-year normal simulation, relative to initial values.

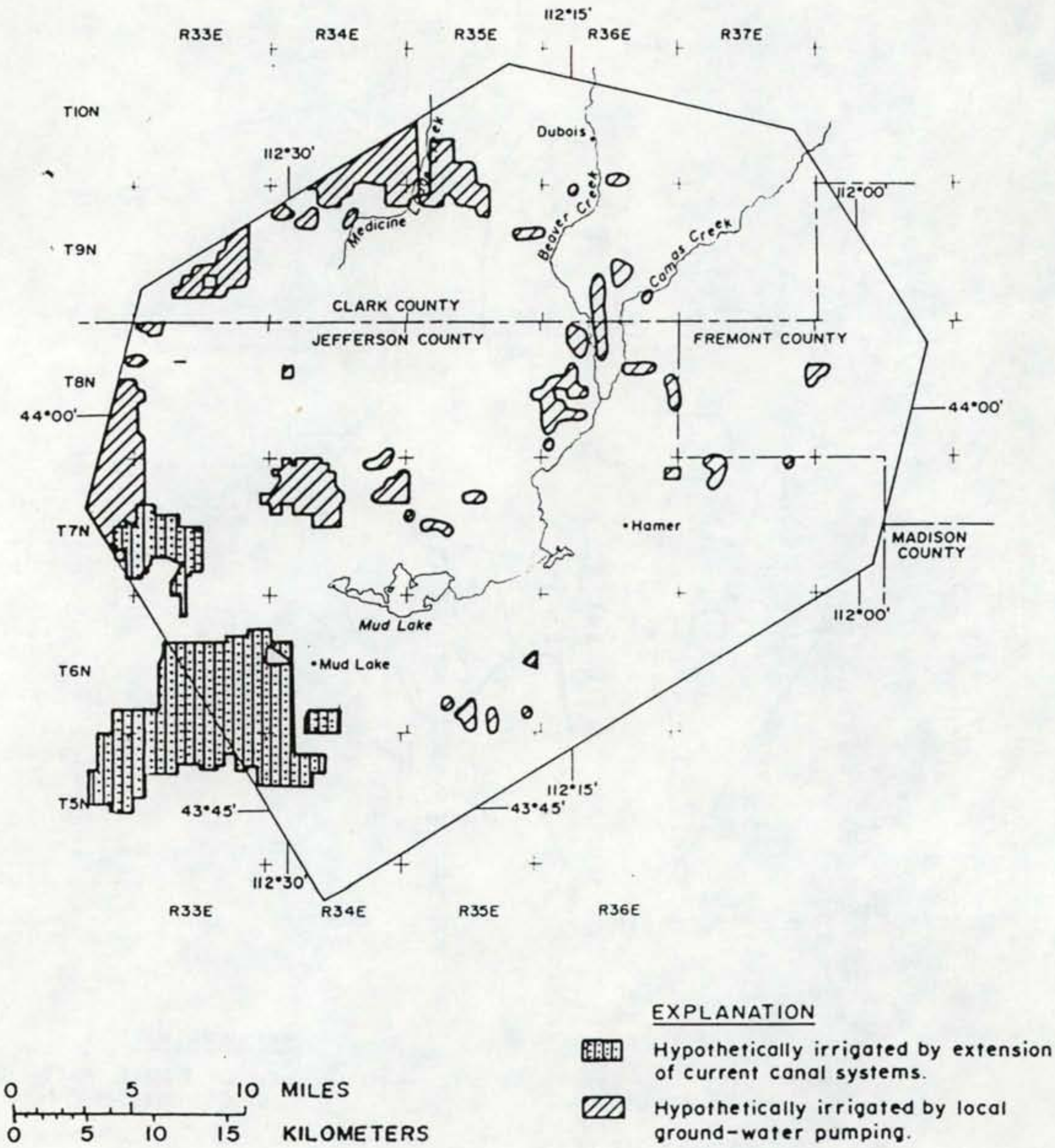

FIGURE 24.--Potentially irrigable lands included in hypothetical case #1.

cells containing local ground-water irrigation was increased in proportion to the area irrigated. Withdrawals were also increased throughout areas of local ground-water pumping. Pumping from the canal company wells at nodes (15,19), (15,20), (16,20), was hypothetically increased 76 percent of normal, resulting in total annual withdrawals of about 53,100 acre-feet. Pumping at nodes (17,16), (17,17), (21,12), was increased 154 percent of normal, resulting in annual withdrawals of about 171,200 acre-feet.

Water levels resulting from long-term simulation of hypothetical case #1 are compared to the those resulting from the long-term normal simulation. The areal distribution of water-level differences is shown in figures 25 and 26 for spring and summer. Increased irrigation pumping is responsible for the much larger decline of water levels in August. The greatest changes occur in nodes closest to well fields where increased pumping causes an additional 50 to 80 feet of drawdown. By the following spring, water levels recover to nearly 20 feet below the "normal" spring water level. Effects of increased pumping are least apparent in areas down-gradient from the band of low transmissivity shown in figure 18.

Hydrographs of simulated ground-water response at specific nodes are shown by lines labeled "Case #1" in figures 21 and 22. Effects of additional pumping are best illustrated by comparison to the normal simulation. Graphs indicate that at most locations water-level declines resulting from pumping stabilize after about 10 years. Changes in hydraulic gradients result in more flow to areas of increased withdrawals and less discharge across boundaries to the south and west. Seasonal water-table fluctuations are greatest near the well fields.

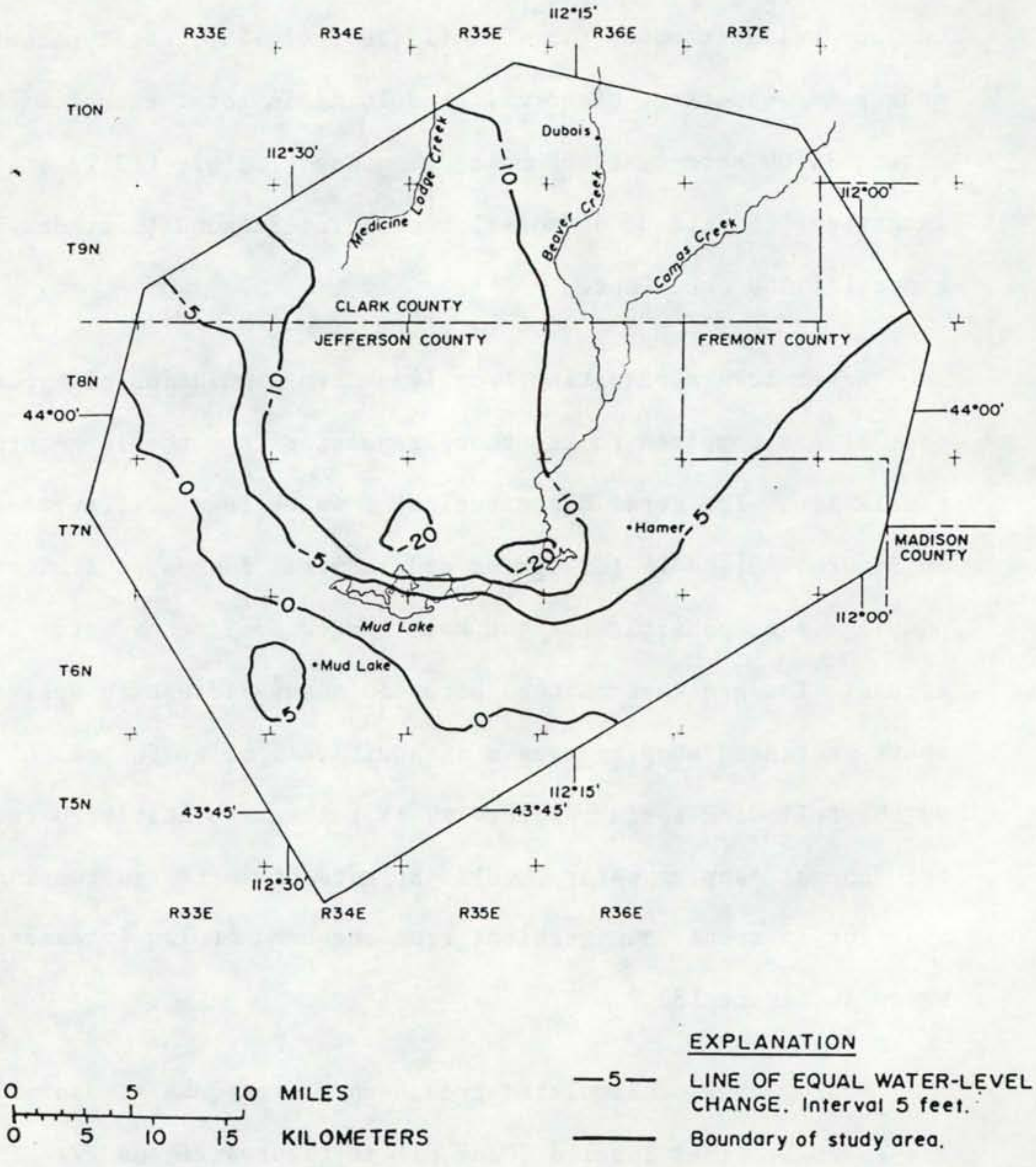

FIGURE 25.--April water-level differences resulting from the 20-year simulation of hypothetical case #1, relative to the 20-year normal simulation.

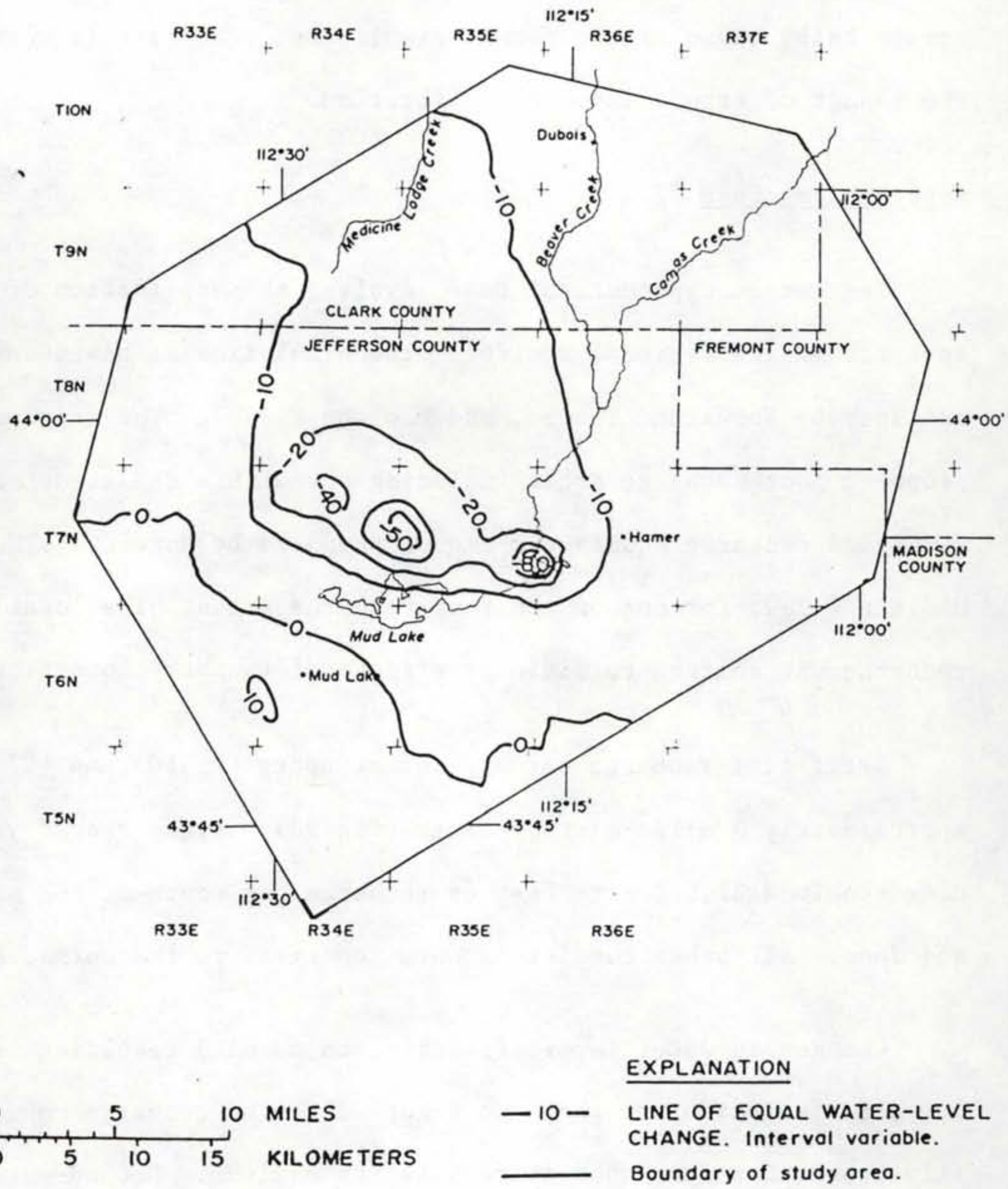

FIGURE 26.--August water-level differences resulting from the 20-year simulation of hypothetical case #1, relative to the 20-year normal simulation.

The drawdown at node (17,15), 1 mile from the well field is 38 feet, more than twice the historic fluctuation. In some locations water levels increase with time, however, in every instance water levels remain below those of the normal simulation. The rise is most likely the result of errors in model calibration.

Hypothetical Case #2

The second hypothetical case involves the application of artificial recharge to the regional aquifer. The simulation is based on a plan outlined by Norvitch, Thomas, and Madison (1969). The original plan proposed four recharge areas including one within the study area. Simulated recharge equals the amount proposed by Norvitch, Thomas, and Madison (1969) for one of the four recharge areas. The location of the recharge was shifted to minimize effects of boundary conditions.

Artificial recharge was applied at nodes (27,10) and (27,11) approximately 3 miles east of Hamer (fig 20). Every second year each node received 31,000 acre feet of recharge per month during April, May, and June. All other conditions were identical to the normal simulation.

Changes in water levels (relative to normal) resulting from a 20-year simulation are shown in figure 27. The recharge mound is illustrated for June when it reaches its maximum. Ground-water mounding would result in temporary flooding of depressions near the recharge area and several miles to the east.

Hydrographs of the simulated recharge (case #2) are shown in figures 21 and 22 along with the normal and case #1 simulations. The hydrograph for node (29,11) only 2 miles from the recharge area, shows

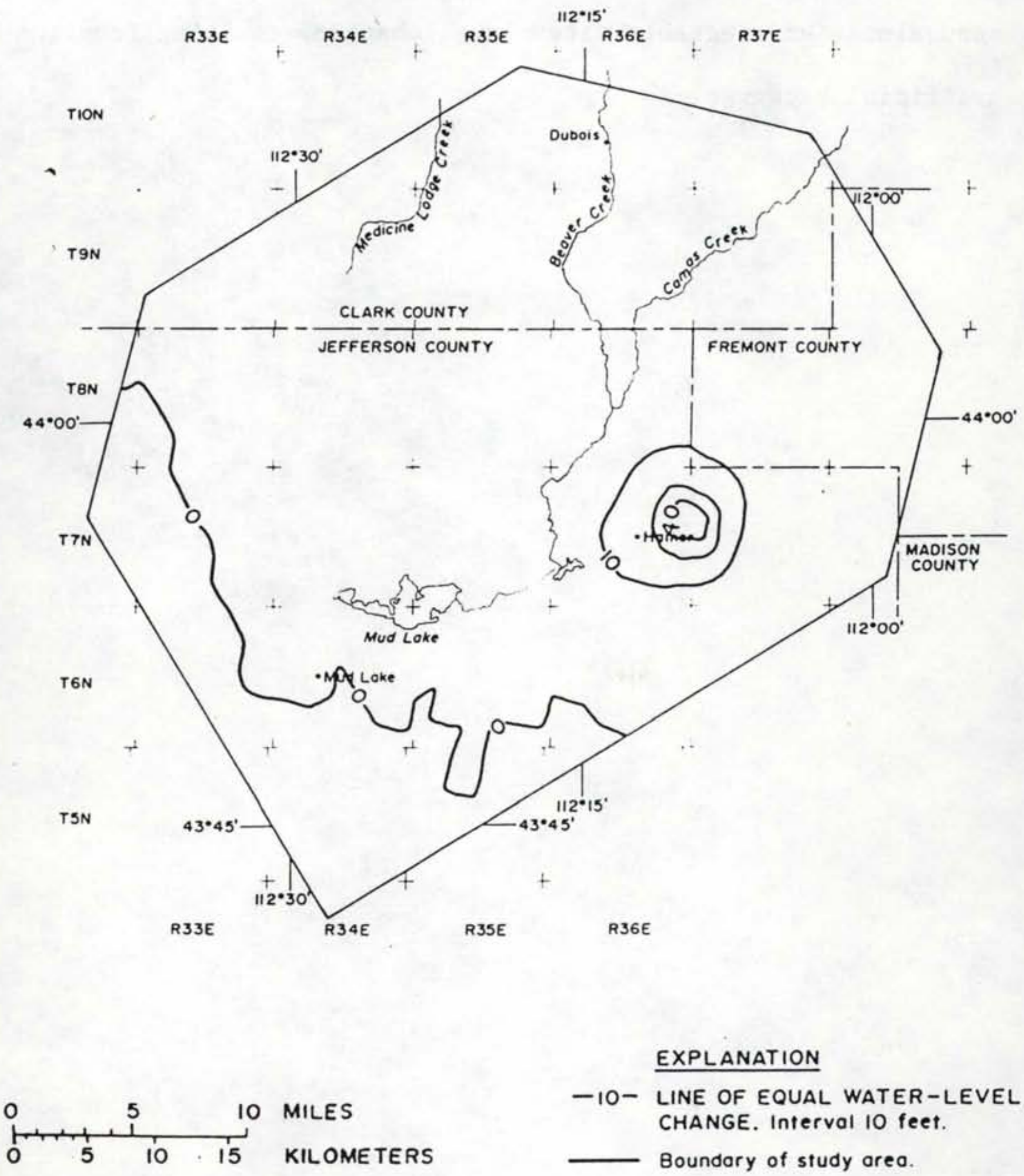

FIGURE 27.--June water-level differences resulting from the 20-year simulation of hypothetical case #2, relative to the 20-year normal simulation.

the greatest change. During periods of artificial recharge, water levels rise as much as 40 feet, however, within 2 years they return to approximately their original levels. Several of the other hydrographs show almost undetectable water-level changes resulting from the artificial recharge.

SUMMARY

The Snake Plain aquifer is the principle source of water in the Mud Lake area. In most of the study area it reacts to stress as an unconfined (water-table) system. Ground-water pumping is the major form of discharge (excluding boundary flow) in the area. Leakage from saturated alluvium recharges 113,000 acre-feet annually. Underflow into the area from the north and northwest is estimated at 247,000 acre-feet per year. Ground-water discharges across study boundaries to the south and west.

The water table has a hydraulic gradient of less than 10 feet per mile throughout most of the Mud Lake area, except in a southeasterly trending band, where the hydraulic gradient is 30 to 60 feet per mile. In most locations water levels vary less than 10 feet during a normal year. However, in an area about 7 miles west of Mud Lake, water levels decline more than 50 feet during the irrigation season.

A numerical ground-water flow model was used to calibrate transmissivities and storage coefficients at established grid locations. Calibration based on hydrologic data indicates that the band of steep hydraulic gradient results from a corresponding band of low transmissivity (less than 4×10^4 ft^2/day). Refined calibration of the storage coefficient was not achieved due to the small seasonal changes in water levels at many locations.

The calibrated model was used to simulate long-term (20 year) regional water levels for current water use conditions and two hypothetical situations. Simulation of current water use with normal hydrologic conditions results in a decline of water levels in most areas. The simulation is generally consistent with historical trends thereby indicating that a reasonably valid calibration has been achieved. Simulation of aquifer response to increased ground-water withdrawals for irrigation expansion results in further water-level declines. The effect is most pronounced near canal company well fields if current pumping rates are doubled . Simulated drawdowns increased by as much as 80 feet. The second hypothetical simulation predicted response to alternate year application of 186,000 acre-feet of artificial recharge over a 2 square-mile area east of Hamer. The simulation indicates a local increase in the water table of over 40 feet resulting in intermittent local flooding of depressions several miles to the east.

REFERENCES CITED

Barraclough, J. T., Lewis, B. D., and Jensen, R. G., 1981, Hydrologic conditions at the Idaho National Engineering Laboratory, Idaho emphasis: 1974-1978: U. S. Geological Survey Open-File Report 81-526, 77 p.

Brockway, C. E. and Grover., K. P., 1978, Evaluation of urbanization and changes in land use on the water resources of mountain valleys: Idaho Water Resources Research Institute, Technical Completion Report, Project B-038-IDA, 105 p.

Crosthwaite, E. G., 1973, A progress report on results of test-drilling and ground-water investigations of the Snake River Plain aquifer, southeastern Idaho. Part 1, Mud Lake region, 1969-1970: Idaho Department of Water Administration Water Information Bulletin No. 32, 60 p.

deSonneville, Joseph, 1972, Development of a mathematical groundwater model: Unpublished thesis on file with University of Idaho Library, 228 p.

deSonneville, Joseph, 1974, Development of a digital ground water model with applicaiton to aquifers in Idaho: Unpublished dissertation on file with University of Idaho Library, 226 p.

Doorenbos, J., and Pruitt, W. O., 1977, Guidelines for predicting crop water requirements: Food and Agriculture Organization of the United Nations Irrigation and Drainage Paper 24, pp. 3-7.

Haskett, G. I., and Hampton, L. D., 1979, Groundwater investigations for Salmon Falls division, Idaho and aspects of pumping from the Snake Plain aquifer: draft copy, U. S. Bureau of Reclamation, Pacific Northwest Region, pp. 22-24.

Johnson, G. S., and Brockway, C. E., 1983, Users manual for the University of Idaho ground-water model and Recharge program: University of Idaho Water and Energy Resources Research Institute, 126 p.

Jorgenson, Wendell, 1979, Soil survey of Jefferson County, Idaho: U. S. Department of Agriculture, Soil Conservation Service, 219 p.

Kjelstrom, L., 1984, Flow characteristics of the Snake River and water-budget analyses for the Snake River Plain, Idaho and eastern Oregon: U. S. Geological Survey Open File Report 84-052, 2 sheets.

Luttrell, S. P., 1982, Ground-water flow characteristics in the Mud Lake area, southeastern Idaho: Unpublished thesis on file with University of Idaho Library, 69 p.

Lindgren, John, 1982, Application of a ground water model to the Boise valley aquifer in Idaho: Unpublished thesis on file with University of Idaho Library, 94 p.

Mundorff, M. J., Crosthwaite, E. G., and Kilburn, Chabot, 1964, Ground water for irrigation in the Snake River Basin in Idaho: U. S. Geological Survey Water Supply Paper 1654, 224 p.

Netz, K. E., 1980, Evaluation of canal seepage in the Snake River Fan Bonneville and Bingham Counties, Idaho: Unpublished thesis on file with University of Idaho Library, 79 p.

Newton, G. D., 1978, Application of a simulation model to the Snake Plain aquifer: Unpublished thesis on file with University of Idaho Library, 82 p.

Norvitch, R. F., Thomas, C. A., and Madison, R. J., 1969, Artificial recharge to the Snake Plain aquifer, an evaluation of potential and effect: Idaho Department of Reclamation Water Information Bulletin No. 12, Boise, Idaho, 59 p.

Ralston, D. R. and Chapman, S. L., 1969, Water level changes in the Mud Lake area, Idaho, 1958-68: Idaho Dept. of Reclamation, Water Information Bulletin No. 7, 29 p.

Robertson, J. B., Schoen, Robert, and Barraclough, J. T., 1974, The influence of liquid waste disposal on the geochemistry of water at at the National Reactor Testing Station, Idaho, 1952-1970: U. S. Geological Survey Open-File Report IDO-22053, 231 p.

Skibitzke, H. E. and da Costa, J. A., 1962, The ground water flow system in the Snake River Plain, Idaho - an idealized analysis: U. S. Geological Survey Water-Supply Paper 1536-D, 67 p.

Stearns, H. T., Bryan, L. L., and Crandall, Lynn, 1939, Geology and water resources of the Mud Lake region, Idaho: U. S. Geological Survey-Water Supply Paper 818, 122 p.

Stearns, H. T., Crandall, Lynn and Steward, W. G., 1938, Geology and ground-water resources of the Snake River Plain in southeastern Idaho: U. S. Geological Survey Water Supply Paper 774, 268 p.

Whitehead, R. L., 1978, Water resources of the upper Henry's Fork basin in eastern Idaho: Idaho Department of Water Resources Water Information Bulletin No. 46, 91 p.

Worstell, R. V., 1976, Estimating seepage losses from canal systems: Journal of the Irrigation and Drainage Division, Proceedings of Am. Soc. Civil Engr., Vol. 102, No. IR1, Mar., 1976, pp. 137-147.

Wright, J. L., 1981, Crop coefficients for estimates of daily crop evapotranspiration: Proceedings of the American Society of Agricultural Engineers Irrigation Scheduling Conference, pp. 18-26.

Wytzes, Jetze, 1980, Development of a groundwater model for the Henry's Fork and Rigby fan areas, upper Snake River basin, Idaho: Unpublished dissertation on file with University of Idaho Library, 196 p.