GROUND-WATER HYDROLOGY AND SIMULATED EFFECTS OF DEVELOPMENT IN THE MILFORD AREA, AN ARID BASIN IN SOUTHWESTERN UTAH

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. Department of the Interior
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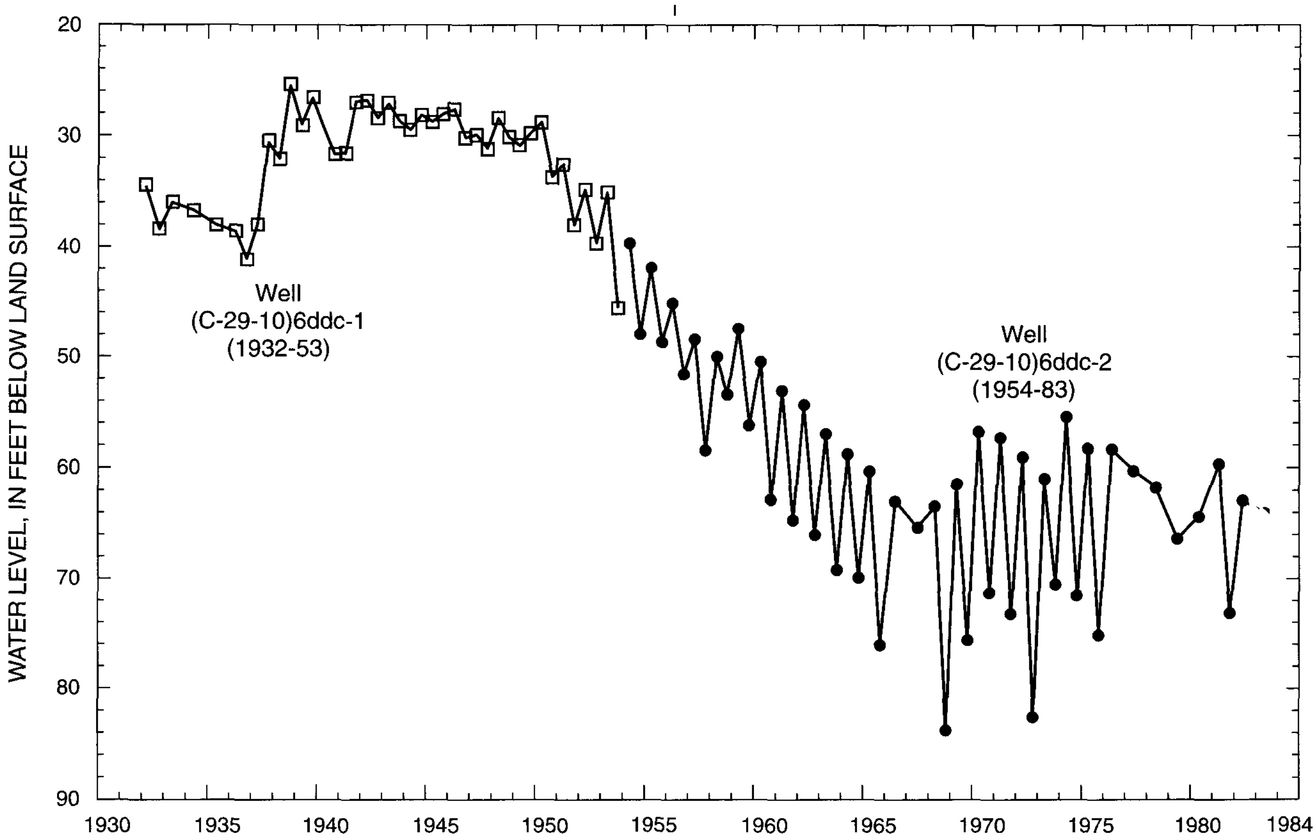

FIGURE 7.—Water levels in two adjacent observation wells in the Milford area, 1932-83.

cific yield, assuming that the water levels will have been lowered so that the confined aquifers have been dewatered; thus, a specific yield representative of water-table conditions will determine the amount of water released from storage. The ground-water system in the Milford area covers approximately 550 mi^2. Assuming an average saturated thickness of 400 ft for the basin fill and a specific yield of 0.15, the amount of recoverable water in storage was estimated to be 21 million acre-ft.

SIMULATION OF GROUND-WATER FLOW

A three-dimensional, finite-difference computer program developed by McDonald and Harbaugh (1988) was used to describe ground-water flow in the basin fill in the Milford area. The finite-difference algorithm used in their computer program can generate only an approximate solution to the partial-differential equation that describes ground-water flow; therefore, the model needs to be considered a tool to help describe the ground-water system. The ground-water model was used to (1) verify or improve estimates of recharge and discharge and the hydraulic properties that describe the basin-fill aquifer; and (2) simulate past and future stresses on the basin-fill aquifer. Both short-term stresses using the present pattern of ground-water withdrawal, and long-term stresses using hypothetical distributions of ground-water withdrawal were simulated. The short-term simulations projected effects of present withdrawals and potential increases in withdrawals. The long-term simulations tested the effects of three kinds of ground-water development: "sustained" yield, ground-water mining, and the capture of natural discharge.

MODEL DESIGN

MODEL GRID AND LAYERS

The three-dimensional, ground-water flow model uses a block-centered or cell-oriented grid system as described by McDonald and Harbaugh (1988, p. 5-1). The grid used for the Milford model consists of 55 rows and 29 columns. Cell spacing ranged from 0.5 to 1.5 mi. The smallest cells, located in the area with numerous wells, cover 0.25 mi^2; largest cells cover 1.5 mi^2. The area of active cells (those actually included in the model calibrations) used in the final model design are shown on plate 1.

The original model design included two layers; an unconfined upper layer and a confined lower layer, except unconfined at the margins of the basin where water levels were below the base of the upper layer. Thickness for the upper layer was held constant at 200 ft, conforming to surface topography.

In some cells along the active edge of the upper layer, the saturated thickness of the porous medium was small. During initial transient simulations, many of these active cells went dry within a short period of time, probably causing adjacent cells to go dry prematurely. In the actual system, water levels do not decline this abruptly, so the ground-water model was redesigned to prevent cells from going dry prematurely by including three layers. The upper layer (layer 1), which represents the unconfined basin fill, was deepened to 250 ft below land surface along the axis of the basin. The bottom of layer 1 was inclined in a north direction to parallel the natural inclination of the land surface. Also, the bottom of layer 1 was made uniform in an east and west direction from the axis, thus increasing its thickness toward the mountains. This design eliminated the problem of cells going dry because almost all cells in layer 1 have large saturated intervals, except in the extreme northwest corner. Due to the steep hydraulic gradient in that area, a few cells have small saturated intervals, and some cells are inactive because the water level is below the bottom of layer 1.

The middle layer (layer 2) generally represents a confined aquifer. Only those cells in the extreme northwest corner, which lie under the inactive (dry) cells of layer 1, simulate unconfined conditions. The confining bed is not simulated; therefore, the top of layer 2 coincides with the bottom of layer 1. Because of the lack of data defining the base of the ground-water system, the thickness of layer 2 was not specified; therefore, constant transmissivity had to be used for this layer during all simulations, rather than computing transmissivity from hydraulic conductivity and saturated thickness.

After initial transient simulations, the bottom layer (layer 3) was added in order to provide a source of water for upward leakage into layer 2. Without this layer, computed water-level declines in layer 1 and layer 2 were almost twice the historical water-level declines. Layer 3 represents a confined aquifer with a constant transmissivity as in layer 2; thus, no top or bottom surfaces had to be specified.

BOUNDARY CONDITIONS

Boundary conditions during model simulations are of three types: constant head, constant flux, and mixed. Constant-head cells maintain the specified head for the entire simulation. Fluxes entering or leaving the ground-water system through the constant-head cells are calculated based on the head gradient and transmissivity between the boundary and interior cells. Constant-flux cells maintain the specified flux for the entire simulation; the heads are calculated. An impermeable or no-flow boundary can be simulated by constant-flux cells with a specified flux of zero. Mixed boundary conditions are handled by the general-head boundary module in the ground-water flow model of McDonald and Harbaugh (1988, p. 11-1), wherein the head and flux are calculated at the model boundary using a specified conductance and head at some distance outside the boundary. The conductance can be determined by multiplying the hydraulic conductivity along the flow path from the model boundary to the specified head by the cross-sectional area of the cell at the model boundary and dividing by the length of the flow path. If transmissivity is used instead of hydraulic conductivity, then transmissivity is multiplied by the length of the cell rather than cross-sectional area.

During steady-state calibration, constant-head, constant-flux, and mixed (general-head) cells were used at the boundaries. Constant-head cells were used in layer 1 along recharge boundaries where the estimated head and hydraulic conductivity were considered to be more accurate than any estimate of recharge. As shown in plate 1, constant-head cells were placed along the entire length of the eastern boundary from the north end of the simulated area near Black Rock to south of the Beaver River near Minersville.

Constant-flux cells in layer 1 were placed along boundaries where the potentiometric surface indicated that no appreciable subsurface inflow enters the ground-water system, and were assigned a flux value of zero. This type of boundary condition generally exists along the Black Mountains/basin-fill interface in the south and most of the San Francisco Mountains/basin-fill interface on the west.

The general-head (mixed) boundary was used along two model boundaries where subsurface flow into or out of the basin occurs. This type of boundary was chosen in order to quantify any changes in basin inflow and outflow due to declining water levels during transient simulations. All flow was assumed to enter or leave through layer 2; therefore, cells in layers 1 and 3 along this type of boundary were specified as no-flow. One general-head boundary was placed along the southwest edge where basin inflow enters from the Beryl-Enterprise area. The other general-head boundary was placed along the northwest boundary at the San Francisco Mountains/basin-fill interface and along the north edge of the simulated area near Black Rock. The general-head boundary was not used where interbasin flow enters the basin along Cove Creek and the Beaver River. Because of the small number of cells and the small area involved, any change in head would not make a substantial difference in computed inflow. All boundary cells in layer 2 not specified as general head, and all boundary cells in layer 3 were simulated as no-flow.

INITIAL CONDITIONS

WATER LEVELS

Within the area of ground-water development, initial water levels used for steady-state calibration were measured in 1927 and reported by White (1932, p. 58). Additional water levels measured in wells through 1983 were used outside the developed area, where data were few, and at margins of the basin where steady-state conditions were assumed at the time of measurement.

RECHARGE

Simulated recharge includes subsurface inflow from consolidated rocks, seepage losses from canals and unconsumed irrigation water, infiltration from perennial and ephemeral streams, subsurface inflow from adjoining basins through basin-fill deposits, and precipitation on basin-fill deposits at margins of the basin. On the basis of estimated-head values along the southern and western margins of the basin, no appreciable subsurface inflow enters from the consolidated rocks of the Black Mountains and San Francisco Mountains. Apparently, all subsurface inflow from consolidated rocks is from the Mineral Mountains and from the basalt east of Black Rock. Although this inflow was estimated to be 15,000 acre-ft/yr, the model computed this inflow using a constant-head boundary during steady-state calibration.

Seepage losses from canals and unconsumed irrigation water is another major component of recharge to the ground-water system. Seepage losses from canals, reported by Mower and Cordova (1974, p. 18), vary with the quantity of flow in the canals. On the basis of annual diversions, they reported a weighted average loss of 14 percent of the water diverted in a test reach, or 1.5 percent per mile. This average rate of loss was assumed to apply for the total 23 mi of canals in the area; thus 34 percent of all diversions from the Beaver River is lost. An estimated 4 percent of the loss is assumed to be transpired by vegetation, leaving 30 percent to recharge the ground-water system. The rate of recharge to the ground-water system is thus estimated to be 1.3 percent per mile of canal. This rate, which is used for all major canals, is assumed to be constant for both steady-state and transient conditions despite yearly changes in flow. Losses from major canals were calculated by multiplying 0.013 times the continually decreasing flow for each mile of canal. These losses were then distributed to the appropriate cells based on the length of the canal in a cell. Losses from small canals were assumed to be part of unconsumed water applied to irrigated lands.

The amount of discharge in the Beaver River, which is regulated upstream at the Rocky Ford Dam, determines the distribution for irrigation. By prior rights, the area near Minersville is allocated 13.8 ft^3/s (10,000 acre-ft/yr). The remainder of the total discharge is available for use in the area near Milford, up to 21.4 ft^3/s (15,500 acre-ft/yr), the maximum quantity that can be transported in the Low Line Canal. Discharge exceeding the total 35.2 ft^3/s (25,500 acre-ft/yr) is allowed to flow down the Beaver River channel.

For steady-state calibration, the total 1927 discharge of 31.8 ft^3/s (23,000 acre-ft/yr) (Mower and Cordova, 1974, fig. 3) was used. As mentioned above, 13.8 ft^3/s (10,000 acre-ft/yr) is allotted to the area near Minersville. Along the 4-mile reach of the Minersville Canal, 0.7 ft^3/s (510 acre-ft/yr) was lost to the ground-water system, thus leaving 13.1 ft^3/s (9,480 acre-ft/yr) for irrigation. Assuming 30 percent infiltration of applied irrigation water, 3.9 ft^3/s (2,820 acre-ft/yr) infiltrated to the ground-water system and was distributed to cells that cover the currently irrigated area.

After subtracting canal losses of 2.2 ft^3/s (1,600 acre-ft/yr) from the 18.0 ft^3/s (13,000 acre-ft/yr) diverted into the Low Line Canal, the amount of Beaver River water available for irrigation in the area near Milford was 15.8 ft^3/s (11,400 acre-ft/yr), of which 4.7 ft^3/s (3,400 acre-ft/yr) infiltrated to the ground-water system. This water was divided among cells that coincide with lands irrigated with surface water as shown by Nelson (1950, fig. 11).

Seepage from irrigated lands using ground water was assumed to be 30 percent as suggested by Mower and Cordova (1974, p. 21). White (1932, p. 88) reported 6.9 ft^3/s (5,000 acre-ft/yr) of ground water was used for irrigation in 1927, of which 30 percent or 2.1 ft^3/s (1,500 acre-ft/yr) was assumed to be recharge. This seepage was applied by using the distribution shown by White (1932, fig. 2).

Stream infiltration from the Beaver River is a minor source of recharge except in wet years, when the flow downstream from Rocky Ford Reservoir is substantially greater than the 35.2 ft^3/s (25,500 acre-ft/yr) diverted for irrigation. Winter minimum flows are 5 ft^3/s or less as reported by Nelson (1950, p. 185). For steady-state simulation purposes, an average of 2.1 ft^3/s (1,500 acre-ft/yr) is assumed to infiltrate to the ground-water system and is distributed along 5 mi of river channel.

All recharge from stream and canal losses and seepage from irrigated lands was simulated as recharging wells with a fixed flux. Stream and canal losses were not simulated as head dependent because water levels are below the streambeds or canals.

Subsurface inflow from the Beryl-Enterprise area was computed during model simulations using the general-head boundary module of McDonald and Harbaugh (1988, p. 11-1). This inflow increased during transient simulations when water-level declines extended to the boundary.

Recharge from precipitation was simulated using the recharge module (McDonald and Harbaugh, 1988, p. 7-1) in the eastern part of the basin where precipitation is greater than 10 inches/yr. Only 5 percent of the precipitation is assumed to infiltrate to the ground-water system because of high evaporation rates and consumption by vegetation. Recharge from precipitation is estimated to be 5.1 ft^3/s (3,700 acre-ft/yr), which is greater than the 2.8 ft^3/s (2,000 acre-ft/yr) reported by Mower and Cordova (1974, p. 21). Recharge for each cell was entered as ft^3/s per ft^2 of area, or ft/s for each cell.

DISCHARGE

Initial discharge from the ground-water system was simulated as evapotranspiration, withdrawal from wells, and basin outflow through the northwest general-head boundary. Evapotranspiration was simulated in cells where the water table is within 30 ft of land surface. Using this extinction depth, the area simulating evapotranspiration (pl. 2) corresponds to the phreatophyte area mapped by Mower and Cordova (1974, pl. 3). A maximum evapotranspiration rate is assigned to each cell within the phreatophyte area. The computed evapotranspiration is based on a linear proportion of the maximum rate and the depth of water below land surface at each cell (McDonald and Harbaugh, 1984, p. 317). Evapotranspiration rates used during model simulations ranged from 8.2×10^{-9} ft/s (3 inches/yr) to 9.8×10^{-8} ft/s (37 inches/yr). These rates are similar to the rates determined by White (1932, p. 86).

Ground-water withdrawal from wells was estimated by White (1932, p. 88) to be 5,000 acre-ft for 1927. The areal distribution for withdrawal from wells determined by White (1932, fig. 2) was used for steady-state calibration.

Basin outflow to the northwest was calculated during steady-state simulations using a general-head boundary. By calibrating computed heads to estimated heads and assuming the estimated distribution of transmissivity approximates reality, a reasonable estimate for basin outflow can be determined.

HYDRAULIC CONDUCTIVITY

Horizontal hydraulic-conductivity values for the upper water-table layer (layer 1) were estimated from drillers' logs using a weighted average for all lithologies, and from specific capacities determined from tests conducted at time of drilling. These hydraulic-conductivity values are multiplied by the saturated thickness in each cell during model computations. Hydraulic-conductivity values range from 2.3×10^{-5} ft/s (2 ft/d) to 6.9×10^{-4} ft/s (60 ft/d) and were changed by trial-and-error during steady-state calibration. The middle layer (layer 2) was simulated as a confined-unconfined aquifer using a constant transmissivity rather than calculating transmissivity from the saturated thickness. Transmissivity values in layer 2 range from 9.3×10^{-3} ft^2/s (800 ft^2/d) to 5.4×10^{-1} ft^2/s (47,000 ft^2/d) and were similar to the distribution reported by Mower and Cordova (1974, fig. 4). The bottom layer was simulated as a confined system. Hydraulic properties of this layer are unknown because this part of the aquifer lies below the level of present development. Transmissivity values were arbitrarily assumed to be one-third of those used in the middle layer.

Most irrigation wells in the Milford area are completed in multiple permeable zones, thus maximizing production. The lack of wells with completion in a specific zone limits the ability to determine vertical hydraulic conductivity and vertical head gradient. For this reason, no estimates of vertical hydraulic conductivity and vertical head gradient were reported by Mower and Cordova (1974) and no aquifer tests were designed to determine vertical hydraulic conductivity during this study. From limited water-level data representative of specific zones, the difference in head between the water-table or semi-confined aquifer (layer 1) and the underlying confined aquifer (layer 2) was estimated to range from 1 to 10 ft in the center of the basin.

The model calculates vertical flow between layers from data incorporating vertical hydraulic conductivity and aquifer thickness. The resulting term, known as vertical leakance (McDonald and Harbaugh, 1988, p. 5-12), is calculated by dividing the estimated vertical hydraulic conductivity by the distance between the centers of adjoining model layers. Vertical hydraulic conductivity can be assumed to be one to two orders of magnitude smaller than horizontal hydraulic conductivity. Although the model was designed with unspecified thicknesses for layers 2 and 3, the distance between the centers of adjoining model layers can be assumed to be a few hundred feet. Initial vertical leakance between layers, therefore, was estimated to range between 1.0 to 5.0×10^{-9}/s (8.6×10^{-5} to 4.3×10^{-4}/d). The larger values were distributed at the basin margin where vertical leakage is larger than in the center of the basin. These values were adjusted during calibration to maintain the estimated head differences between layers 1 and 2.

STORAGE COEFFICIENTS

An average value of 0.20 for specific yield as reported by Mower and Cordova (1974, p. 15) was used for the entire upper layer (layer 1). In the middle layer (layer 2), two storage-coefficient arrays are necessary because it is simulated as a confined-unconfined system. The primary array contains the storage-coefficient values for the confined aquifer that range from 5.0×10^{-4} to 1.5×10^{-3}. The secondary storage-coefficient array is necessary when cells in layer 1 become dry and the underlying cells in layer 2 simulate an unconfined aquifer. Because of greater compaction at depth, the secondary storage-coefficient array was assigned an average value of 0.10. The bottom layer (layer 3) is confined throughout the simulated area and was assigned values an order of magnitude less than the primary storage-coefficient array for layer 2.

STEADY-STATE CALIBRATION

The model was first calibrated to steady-state heads known to exist prior to large-scale ground-water development. Hydraulic conductivity and transmissivity in all layers were varied along with vertical leakance during steady-state calibration to obtain a best fit to initial heads. Evapotranspiration rates also were adjusted to get a better match to initial heads; however, all adjusted values remained within the initial ranges.

In the calibration process, the vertical leakance values between layers were adjusted so that the estimated head differences between layers 1 and 2 were maintained and the necessary quantity of water from layer 3 moved upward. Calibrated vertical leakance values between layers 1 and 2 range from 4.8×10^{-10} to 6.1×10^{-8}/s (4.1×10^{-5} to 5.3×10^{-3}/d). Calibrated vertical leakance values between layers 2 and 3 range from 1.2×10^{-9} to 7.2×10^{-8}/s (1.0×10^{-4} to 6.2×10^{-3}/d).

Initial conductance values were varied at both general-head boundaries. At the southwestern boundary, the conductances were varied to match computed heads and fluxes to known heads and to the total estimated subsurface inflow of 2.9 ft^3/s (2,100 acre-ft/yr) (Mower, 1982, p. 47). At this boundary, the final conductances range from 4×10^{-3} to 2.5×10^{-2} ft^2/s (345 to 2,160 ft^2/d). Through this process, the computed steady-state heads were within 20 ft of the initial heads estimated from field data and the total computed subsurface inflow was only 0.2 ft^3/s (145 acre-ft/yr) below the initial estimate.

Because no estimate had ever been made for subsurface outflow along the northwestern general-head boundary, the conductances were varied in order to match estimated heads in the boundary cells. Most of the computed heads were within 30 ft of the estimated heads in this area; a 65-ft difference was the largest deviation. These computed heads are considered to be within calibration limits given that the estimated head gradient is greater than 100 ft/mi, which is based on very limited data and large cell size in this area. The model-calculated, steady-state subsurface outflow for this boundary is 15.8 ft^3/s (11,400 acre-ft/yr). In most cases, the conductance for each cell was increased by three times the original estimate. The final conductances at the northwest boundary range from 1×10^{-3} to 4×10^{-2} ft^2/s (86 to 3,460 ft^2/d).

Steady-state calibration criteria included a match to within 5 ft of initial heads in the area of ground-water development and a reasonable match in other areas depending on hydraulic gradient, topography, and quantity of data. Within a 309-cell area, where sufficient data were available to make a reasonable estimate of steady-state water levels or where the estimated hydraulic gradient was small, the average difference between initial head and computed head was less than 2 ft for all three layers. Along most of the margins of the basins, a match to within 20 ft was considered to be within calibration limits. In a few areas, such as the southeast recharge area near Minersville and in the northwestern outflow area, the difference between initial and computed heads was greater than 20 ft. This can be attributed to the steep hydraulic gradient, large cell size, and uncertainty in estimated initial heads.

Recharge from consolidated rocks was simulated by using constant-head cells along the eastern boundary of the area. The steady-state calibrated model is just one of many possible solutions because recharge is unbounded. The calibrated model, however, is a reasonable approximation of the ground-water system under steady-state conditions because hydraulic conductivity was varied within reasonable limits along this boundary in order to match initial heads. The original estimate for recharge from consolidated rocks was calculated to be 20.7 ft^3/s (15,000 acre-ft/yr). By not assigning this recharge as constant flux along this boundary, the model calculated the recharge to be 32.9 ft^3/s (23,800 acre-ft/yr).

Computer-generated steady-state contours for initial and computed heads in layers 1 and 2 are shown in figures 8 and 9. The similarity in the figures is due to the relatively small head differences between the two layers, except in the center of the basin, where estimated head differences are assumed to be 1 to 10 ft.

Prior to transient calibration, all constant-head cells along the eastern boundary of the simulated area were

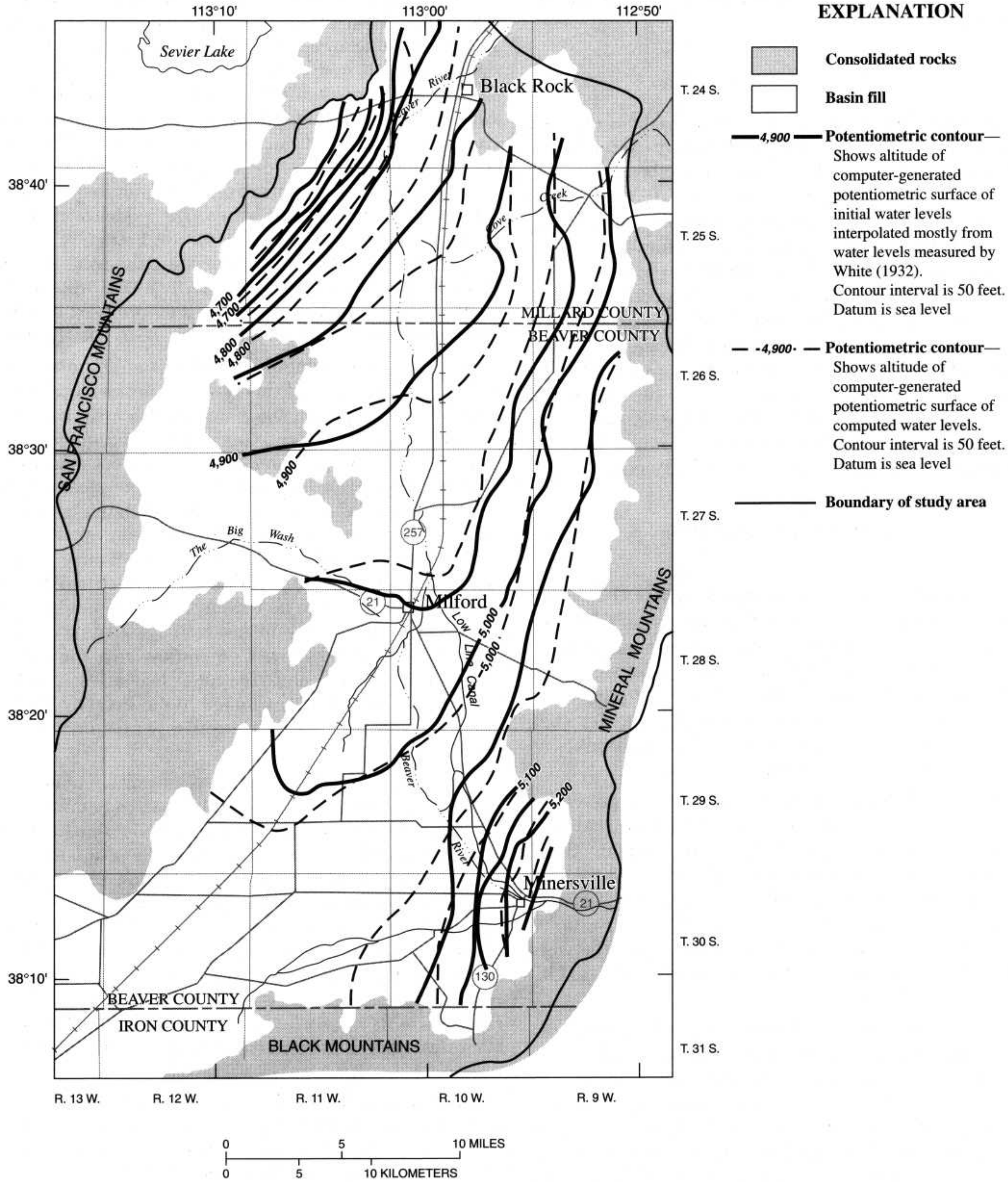

FIGURE 8.—Steady-state contours for initial and computed water levels in layer 1.

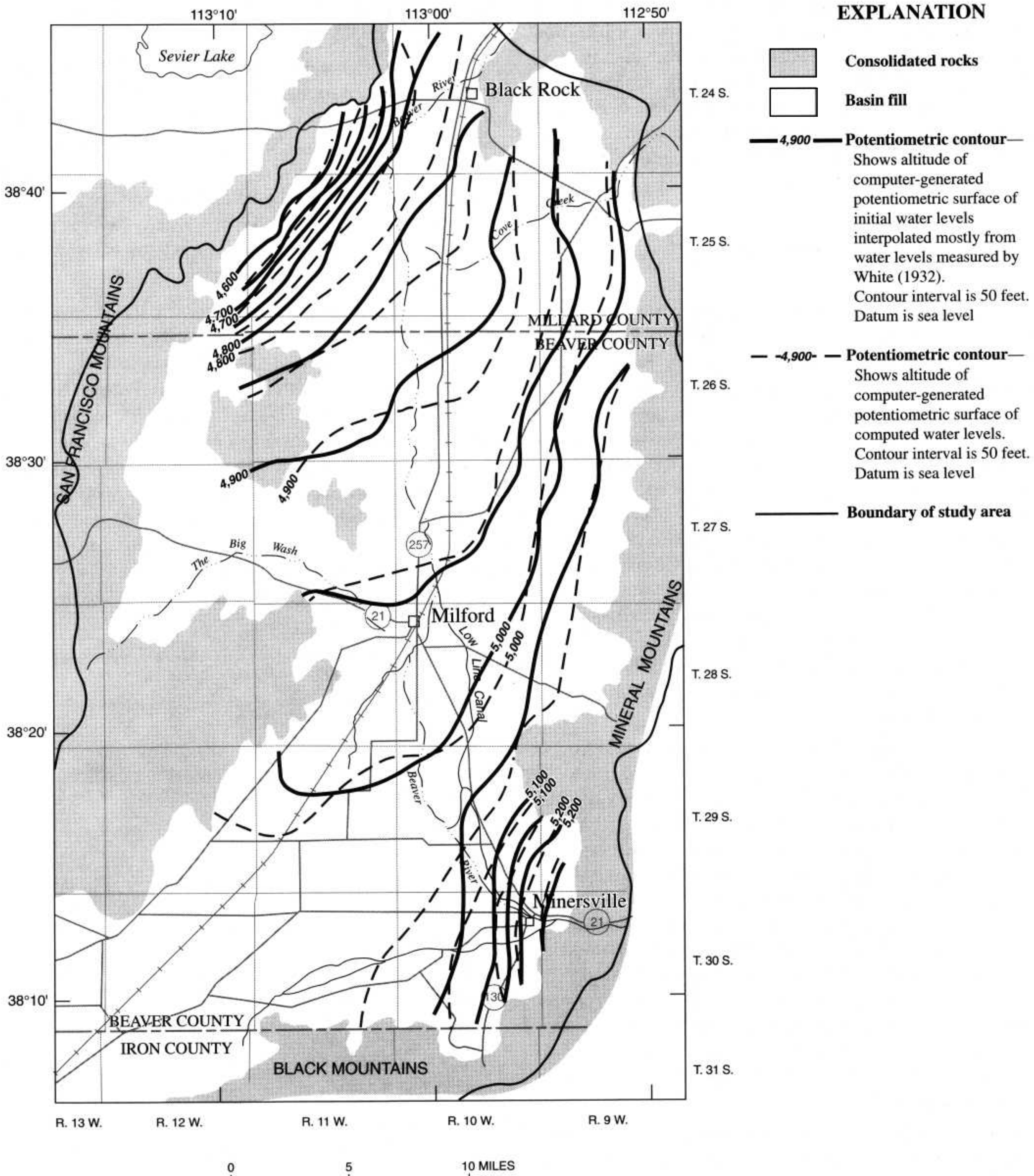

FIGURE 9.—Steady-state contours for initial and computed water levels in layer 2.

converted to constant-flux cells with all model-calculated fluxes entered into the same recharge array as infiltration from precipitation. As a transition step from steady-state to transient calibration, arrays containing the necessary storage-coefficient values were entered and a 100-year simulation was made using steady-state conditions. This simulation verified that the model-calculated boundary fluxes did not vary with time.

TRANSIENT CALIBRATION

Transient calibration involved the use of time-dependent data such as storage, known ground-water withdrawals, and varying seepage from the Beaver River channel and irrigated lands. Storage was adjusted in order to match known water-level fluctuations. Steady-state conditions were assumed to have prevailed in the Milford area until 1950 when ground-water withdrawals began to increase rapidly. Withdrawals varied from 15.0 ft^3/s (10,900 acre-ft/yr) to 31.5 ft^3/s (22,800 acre-ft/yr) during 1931-49. After 1950, withdrawals did not drop below 44.2 ft^3/s (32,000 acre-ft/yr) (fig. 6). Coincident with this increase, water levels in observation wells (C-29-10)6ddc-1 and (C-29-10)6ddc-2 began to decline after 1950 (fig. 7). The model-calculated, steady-state heads were assumed to be representative of the ground-water system prior to 1950 and were used as initial heads for transient calibration.

Seven pumping or stress periods were selected for the transient calibration: 1950–52, 1953–60, 1961–67, 1968–72, 1973, 1974–78, and 1979–82. During these intervals, discharge from wells was relatively constant. The fifth stress period, 1973, was a year of high flows in the Beaver River and decreased pumping. By defining a single-year stress period, the model could be tested for its response to increased recharge and decreased ground-water withdrawals. Water-level changes were computed for the end of each stress period within the 33-year transient-simulation period, starting from the calibrated, steady-state water levels. The computed changes in water levels for each stress period were compared to water-level changes calculated from measurements made in March of the year after each stress period.

Differences between measured and computed water-level changes may reflect the response of the aquifer to a large change in the last year of a stress period rather than the overall trend for the entire stress period. Also, in some observation wells, the measured water levels are representative of both the unconfined and confined aquifers, depending on the location of the perforated intervals; whereas computed water levels are representative of either the unconfined or confined aquifers, depending on location and model layer.

The average ground-water withdrawals applied during each stress period are shown in table 1. Annual ground-water withdrawals for each well were averaged over each stress period. If a well penetrated more than one model layer, the average withdrawal was divided proportionally between the layers based on the percent of perforated interval in each layer. Finally, average withdrawals for all wells were combined for each cell in a model layer.

Two simulations were made for the transient calibration, each treating recharge from seepage to the ground-water system differently. In both cases, seepage was based on the mean annual flow in the Beaver River as measured at Rocky Ford Dam. For the first transient simulation, seepage from surface-water irrigation, canal losses, and infiltration from the Beaver River were averaged for each stress period. As mentioned previously, the 13.8 ft^3/s (10,000 acre-ft/yr) of surface water diverted to the area near Minersville remained constant due to prior rights. In 1960, the flow in the Beaver River was slightly below the amount allocated to the Minersville area (fig.

TABLE 1.—*Variation in water-budget components during historical transient simulations in the Milford area, Utah*

[Data are in cubic feet per second; acre-feet per year shown in parentheses]

Stress period (years)	Ground-water withdrawals	Recharge from seepage to ground-water system[1] Variable[2]	Constant[3]
1950–52	42.8 (31,000)	30.4 (22,000)	26.4 (19,100)
1953–60	56.5 (40,900)	28.6 (20,700)	30.5 (22,100)
1961–67	61.5 (44,500)	27.4 (19,800)	32.0 (23,200)
1968–72	74.2 (53,700)	39.9 (28,900)	35.8 (26,000)
1973	69.0 (50,000)	60.6 (43,900)	34.2 (24,800)
1974–78	83.0 (60,100)	42.9 (31,100)	38.4 (27,800)
1979–82	63.6 (46,000)	51.8 (37,600)	32.6 (23,600)

[1]Sum of seepage from irrigated lands using surface and ground water, canal losses, and infiltration from the Beaver River.

[2]Based on the mean annual flow in the Beaver River and associated canals during each stress period.

[3]Based on the mean annual flow in the Beaver River and associated canals during the entire simulation period.

3); however, the mean annual flow for any one stress period was never lower than that allocation.

A ratio was calculated by which the seepage in the area near Milford was adjusted for each stress period. The diversion to the area near Minersville was subtracted from the mean annual flow in the Beaver River as measured at Rocky Ford Dam. The remaining flow, up to a maximum of 21.4 ft^3/s (15,500 acre-ft/yr), was divided by the flow that was diverted to the area near Milford in the steady-state calibration. This ratio was then multiplied by the steady-state seepage for each affected cell, thus increasing or decreasing the seepage for each stress period from the steady-state seepage. The same distribution of surface-water irrigated lands was used for each stress period. If the mean annual flow in the Beaver River was in excess of the 35.2 ft^3/s (25,500 acre-ft/yr) diverted for irrigation in any stress period, then the entire amount of excess flow was assumed to recharge the ground-water system and was distributed among the cells along the Beaver River channel.

A second simulation was made in which one set of average values was used for seepage to the ground-water system during the entire simulation period rather than a different set of values for each stress period. In this simulation, seepage to the ground-water system in the area near Minersville remained the same as in the previous simulation.

Seepage from land irrigated with ground water was simulated for each stress period by assuming that 30 percent of the water withdrawn from each cell returned to the ground-water system. Because this type of recharge is independent of the surface-water system, the percentage of seepage remained constant for each transient simulation. Computer-generated contours of computed water levels for 1983, at the end of the transient simulation using constant recharge, are reasonably close to water levels measured in wells for that year (fig. 10).

Differences in simulated water levels for the two transient simulations using constant and varying recharge from seepage for each stress period are considered to be substantial if greater than 2 ft. These differences are found in or near areas of recharge from seepage of surface water as shown in figure 11. Measured and computed water-level changes for 13 observation wells that have data for most or all of the stress periods are shown in figure 12. These observation wells are located in the north and west parts of the developed area where the effects of seepage from surface-water irrigation are negligible; therefore, only computed water-level changes using constant recharge are compared to measured water-level changes. Measured and computed water-level changes for seven observation wells in which there are some, but minimal, differences are shown in figure 13. Simulated water-level changes for three observation wells where there are substantial differences between the two types of computed water-level changes are shown in figure 14. These observation wells are located in the southeastern part of the developed area, closest to the area of surface-water recharge. The computed water levels from the transient simulation using constant seepage from surface-water irrigation show a steady decline. This decline might be the result of withdrawals in the main pumping center to the north. By varying the recharge for each stress period, recharge becomes the dominant influence on the ground-water system in this area, although the long-term decline still occurs. The computed water-level changes follow the same trends as those shown by the measured water-level changes; however, overall water-level declines obtained from constant recharge show a better match to the total measured declines.

When seepage to the ground-water system from surface-water sources was varied, model-calculated water-budget components of evapotranspiration and basin outflow were not substantially different than those calculated when seepage for each stress period was constant (table 2). In the last stress period with varying seepage, however, the large increase in seepage from the Beaver River resulted in a net increase of almost 5 ft^3/s (3,620 acre-ft/yr) of water going into storage. In addition, the three observation wells that show substantial water-level changes are located near the Beaver River. This would indicate that the greatest effect on computed water levels is the result of excess flow in the Beaver River channel recharging the ground-water system.

Historical transient simulations using both varying and constant recharge from seepage of surface-water irrigation show water-level declines of nearly 22 ft along the eastern constant-flux boundary; however, limited water-level data in this area indicate that there have been no actual water-level declines during the 1950–82 simulation period. Adjusting the constant-flux rates based on the variation from average precipitation for each stress period still gave computed water-level declines of nearly 16 ft along the eastern boundary. The simulated boundary effects are therefore probably due to the grid spacing in this narrow basin and large simulated ground-water withdrawals, rather than the flux rates at the boundary.

SENSITIVITY ANALYSIS

Numerous simulations were made to determine the sensitivity of the calibrated, steady-state model to changes in input data. Each parameter was increased and decreased by 20 percent of its final calibrated value for all layers simultaneously and for each layer separately.

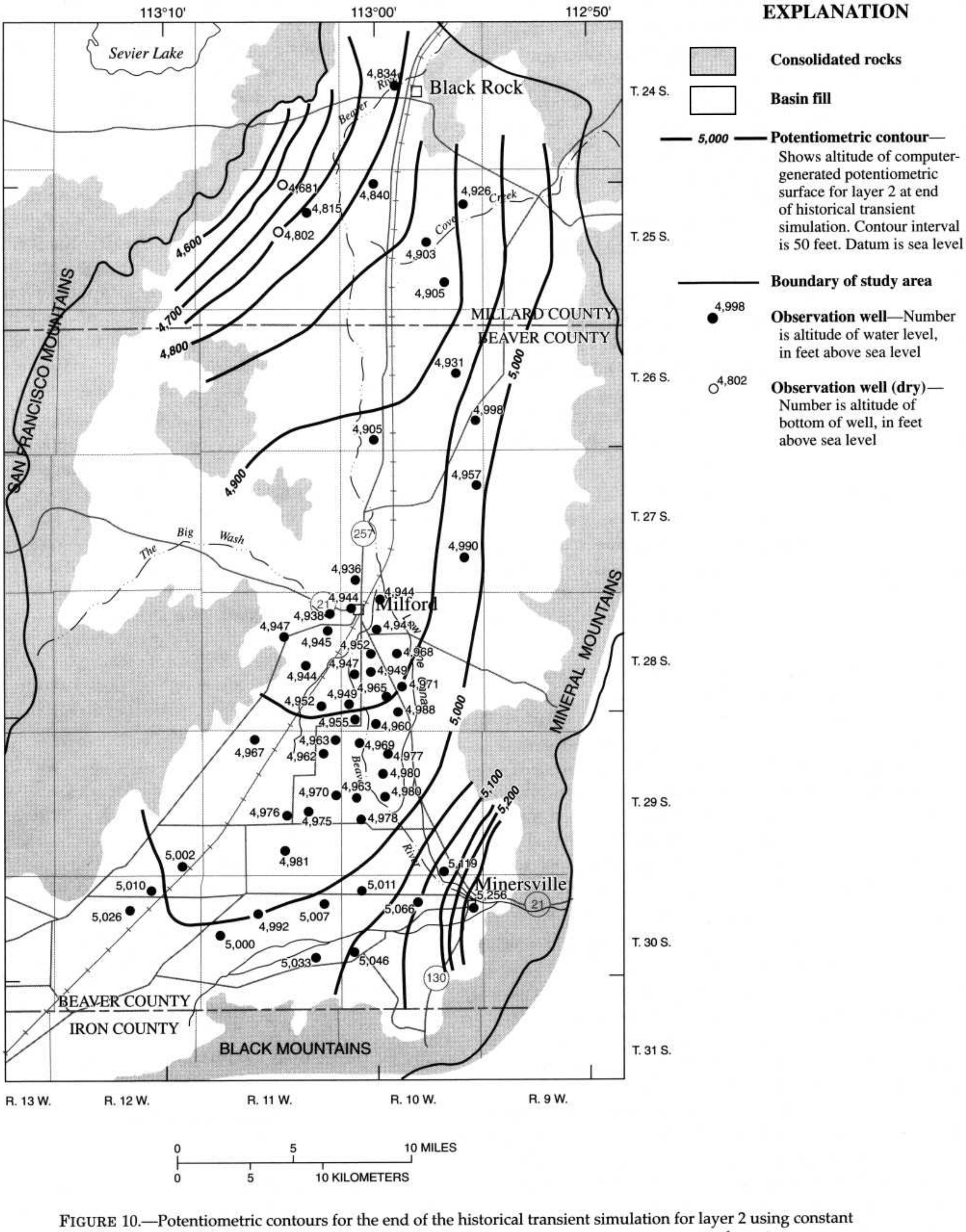

FIGURE 10.—Potentiometric contours for the end of the historical transient simulation for layer 2 using constant recharge from seepage of surface water and measured water levels in wells in March 1983.

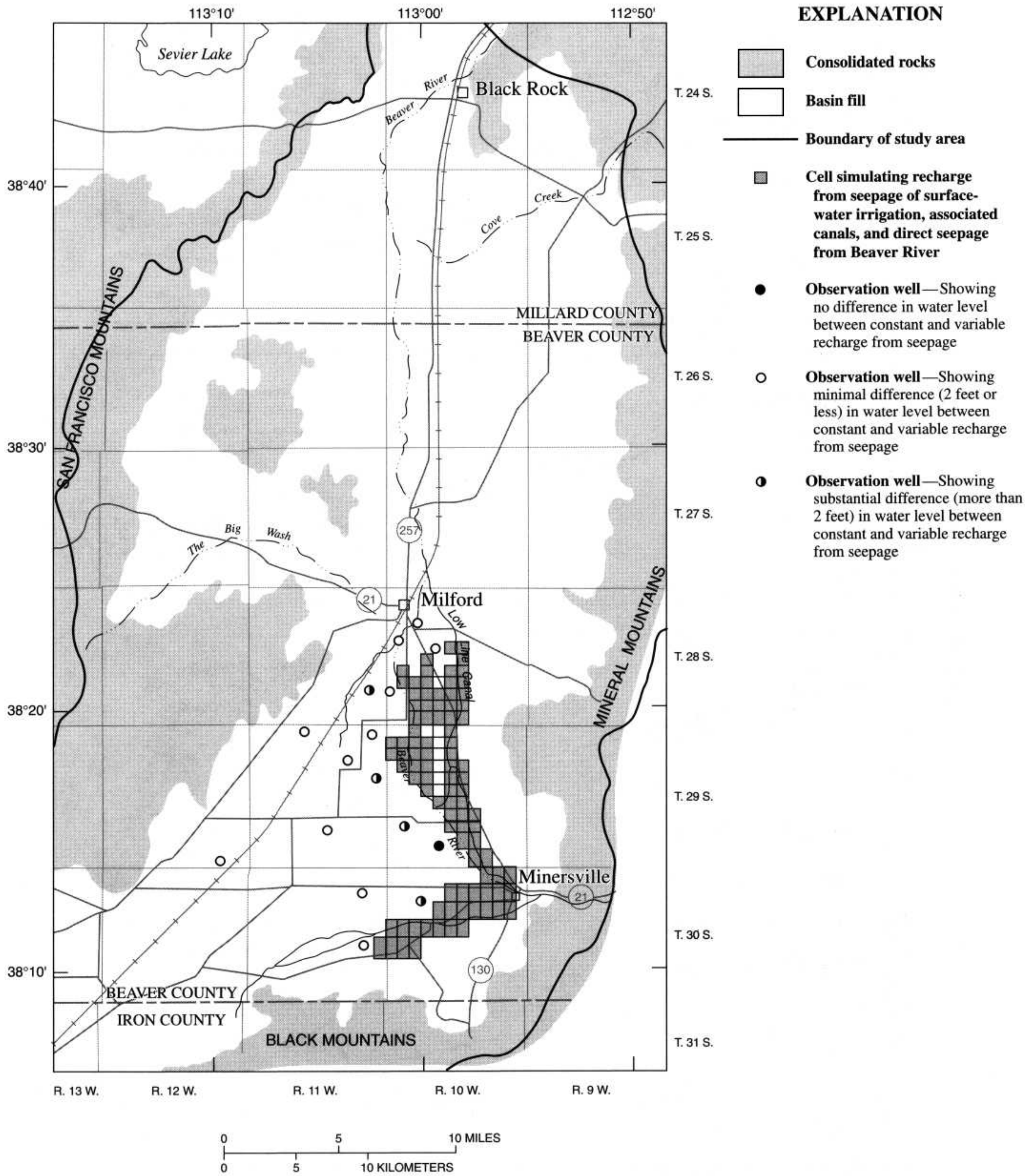

FIGURE 11.—Model cells receiving constant and variable recharge from seepage of surface water during transient simulations and location of observation wells from which measured water levels are compared to computed water levels.

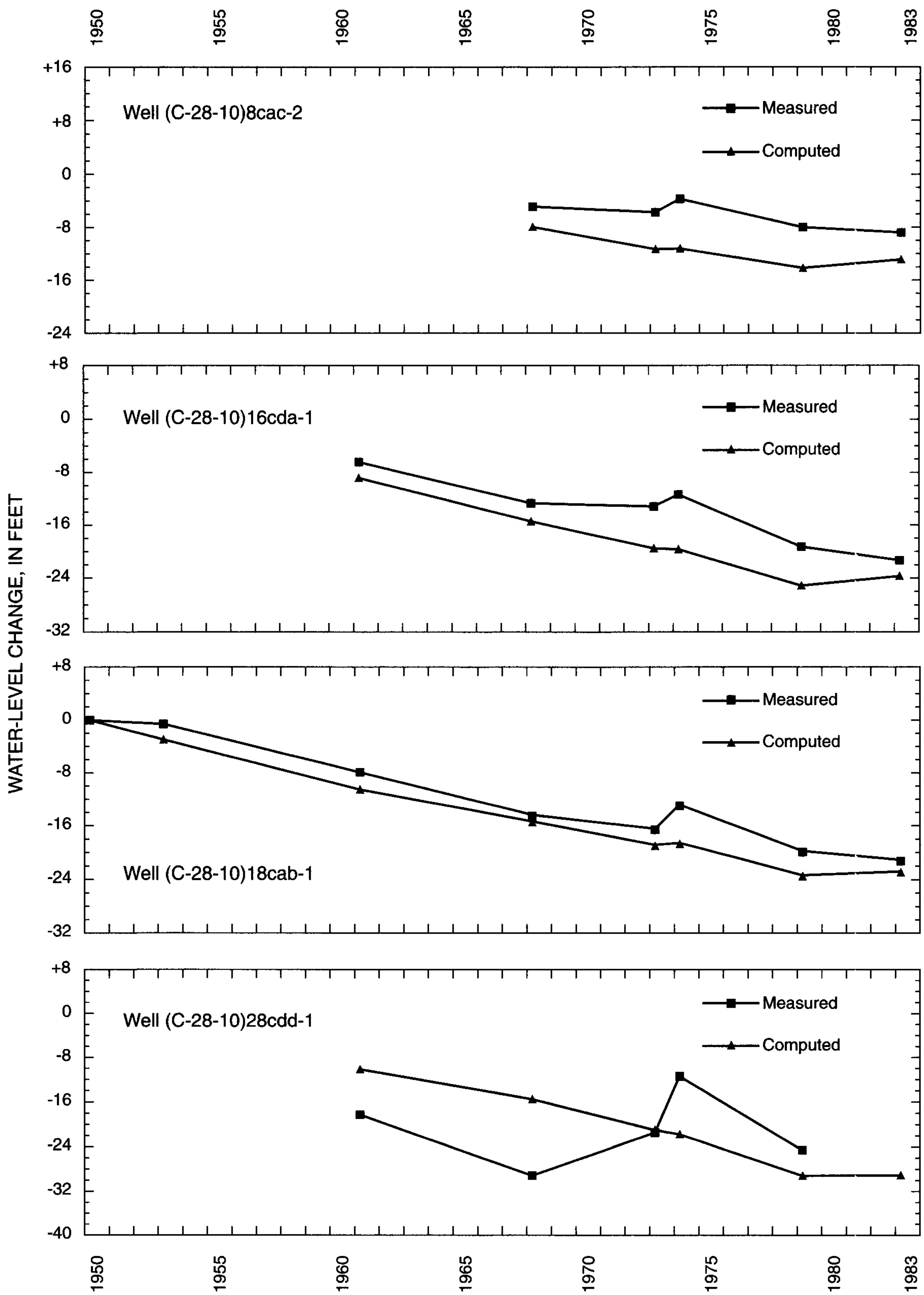

FIGURE 12.—Measured and computed water-level changes during 1950–83 for 13 observation wells in the Milford area.

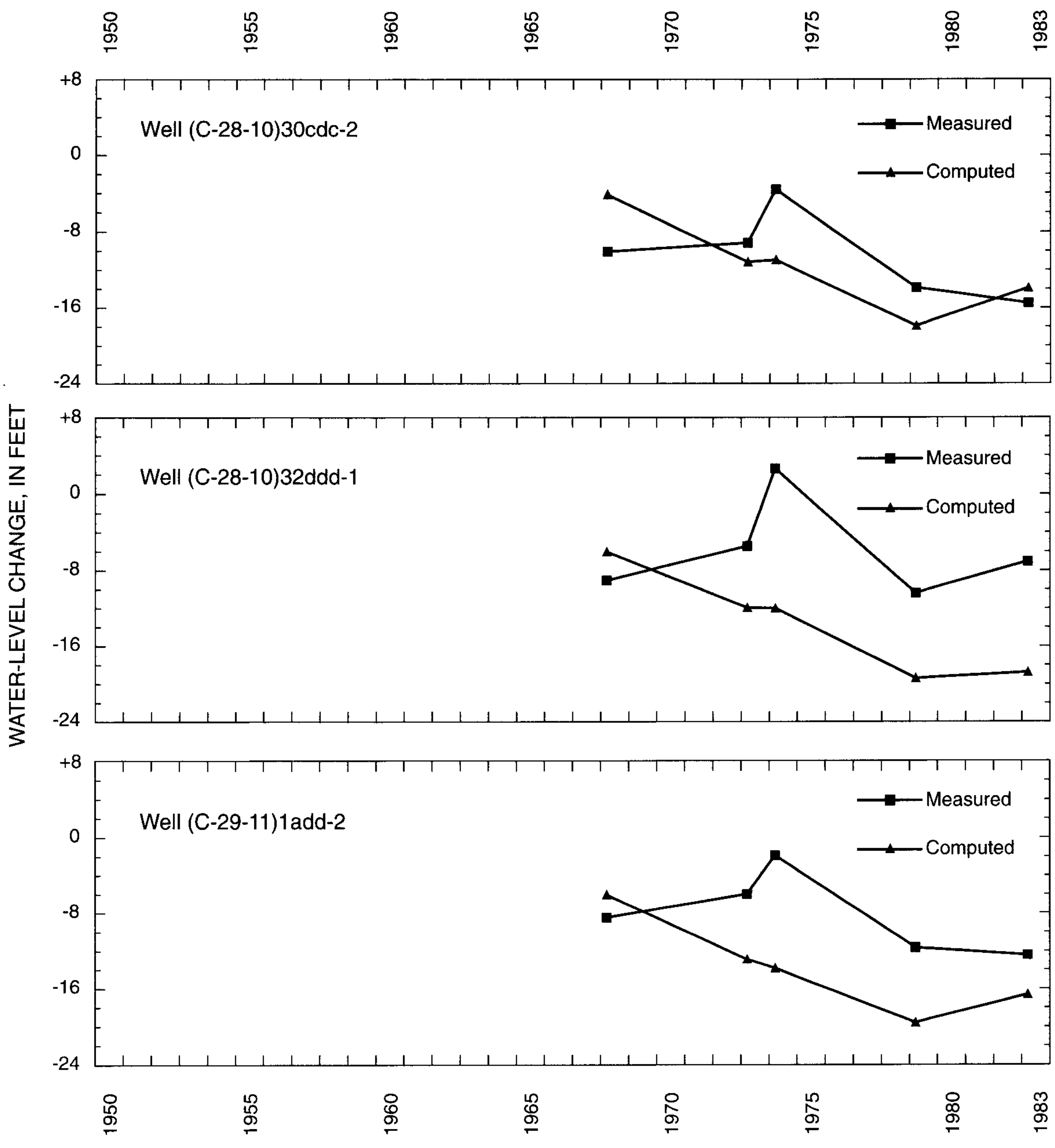

FIGURE 12.—Measured and computed water-level changes during 1950–83 for 13 observation wells in the Milford area—Continued.

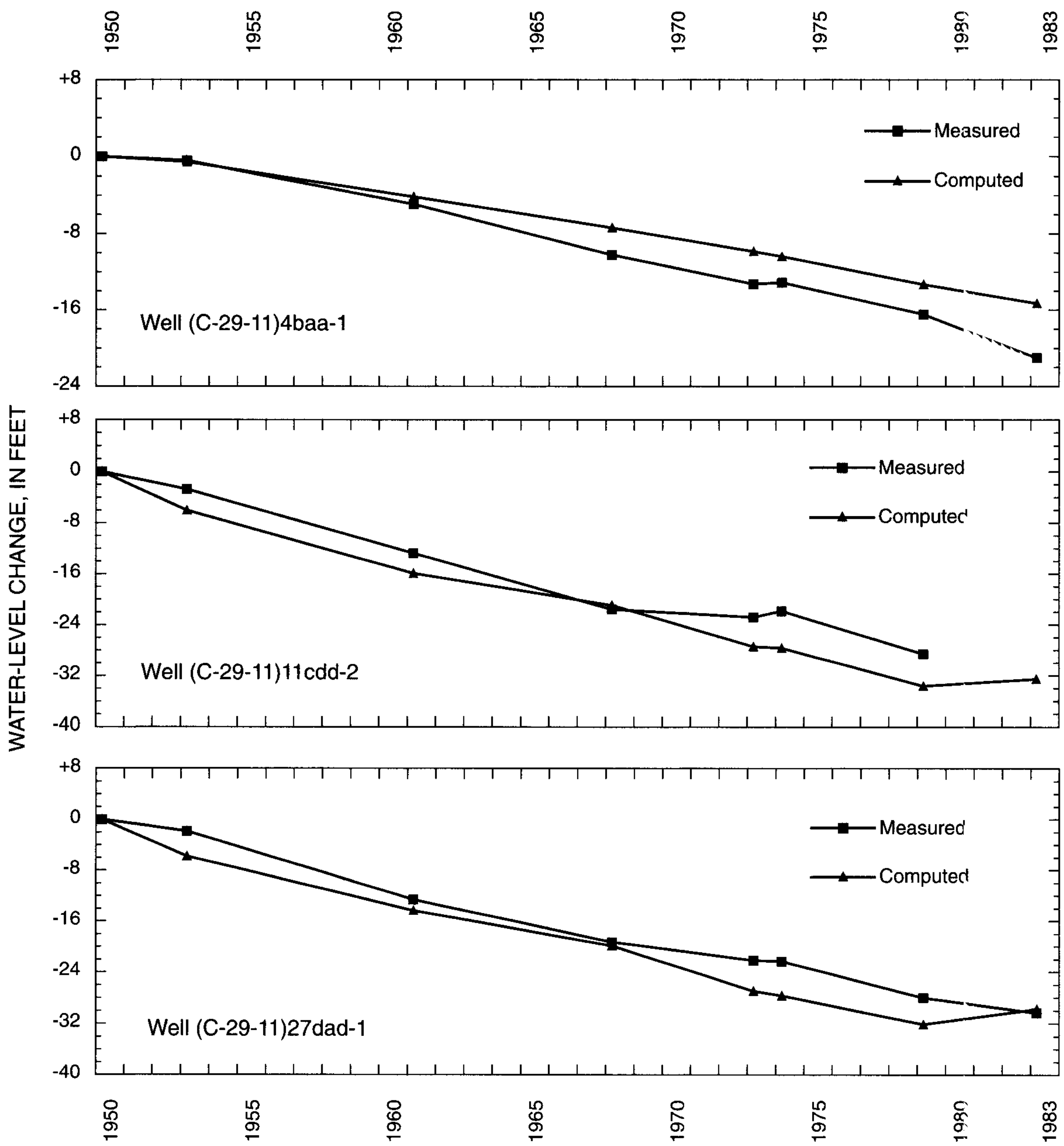

FIGURE 12.—Measured and computed water-level changes during 1950–83 for 13 observation wells in the Milford area—Continued.

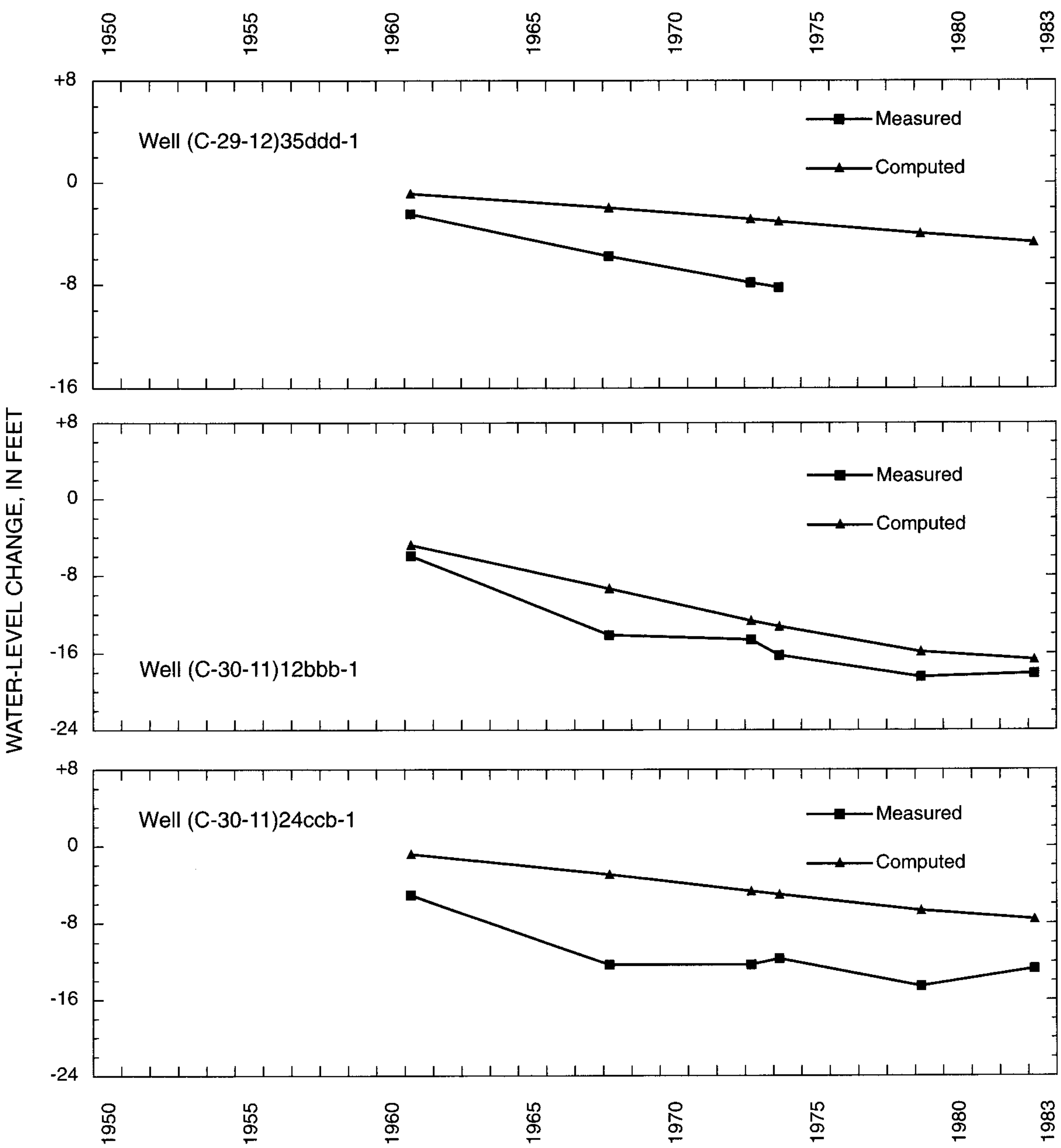

FIGURE 12.—Measured and computed water-level changes during 1950–83 for 13 observation wells in the Milford area—Continued.

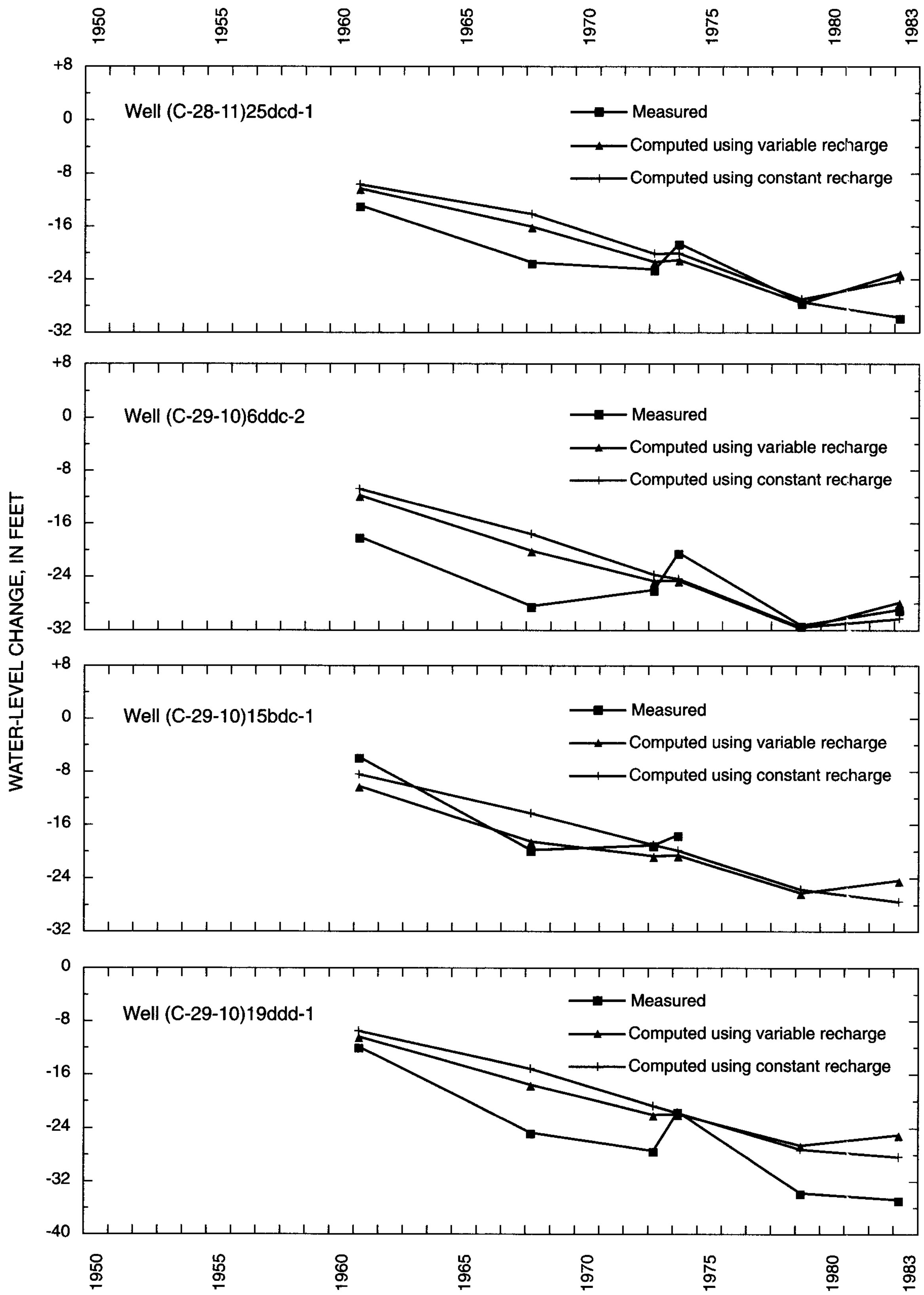

FIGURE 13.—Measured and computed water-level changes during 1950–83 for seven observation wells in the Milford area that show minimal differences (2 ft or less) in computed levels between constant and variable recharge from seepage.

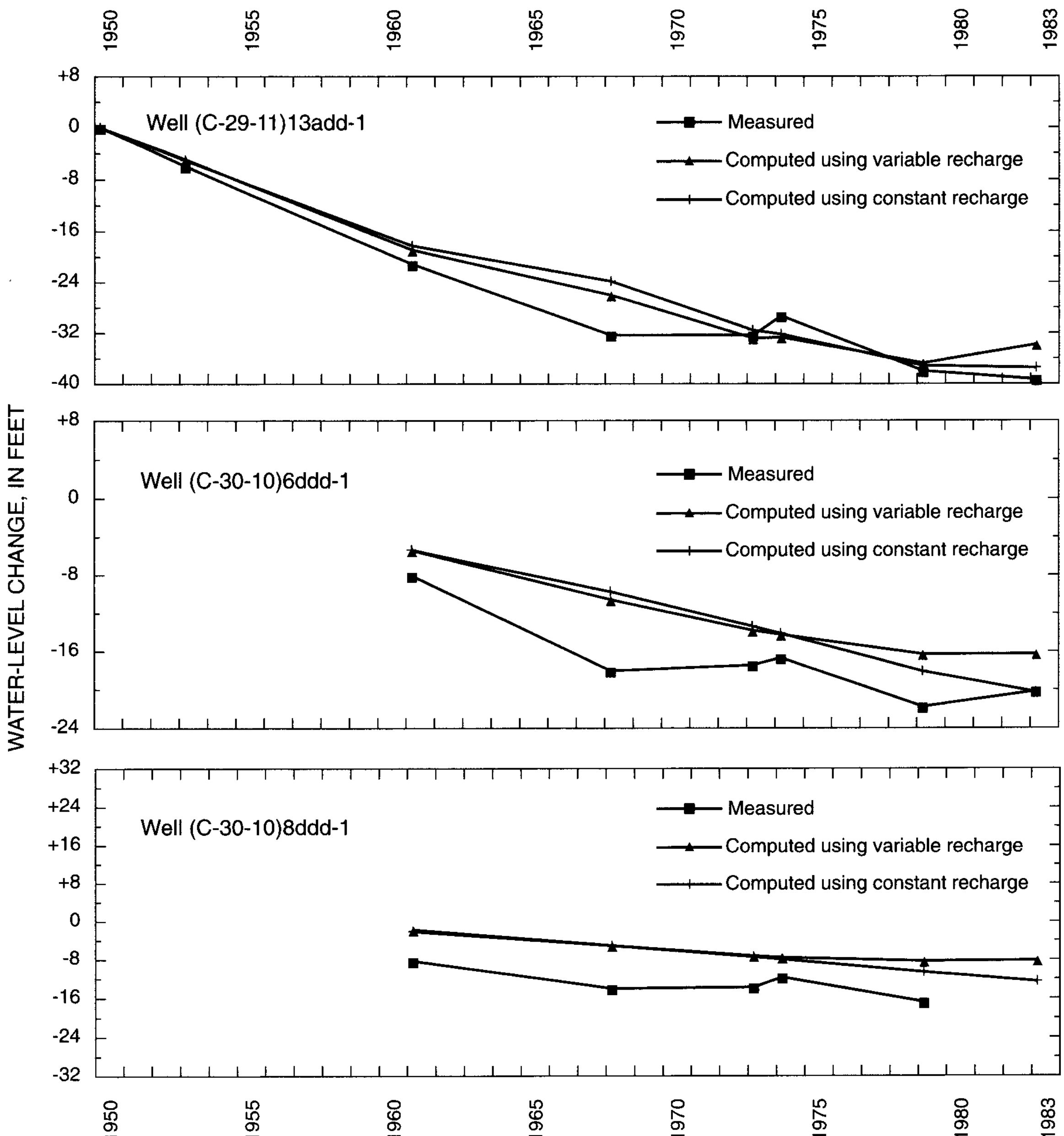

FIGURE 13.—Measured and computed water-level changes during 1950–83 for seven observation wells in the Milford area that show minimal differences (2 ft or less) in computed levels between constant and variable recharge from seepage—Continued.

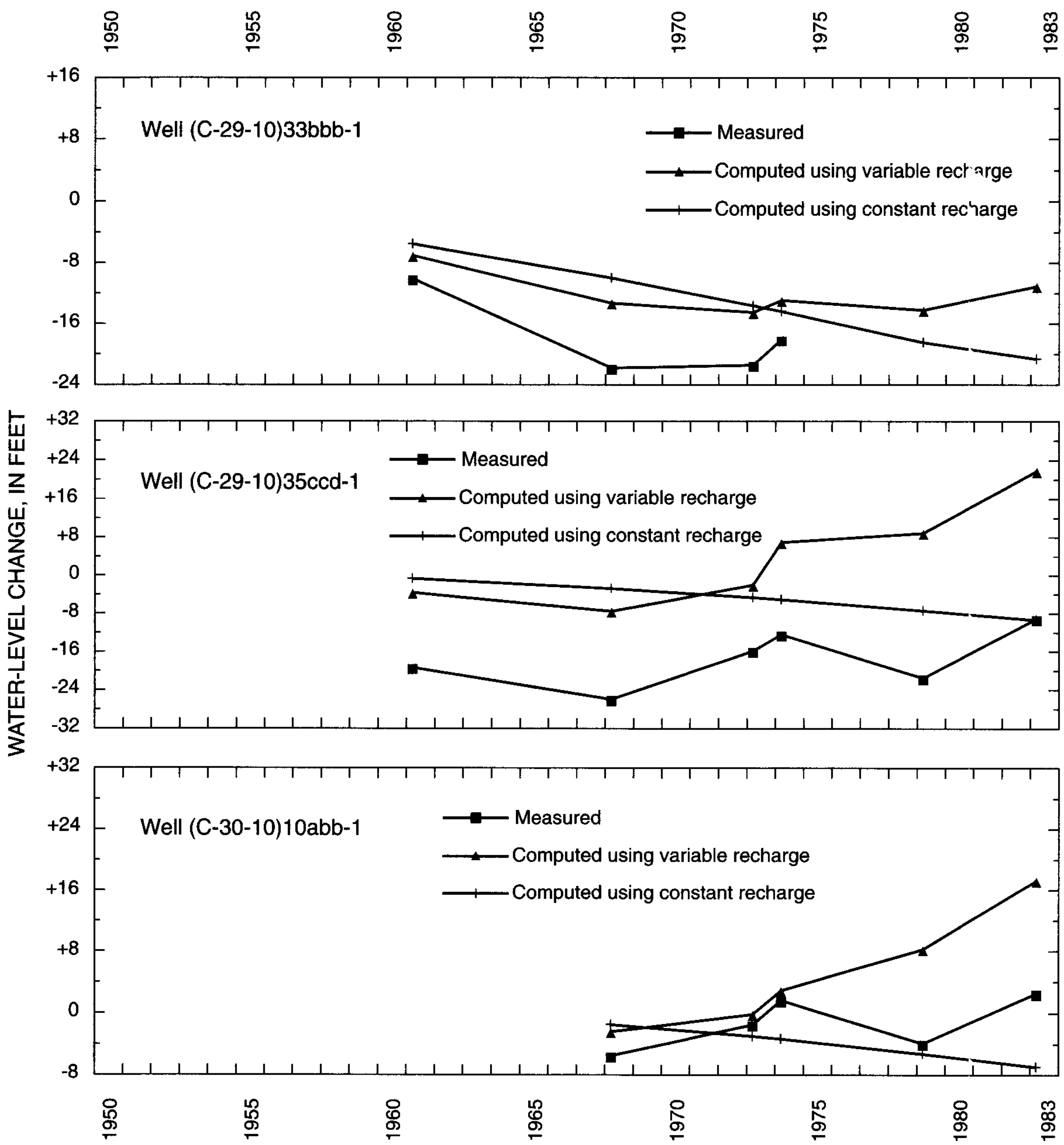

FIGURE 14.—Measured and computed water-level changes during 1950–83 for three observation wells in the Milford area that show substantial differences (more than 2 ft) in computed levels between constant and variable recharge from seepage.

TABLE 2.—*Simulated steady-state (1927) and transient-state (1979–82) ground-water budget for the Milford area, Utah*

[Data are in cubic feet per second; acre-feet per year shown in parentheses. Dashes (—) indicate not applicable]

Budget component	Steady-state (1927)	Transient-state (end of 1979-82 stress period)	
		Variable seepage	Constant seepage
Recharge:			
Subsurface inflow from consolidated rocks in the Mineral Mountains.	32.9 (23,800)	32.9 (23,800)	32.9 (23,800)
Subsurface inflow from adjoining areas:			
Beryl–Enterprise	2.67 (1,930)	3.52 (2,550)	3.44 (2,490)
Beaver Valley	5.02 (3,630)	5.02 (3,630)	5.02 (3,630)
Cove Fort area	2.37 (1,720)	2.37 (1,720)	2.37 (1,720)
Seepage from canals, streams, and unconsumed irrigation water.			
Unconsumed irrigation water derived from ground-water sources.	2.08 (1,500)	19.1 (13,800)	19.1 (13,800)
Unconsumed irrigation water derived from surface-water sources.	6.24 (4,520)	9.80 (7,100)	8.78 (6,360)
Seepage from canals	1.92 (1,390)	2.78 (2,010)	2.66 (1,930)
Seepage from the Beaver River	2.07 (1,500)	20.1 (14,500)	2.07 (1,500)
Infiltration from precipitation	4.77 (3,450)	4.77 (3,450)	4.77 (3,450)
Total recharge	60.0 (43,400)	100 (72,400)	81.1 (58,700)
Discharge:			
Wells	6.95 (5,000)	63.6 (46,000)	63.6 (46,000)
Evapotranspiration	37.2 (26,900)	16.8 (12,200)	17.0 (12,300)
Subsurface outflow to adjoining areas on northwestern boundary.	15.8 (11,400)	15.2 (11,000)	15.2 (11,000)
Total discharge	60.0 (43,400)	95.6 (69,200)	95.8 (69,400)
Change in storage:			
Water entered into storage	—	13.8 (9,990)	0.45 (325)
Water removed from storage	—	9.00 (6,520)	15.1 (10,900)
Net change in storage equals difference between water entered into storage and water removed from storage.	—	-4.8 (-3,480)	14.6 (10,600)

These changes are considered to be a reasonable estimate of error for each parameter, although the estimate of error may be much greater in areas with few data. After each simulation, the average difference between computed and calibrated steady-state heads was determined for each layer in a 309-cell area (fig. 15). The average differences obtained using the adjusted data were compared to the average differences that existed in the calibrated model between the calibrated steady-state heads and the initial heads. In addition, the new head-dependent fluxes were compared to the fluxes computed by the calibrated model. The results of all simulations are summarized in table 3.

The largest changes in computed-head distributions were due to variations in recharge at the eastern boundary, maximum evapotranspiration rates, and evapotranspiration extinction depths; however, the difference between the newly computed and calibrated steady-state heads within the 309-cell area show relatively minor changes compared to the calibrated-model values. The largest changes in flux at head-dependent boundaries were due to variations in recharge at the eastern boundary, maximum evapotranspiration rates, and extinction depths, in addition to transmissivity. Variations in recharge caused substantial changes in evapotranspiration. Variations in evapotranspiration rates and extinction depths changed the flux at the basin inflow boundary and variations in transmissivity caused changes in flux at both the inflow and outflow boundaries. In conclusion, the flux at head-dependent boundaries seems to be moderately sensitive to variations in some of the data. Consequently, even though the changes in computed head that result from errors in estimation of data are relatively small, the model-calculated water-budget components might change.

LIMITATIONS OF MODEL

The limitation to the ground-water model of the Milford area is the uncertainty of water levels and values for hydraulic properties in the northern one-half of the basin and along the margins of the basin. Because of the uncertainty in the potentiometric surface and hydraulic properties, our understanding of ground-water flow in these parts of the basin is limited. Constant-head cells were used initially along the eastern recharge boundary to determine the flux entering the system from the consolidated rocks and any uncertainty in the values for head and hydraulic conductivity would lead to uncertainty in the computed flux. By maintaining all parameters within reasonable limits, the computed flux along the recharge boundary was similar to the value estimated previously.

Steep hydraulic gradients and larger cell size in the southeast and the northwest parts of the simulated area cause some differences between initial and computed heads. These values were much larger than the generally accepted calibration limits. Although the computed heads were considerably different from measured heads, the initial and computed hydraulic gradients in these areas are similar. Consequently, during transient simulations, computed head changes between stress periods were similar to actual head changes in the southeast. If future model simulations were to consider the effects of seepage from surface-water irrigation in the southeast part of the simulated area and more data were available, then model cell size in that part of the basin could be made smaller to ensure greater accuracy.

On the basis of canal-loss studies, 30 percent of all surface water and ground water used for irrigation was assumed to seep into the ground-water system. In making this simplistic assumption, variations in seepage due to the use of different methods of irrigation and differences in soil conditions were not taken into account. This assumption probably does not make a substantial difference in the overall accuracy of the calibrated model, but could make a difference in the accuracy of a computed head of a specific cell.

The effects on layer 1 due to increased withdrawals in layer 2 could not be tested accurately without additional water-level and aquifer-test data from wells completed only in deeper zones of the basin-fill aquifer. These data would be necessary to define and calibrate the vertical-head gradient and flux within the area of ground-water development.

SHORT-TERM PREDICTIVE SIMULATIONS

The model of the Milford area was used to project the effects of ground-water withdrawals from 1983 to 2020. Computed water levels at the end of the 1979–82 stress period, derived from the historical transient simulation, were used as the starting point for the short-term predictive simulations. As in the transient simulation, average values for seepage were used for all stress periods. Because the model was designed principally to simulate relative declines and possible trends in ground-water levels, no attempt was made to simulate future variations of flow in the Beaver River and its associated irrigation-canal system. Also, the extremely high flows in the Beaver River for 1983 and 1984 were not simulated.

The model was used to simulate the response of the ground-water system to three rates of ground-water withdrawal. The areal and vertical distributions of the withdrawals for 1979–82 were used for all simulations. In

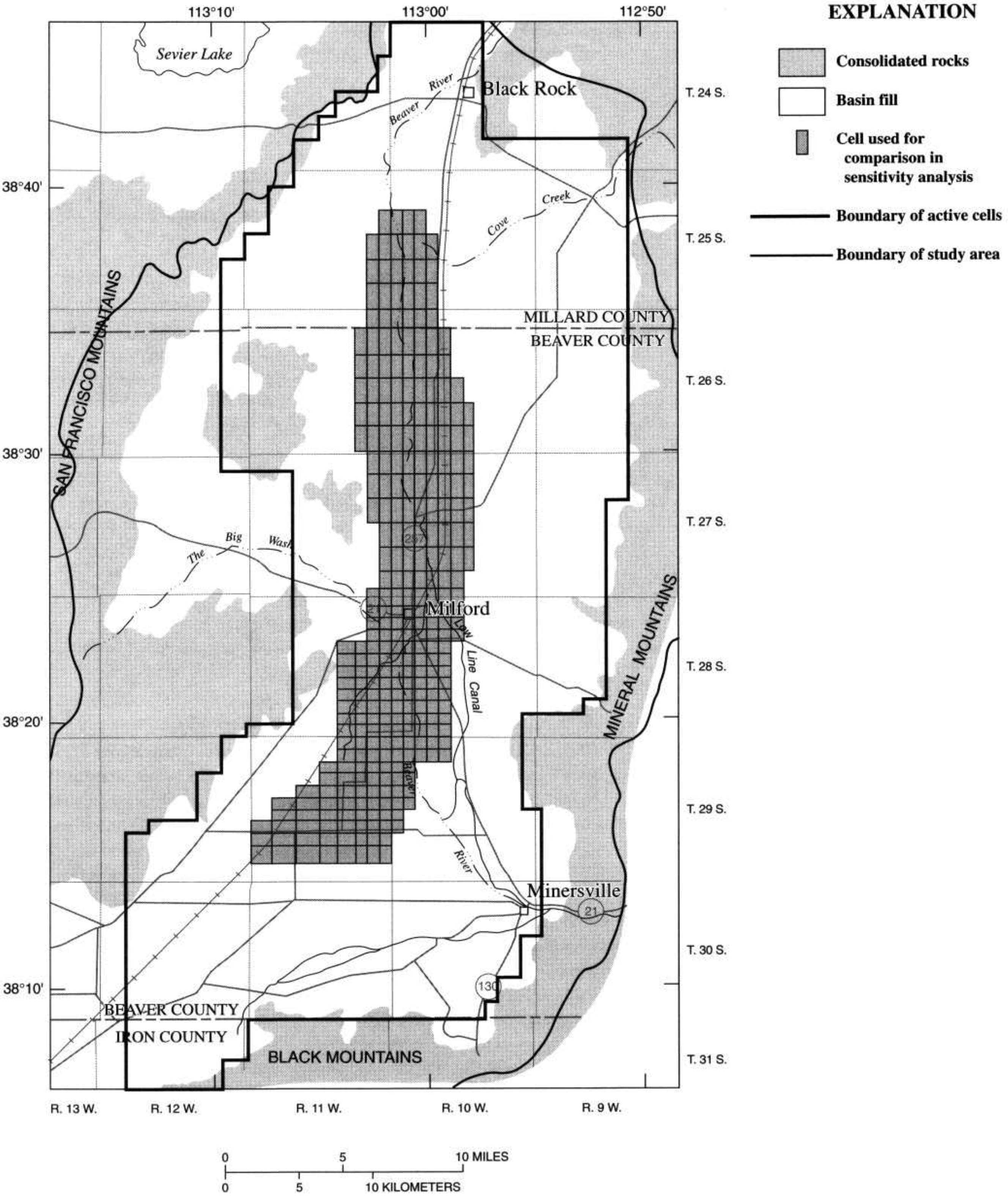

FIGURE 15.—Boundary of active cells during model simulations and areal distribution of cells used for comparison during sensitivity analysis.

TABLE 3.—*Results of sensitivity analysis for ground-water model, Milford area, Utah*

[Percent change of new computed value, compared to calibrated model given in parentheses]

Hydraulic property	Percent change	Difference between initial and computed head difference[1] (feet)						Flux at head-dependent boundaries (cubic feet per second)					
		Layer 1		Layer 2		Layer 3		Basin inflow		Basin outflow		Evapotranspiration	
Calibrated model	0	1.63	(0.0)	1.76	(0.0)	1.53	(0.0)	2.67	(0.0)	15.82	(0.0)	37.25	(0.0)
Transmissivity	-20	0.41	(74.8)	0.08	(95.5)	-0.18	(111.0)	2.48	(7.1)	13.76	(13.0)	39.12	(5.0)
(all layers).	+20	2.61	(60.1)	3.12	(77.3)	2.91	(90.2)	2.85	(6.7)	17.60	(11.3)	35.66	(4.3)
Hydraulic conductivity	-20	1.11	(31.9)	0.77	(56.3)	0.52	(66.0)	2.57	(3.7)	15.33	(3.1)	37.65	(1.1)
(layer 1).	+20	2.04	(25.2)	2.56	(45.5)	2.35	(53.6)	2.77	(3.7)	16.32	(3.2)	36.86	(1.0)
Transmissivity	-20	1.09	(33.1)	1.25	(29.0)	0.99	(35.3)	2.61	(2.2)	14.63	(7.5)	38.37	(3.0)
(layer 2).	+20	2.11	(29.4)	2.24	(27.3)	2.05	(34.0)	2.75	(3.0)	16.89	(6.8)	36.26	(2.7)
Transmissivity	-20	1.50	(8.0)	1.63	(7.4)	1.43	(6.5)	2.65	(0.7)	15.54	(1.8)	37.52	(0.7)
(layer 3).	+20	1.75	(7.4)	1.89	(7.4)	1.63	(6.5)	2.69	(0.7)	16.10	(1.8)	36.99	(0.7)
Vertical conductivity	-20	1.60	(1.8)	1.28	(41.3)	1.02	(33.3)	2.67	(0.0)	15.87	(0.3)	37.20	(0.1)
(all layers).	+20	1.65	(1.2)	2.13	(21.0)	1.92	(25.5)	2.68	(0.4)	15.78	(0.3)	37.30	(0.1)
Vertical conductivity	-20	1.60	(1.8)	1.27	(27.8)	1.07	(30.1)	2.67	(0.0)	15.87	(0.3)	37.20	(0.1)
(layers 1–2).	+20	1.65	(1.2)	2.13	(21.0)	1.89	(23.5)	2.68	(0.4)	15.78	(0.3)	37.29	(0.1)
Vertical conductivity	-20	1.63	(0.0)	1.76	(0.0)	1.49	(2.0)	2.67	(0.0)	15.82	(0.0)	37.25	(0.0)
(layers 2–3).	+20	1.63	(0.0)	1.76	(0.0)	1.56	(2.0)	2.67	(0.0)	15.82	(0.0)	37.25	(0.0)
Recharge	-20	4.67	(186)	5.26	(199)	5.06	(231)	2.76	(3.4)	15.04	(4.9)	30.60	(17.9)
	+20	-1.40	(186)	-1.70	(197)	-1.95	(228)	2.64	(1.1)	16.57	(4.7)	43.98	(18.1)
Maximum	-20	-1.65	(201)	-1.34	(176)	-1.56	(202)	2.51	(6.0)	16.38	(3.5)	36.54	(1.9)
evapotranspiration.	+20	3.82	(134)	3.83	(118)	3.59	(135)	2.82	(5.6)	15.48	(2.1)	37.75	(1.3)
Evapotranspiration	-20	-1.19	(173)	-1.07	(161)	-1.29	(184)	2.41	(9.7)	16.37	(3.5)	36.43	(2.2)
extinction depth.	+20	4.39	(169)	4.53	(157)	4.30	(181)	2.96	(10.9)	15.33	(3.1)	38.04	(2.1)

[1]Average of the absolute difference between the initial-head distribution and the computed-head distribution in the 309-cell area where enough data were available to make a reasonable estimate of steady-state water levels or where the estimated hydraulic gradient was small (fig. 15). Negative values for difference in head indicate computed head is above initial head.

the first simulation, withdrawals equal to the 1979–82 average rate caused water-level declines of more than 12 ft near the south end of the Mineral Mountains and declines of 6 to 10 ft in the area of pumping (fig. 16). The smaller water-level declines in the area of pumping can be attributed to decreased evapotranspiration and to decreased storage depletions (table 4), indicating that the ground-water system could conceivably be approaching a new equilibrium condition. The projected water-level declines along the eastern margin of the simulated area are a continuation of the boundary effects that were simulated at the end of the historical transient simulation.

In the second simulation, ground-water withdrawals were increased to 1.5 times the 1979–82 average rate. This rate of ground-water withdrawal is about 15 percent larger than the maximum average rate applied in the historical, transient simulation for 1974–78 (table 1) and it is equal to the largest annual rate of withdrawal reported for 1974 (fig. 6). Although long-term ground-water withdrawals probably would not remain this large, it could approach this level for short periods as it has in recent years. This simulation resulted in projected water-level declines of more than 35 ft at the center of a well-defined cone of depression that covered the entire southern one-half of the basin (fig. 17). As would be expected with the extent of projected water-level declines, evapotranspiration decreased and storage depletion and basin inflow at the southwest boundary increased (table 4). Minor water-level rises of less than 1 ft were projected for the extreme north end of the basin. These rises probably are due to minor flux imbalances at the northern boundary.

The third simulation used ground-water withdrawals at double the 1979–82 rate. This simulation projected water-level declines of more than 70 ft at the center of a well-defined cone of depression (fig. 18). Water-level declines of more than 40 ft were projected at the eastern and western boundaries of the basin. Storage depletion becomes a large component in the water budget at this rate of withdrawal (table 4). Pumping at this rate with the current (1984) distribution of wells would be a worst-case possibility. Pumping could not approach this rate without considerable development in the north part of the basin. If substantial development did occur in the north part of the basin, the overall water-level declines would be less because withdrawals would be distributed throughout the basin rather than being restricted to the southern one-half.

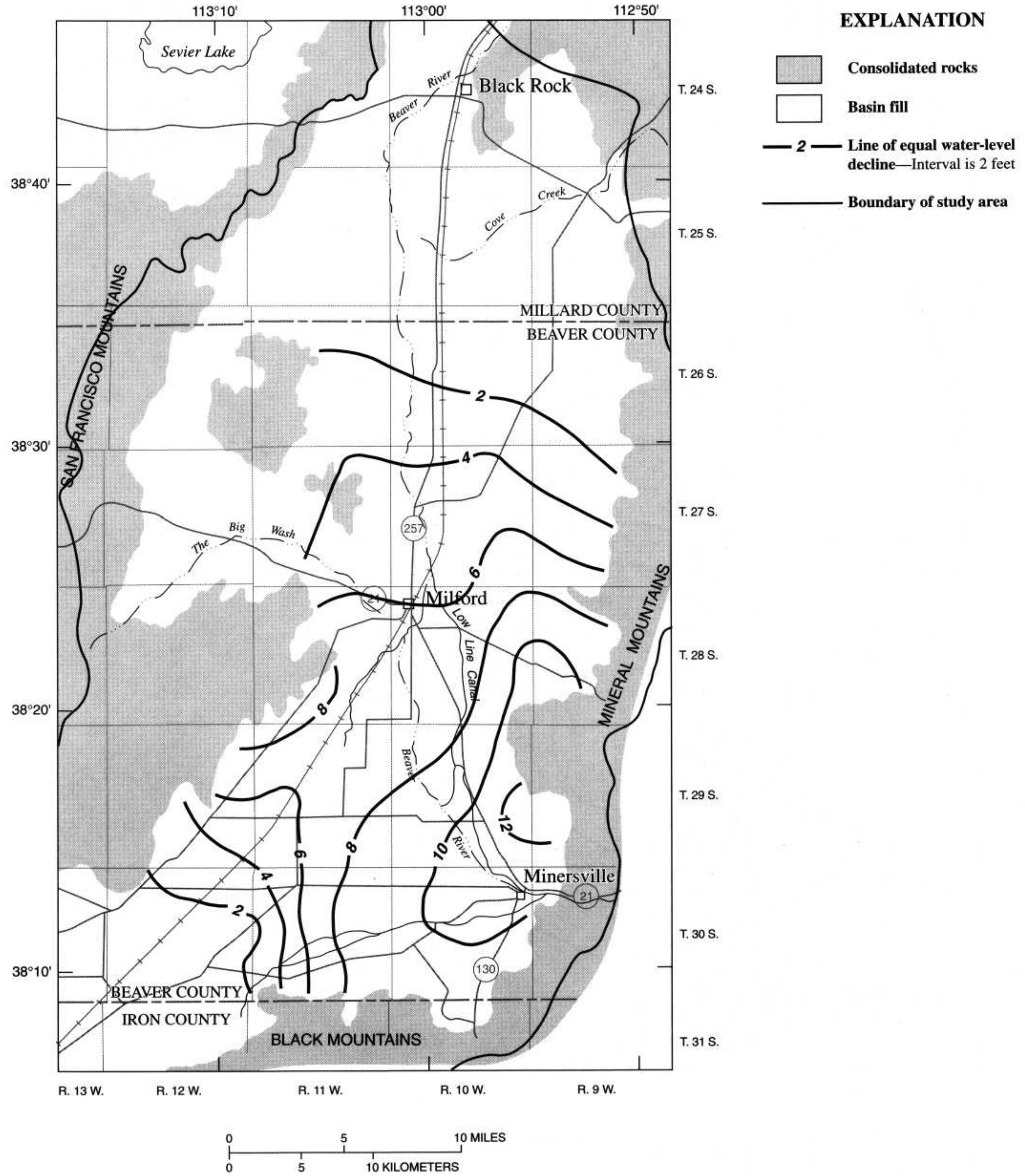

FIGURE 16.—Projected water-level declines in the basin-fill aquifer for 1983-2020, assuming ground-water withdrawals equal to the 1979-82 average rate.

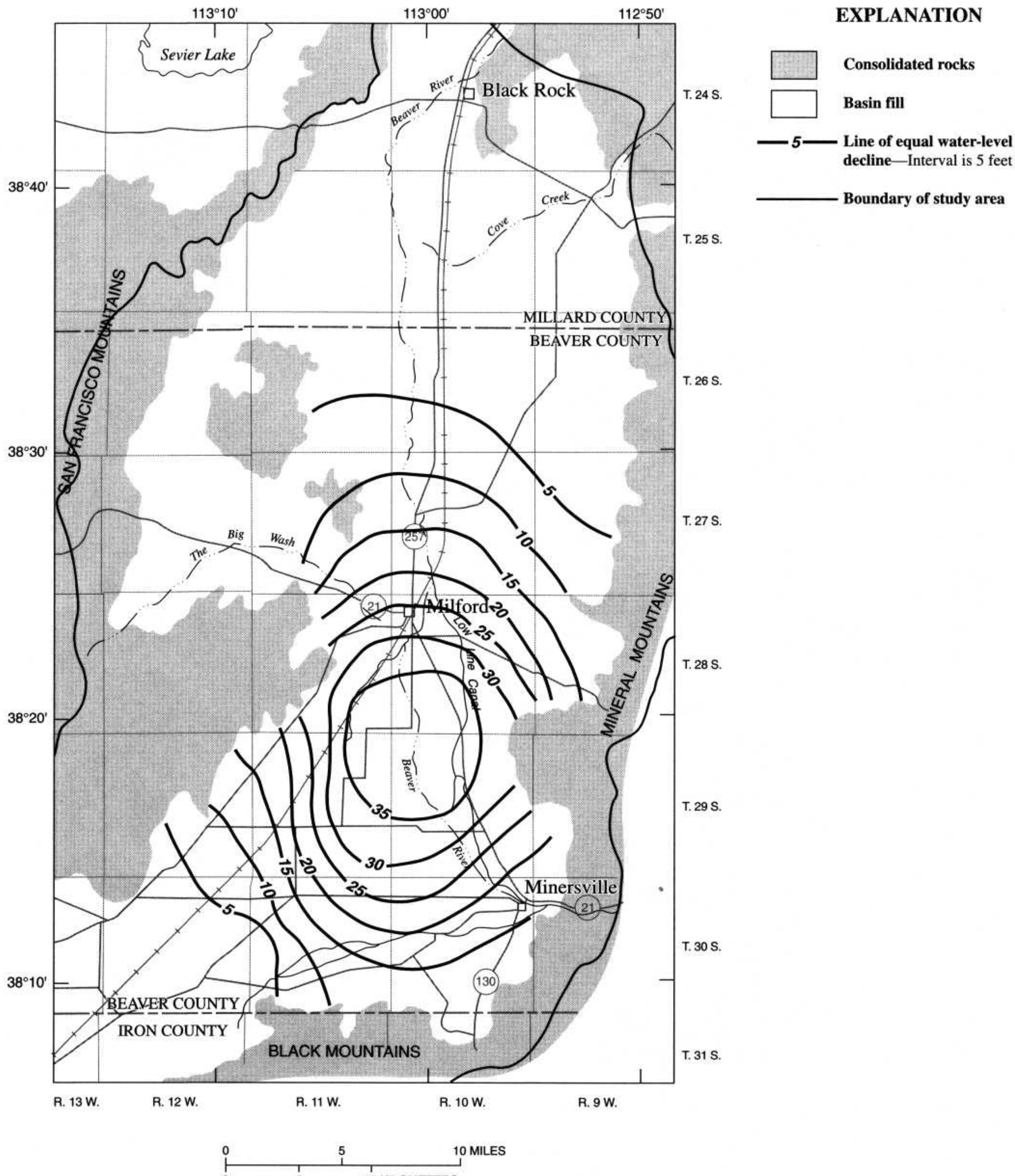

FIGURE 17.—Projected water-level declines in the basin-fill aquifer for 1983-2020, assuming ground-water withdrawals equal to 1.5 times the 1979-82 average rate.

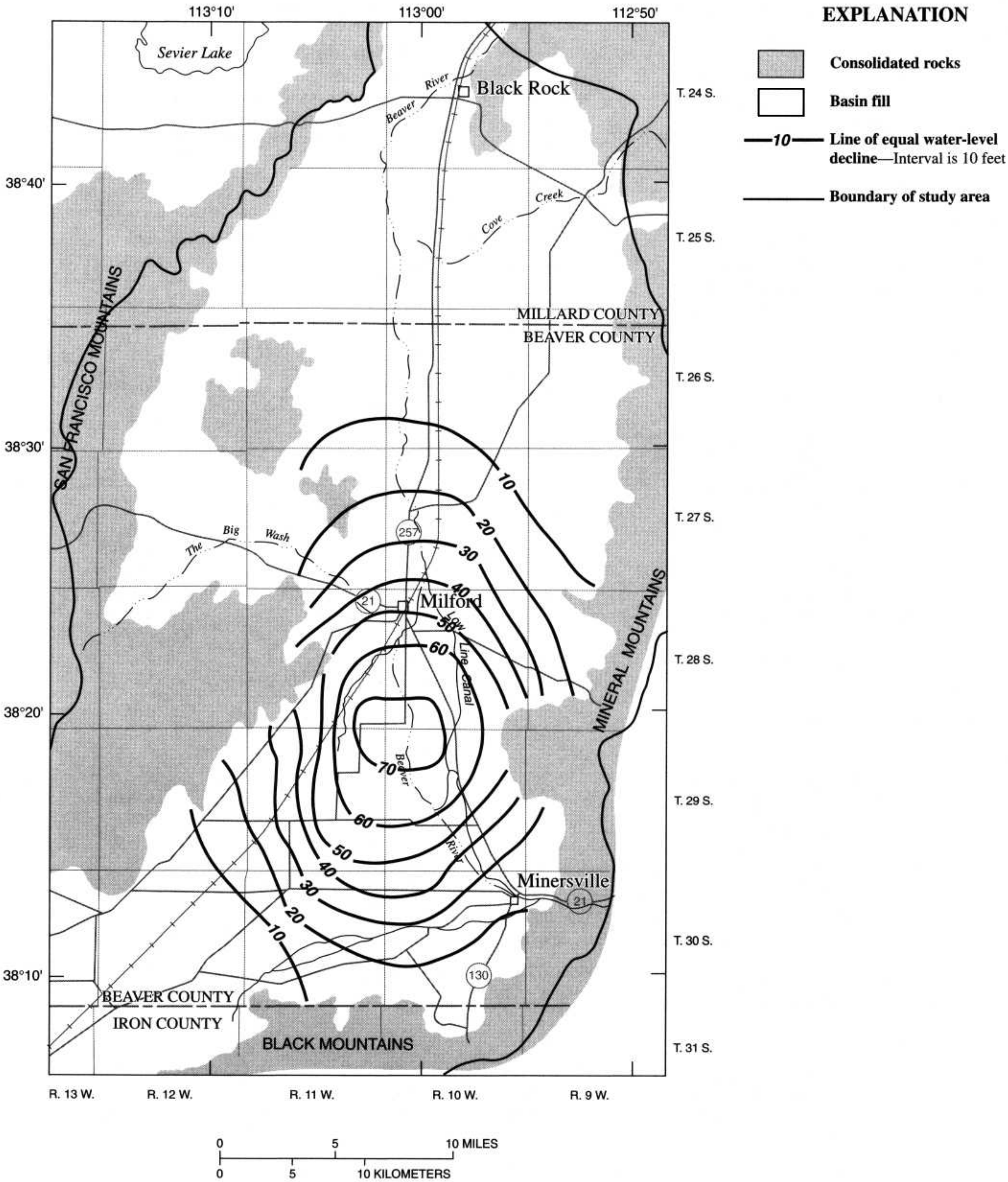

FIGURE 18.—Projected water-level declines in the basin-fill aquifer for 1983-2020, assuming ground-water withdrawals double the 1979-82 average rate.

TABLE 4.—*Projected changes in ground-water budget components due to increased ground-water withdrawals, Milford area, Utah*

[Data are in cubic feet per second; acre-feet per year shown in parentheses]

Budget component	Transient-state (end of 1979–82 stress period) average seepage	Projected water-budget rates at year 2020 using 1979–82 average ground-water withdrawal times		
		1.0	1.5	2.0
Recharge:				
Subsurface inflow from consolidated rocks in the Mineral Mountains.	32.9 (23,800)	32.9 (23,800)	32.9 (23,800)	32.9 (23,800)
Subsurface inflow from adjoining areas:				
Beryl–Enterprise	3.44 (2,490)	4.41 (3,190)	5.85 (4,240)	7.34 (5,310)
Beaver Valley	5.02 (3,630)	5.02 (3,630)	5.02 (3,630)	5.02 (3,630)
Cove Fort area	2.37 (1,720)	2.37 (1,720)	2.37 (1,720)	2.37 (1,720)
Seepage from streams, canals, and unconsumed irrigation water.				
Seepage from the Beaver River	2.07 (1,500)	2.07 (1,500)	2.07 (1,500)	2.07 (1,500)
Seepage from canals	2.66 (1,930)	2.66 (1,930)	2.66 (1,930)	2.66 (1,930)
Unconsumed irrigation water derived from ground-water sources.	19.1 (13,800)	19.1 (13,800)	28.6 (20,700)	38.1 (27,600)
Unconsumed irrigation water derived from surface-water sources.	8.78 (6,360)	8.78 (6,360)	8.78 (6,360)	8.78 (6,360)
Infiltration from precipitation	4.77 (3,450)	4.77 (3,450)	4.77 (3,450)	4.77 (3,450)
Total recharge	81.1 (58,700)	82.1 (59,400)	93.0 (67,300)	104 (75,300)
Discharge:				
Subsurface outflow to adjoining areas on northwestern boundary.	15.2 (11,000)	15.1 (10,900)	15.1 (10,900)	15.1 (10,900)
Wells	63.6 (46,000)	63.6 (46,000)	95.4 (69,100)	127 (92,000)
Evapotranspiration	17.0 (12,300)	13.3 (9,630)	10.4 (7,530)	8.88 (5,950)
Total discharge	95.8 (69,400)	92.0 (66,600)	121 (87,600)	151 (109,000)
Change in storage:				
Water entered into storage	0.45 (325)	0.0	0.0	0.0
Water removed from storage	15.1 (10,900)	9.93 (7,200)	27.8 (20,100)	47.0 (34,000)
Net change in storage equals difference between water entered into storage and water removed from storage.	14.6 (10,600)	9.93 (7,200)	27.8 (20,100)	47.0 (34,000)