Excerpt from McDonald and Harbaugh (1988), page 5-33 specific yields must not be included in the input data; if they are included, the data sequence will be misread. Note that erroneous specification of ISS or of a LAYCON value will also cause misreading of the data array sequence.

Four types of layer are recognized by the model, incoporating various combinations of the options provided by the Block-Centered-Flow Package. These four layer types are identified by their layer-type codes, which are stored in the one-dimensional array LAYCON (NLAY). The code values and the corresponding layer characteristics are given below.

Layer-type 0--In this category there is no provision for modification of transmissivity as water level varies, for storage term conversion, or for limitation of vertical flow from above if water level falls below the top of the cell. This layer type is normally used to simulate confined conditions, but could also be used to simulate a layer in which unconfined conditions will always prevail, provided drawdowns are expected to be a small fraction of layer thickness and flow from the overlying layer (if present) is expected to be negligible. If the simulation is transient, storage coefficient or specific yield values are entered in the input array sf1(NCOL, NROW); then row-direction transmissivities are entered in the input array Tran (NCOL, NROW); and following the transmissivities, unless the layer is the lowermost in the model, vertical leakance values are entered in the input array Vcont (NCOL, NROW). Again, parameter values may be specified by providing the entire array, or by providing a single default value which is applied to all cells of the layer. The parameter values assigned at the beginning of a simulation in this type of layer are retained without change throughout the simulation.

Excerpt from Schaefer and Harrill (1995), page 8

In keeping with the conceptual nature of the model, the simulation provides information about the probable areas that may be affected, the general magnitude of possible water-level declines or other effects, and the general period of time over which changes may be expected to occur. Prediction of specific, detailed water-level changes throughout the area would require that effects of the proposed pumping be superimposed on the effects of existing and other anticipated future pumping. That was beyond the scope of this analysis.

The second assumption was that storage values used for transient simulations for the upper layer were based on the predominant aquifer material in each cell, determined from surficial maps. This distribution may not be totally correct because the material may be different at depth in the zone of saturation. Storage coefficients in the upper layer also assume dewatering of the sediments.

Rock and deposit types were divided into three categories—basin-fill materials, carbonate rocks, and other consolidated rocks. Distribution of these units is shown by Prudic and others (1993, fig. 15). Average values for storage coefficients in layer one were assigned to each of these materials. For basin-fill material, a value of 0.1 was assigned on the basis of average values of specific yield used in U.S. Geological Survey reconnaissance evaluations of ground-water resources in most basins of the study area. For carbonate rocks, a value of 0.05 was assigned on the basis of an average porosity value of 0.047 determined from geophysical logs of five wells in the Coyote Spring Valley area (Berger, 1992, p. 18). For other rocks, a value of 0.01 was assigned on the basis of a range of yalues for fractured rocks given by Snow (1979, table 1).

The storage coefficient for the lower layer was estimated on the basis of the probable average porosity of the rocks present (0.01 to 0.05), the effective thickness of aquifer material (probably between 5,000 and 10,000 ft), the bulk modulus of elasticity of water (3×10^5 lb/in^2), and the bulk modulus of elasticity of the solid skeleton of the aquifer (for limestone, about 4.8×10^6 to 5.4×10^6 lb/in^2; Krynine and Judd, 1957, table 2.5). The following equation from Lohman (1972, p. 9) was used to estimate the coefficients:

$$S = \theta \gamma b \left(\frac{1}{E_w} + \frac{C}{\theta E_s} \right), \qquad (1)$$

where S is storage coefficient (dimensionless);
θ is porosity, as a decimal fraction;
γ is specific weight per unit, 62.4 lb/ft^3 ÷ 144 in^2/ft^2 = 0.434 (lb/in^2)/ft;
b is thickness, in feet;
E_w is bulk modulus of elasticity of water;
C is a dimensionless ratio, which may be considered unity in an uncemented granular material; in a solid aquifer, such as limestone with tubular solution channels, C is apparently equal to porosity; and
E_s is bulk modulus of elasticity of the solid skeleton of an aquifer.

Estimates of storage values based on the above numbers ranged from 7.6×10^{-5} to 1.2×10^{-3}. For purposes of this report, the storage coefficient for the lower layer was set at the midrange of these values, 6×10^{-4}, for the entire layer. The data set for storage values used in the model is listed in appendix 1.

The third major assumption used in the model is from the previous steady-state model and concerns the lower layer. The individual basin-fill aquifers underlying the various ground-water basins can be adequately described in the upper layer as a series of high-transmissivity zones (the basin-fill valleys) separated from each other by low-transmissivity zones (the intervening mountain ranges). The lower layer represents the distribution of carbonate-rock aquifers in the system in a limited way that may affect the calculated drawdowns in that layer.

The fourth and final assumption was that all input values used in the conceptual steady-state model remain constant during the transient simulations. No changes were made to transmissivity, leakance, recharge, or the other input data sets described by Prudic and others (1993) and Schaefer (1993).

RESULTS OF SIMULATIONS

Simulation of Conditions Prior to Proposed Pumping

The steady-state conditions simulated by Prudic and others (1993) represent a conceptualization of ground-water flow in the carbonate-rock province of the Great Basin before ground-water pumping within the province commenced. Figure 2 shows the general distribution of simulated steady-state heads (water

Excerpt from McDonald and Harbaugh (1988), pages 5-11 and 5-12

Vertical Conductance Formulation

Vertical conductance terms are calculated within the model using data from an input array which incorporates both thickness and vertical hydraulic conductivity in a single term, and using horizontal (or map) areas calculated from cell dimensions. In general, the vertical interval between two nodes, i,j,k and and i,j,k+1, may be considered to contain n geohydrologic layers or units, having vertical hydraulic conductivities $K_1, K_2 \ldots . K_n$ and thicknesses $\Delta z_1, \Delta z_2 \ldots . \Delta z_n$. The map area of the cells around nodes i,j,k and i,j,k+1 is $DELR_j*DELC_i$; the vertical conductance of an individual geohydrologic layer, g, in this area is given by

$$C_g = \frac{K_g \, DELR_j*DELC_i}{\Delta z_g} \tag{46}$$

The equivalent vertical conductance, $C_{i,j,k+1/2}$, for the full vertical interval between nodes i,j,k and i,j,k+1 is found by treating the n individual geohydrologic layers as conductances in series; this yields

$$\frac{1}{C_{i,j,k+1/2}} = \sum_{g=1}^{n} \frac{1}{C_g} =$$

$$\sum_{g=1}^{n} \frac{1}{\frac{K_g DELR_j*DELC_i}{\Delta z_g}} = \frac{1}{DELR_j*DELC_i} * \sum_{g=1}^{n} \frac{\Delta z_g}{K_g} \tag{47}$$

rearranging equation (47)

$$\frac{C_{i,j,k+1/2}}{DELR_j*DELC_i} = \frac{1}{\sum_{g=1}^{n} \frac{\Delta z_g}{K_g}} \tag{48}$$

The quantity $\frac{C_{i,j,k+1/2}}{DELR_j*DELC_i}$ has been termed the "vertical leakance " and is designated $Vcont_{i,j,k+1/2}$ in this report; thus we have

$$Vcont_{i,j,k+1/2} = \frac{1}{\sum_{g=1}^{n} \frac{\Delta z_g}{K_g}} \tag{49}$$

Vcont is the term actually used as input in the model described herein. That is, rather than specifying a total thickness and an equivalent (or harmonic mean) vertical hydraulic conductivity for the interval between node i,j,k and node i,j,k+1, the user specifies the term $Vcont_{i,j,k+1/2}$, which is actually the conductance of the interval divided by the cell area, and as such incorporates both hydraulic conductivity and thickness. The program multiples Vcont by cell area to obtain vertical conductance. The values of Vcont must be calculated or determined externally to the program; this is generally done through an application of equation (49). The Vcont values are actually read as the elements of a two-dimensional input array, $Vcont_{i,j}$, for each layer. Each value of $Vcont_{i,j}$ is the vertical leakance for the interval between cell i,j,k and cell i,j,k+1--that is, for the interval between the layer for which the array is read, and the layer below it. It follows that the Vcont array is not read for the lowermost layer in the model. Although values of Vcont are thus read into the model through a series of two-dimensional input arrays, the discussion in this section will continue to be given in terms of three-dimensional array notation, $Vcont_{i,j,k+1/2}$, to emphasize the fact that the Vcont values refer to the intervals between layers.

Excerpt from Prudic et al (1995), pages D28 to D31

INITIAL ESTIMATES OF TRANSMISSIVITY AND LEAKANCE

Initial estimates of transmissivity for the upper model layer are grouped into three geologic units. The estimates were made to provide a starting point for the calibration process in which transmissivities were modified. The geologic units within the modeled area are grouped into three principal types (Harrill and others, 1988; Plume and Carlton, 1988): (1) basin fill, which includes Tertiary tuffs, terrigenous sediments, and Quaternary stream, alluvial fan, and lacustrine deposits; (2) thick sequences of carbonate rocks of Paleozoic and early Mesozoic age; and (3) other consolidated rocks, which include clastic sedimentary rocks, intrusive and extrusive igneous rocks, metamorphic rocks, and locally thick units of Tertiary clay and silt. Figure 15 shows how the principal rock types are distributed in the upper layer. The basin-and-range physiography can be easily distinguished with the resolution provided by the 5-mi by 7.5-mi grid.

Carbonate rocks are assumed to have the highest transmissivity. The initial transmissivity assigned to cells in the upper model layer representing carbonate rocks was 0.25 ft^2/s, within the range of values reported by Winograd and Thordarson (1975, table 3 and p. 73), Bunch and Harrill (1984, p. 119), and Plume (1989). Reported values range from about 0.002 ft^2/s (200 ft^2/d) to about 9 ft^2/s (800,000 ft^2/d). Initial transmissivity assigned to cells representing other consolidated rocks was 0.002 ft^2/s; the initial value assigned to cells representing basin fill was 0.02 ft^2/s, within the range of values presented by Winograd and Thordarson (1975, table 3) and Bunch and Harrill (1984, p. 115). A uniform value of 0.25 ft^2/s was initially assigned to all cells in the lower layer.

Transmissivities of each rock type actually vary widely due to either changes in thickness or differing hydrologic properties of the rocks. The transmissivities for each model cell changed during model calibration. The vertical resistance to ground-water flow is simulated in the model with a vertical leakance term. Vertical leakance is defined as the vertical hydraulic conductivity divided by length of flow path (Lohman, 1972, p. 30). A vertical leakance of 1×10^{-11} per second was initially assumed for all cells. No attempt was made to distinguish leakance values according to hydrogeologic conditions because of the uncertainty of the geologic units at depth and because of uncertainties in estimating the vertical hydraulic conductivity and the length of the flow path. The vertical leakances also changed during model calibration.

MODEL CALIBRATION

Initial model calibration began by assigning an estimated water level to each model cell. In many cells, particularly in the lower layer, the assigned water levels were interpolated and extrapolated from data many miles away. Transmissivities of cells in the upper and lower model layers and vertical leakances of cells between layers were initially adjusted on the basis of comparing simulated water levels to those assigned to the model cells. Two computer programs were written and used to automatically adjust both transmissivities and vertical leakances. The first program adjusted transmissivities in cells where the simulated water levels were either too high or too low compared to the assigned water levels. Transmissivities were increased or decreased depending on the ratio of the simulated water level to the assigned water level. The method worked reasonably well because simulated heads were either too high or too low over large regions of the model.

The second program adjusted vertical leakances between adjacent cells in the upper and lower model layers during alternate simulations. Vertical leakances were adjusted using the ratio of the simulated water-level difference to the assigned water-level difference as expressed in the following equation (Williamson and others, 1989, p. 32):

$$\text{Lnew} = \text{Lold} * \text{FAC} * (\Delta\text{HVmod}/\Delta\text{HVas})$$

where Lnew = the adjusted vertical leakance value;

Lold = the previous vertical leakance value;

ΔHVmod = the simulated water-level difference of adjacent cells between the upper and lower model layers;

ΔHVas = the assigned water-level difference of adjacent cells between the upper and lower model layers; and

FAC = 0.9 when the ratio of ΔHVmod to ΔHVas is less than 1, 1.1 when the ratio is greater than 1, and 1.0 when the ratio is 1.

The computer programs do not correctly adjust transmissivities or vertical leakances on the first

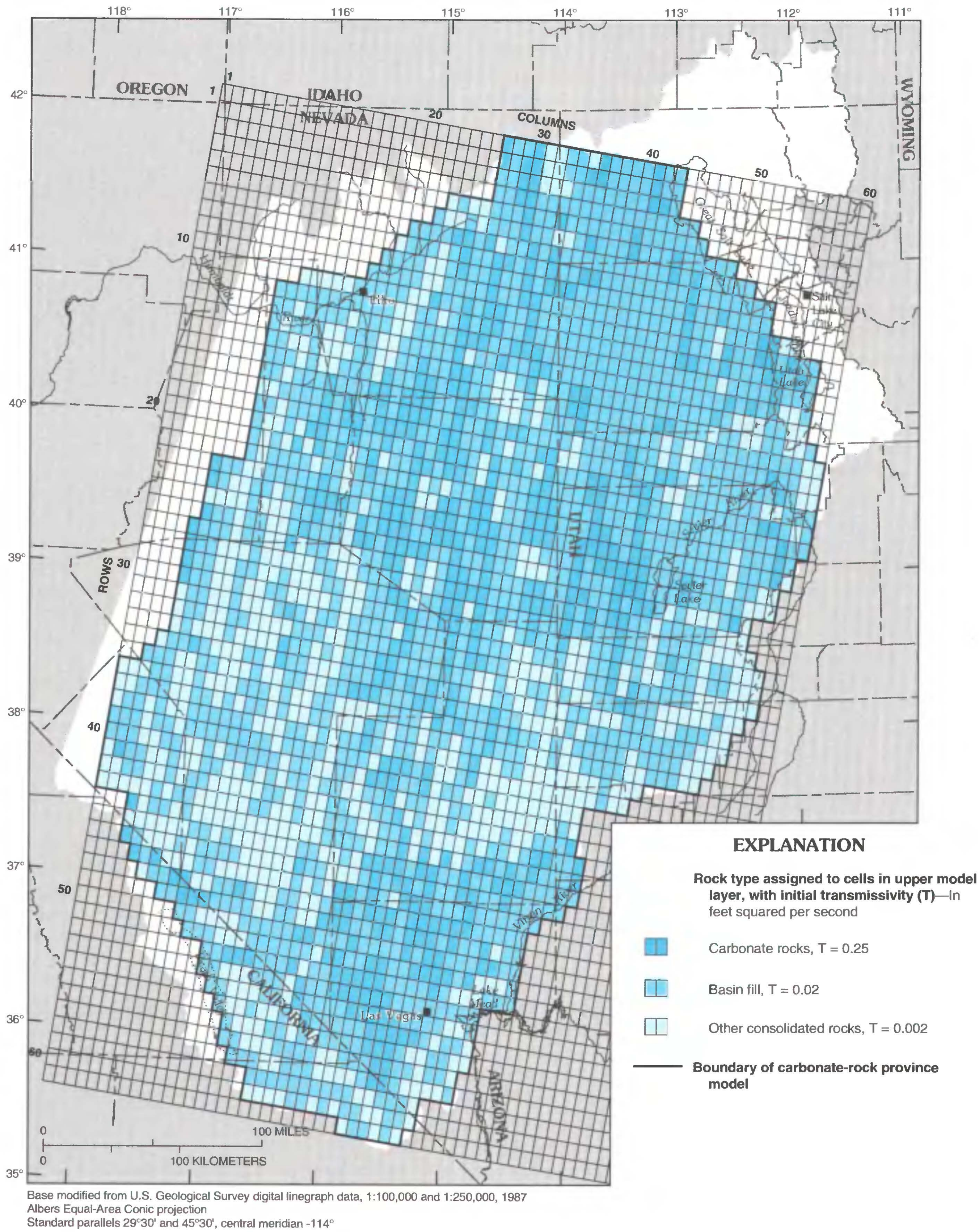


FIGURE 15.—Principal rock types assigned to cells in upper model layer, and initial transmissivities used.

computation because flow to and from a cell may change after adjusting the vertical leakance and the transmissivities in adjacent cells. Thus, the process involved numerous simulations that alternately adjusted transmissivities and vertical leakances. The use of these programs ceased once the simulated water levels over the entire model generally matched the water levels presented by Thomas and others (1986).

The final part of model calibration involved (1) testing the range in transmissivities and vertical leakances calculated from the initial calibration by comparing the simulated water levels in 773 selected cells in the upper layer and 144 cells in the lower layer where water levels had been estimated from the maps by Thomas and others (1986), (2) making regional and local changes to transmissivities and vertical leakances until simulated discharge as evapotranspiration in the upper model layer and regional spring flow in the lower layer approximated estimated values, and (3) adjusting conductance values at head-dependent flow boundaries.

Transmissivities following the initial calibration ranged from 2.5×10^{-4} to 2.5 ft^2/s in the upper layer and from 2.5×10^{-4} to 2.5×10^{-1} ft^2/s in the lower layer. During the final phase of model calibration, both transmissivities and vertical leakances were rounded to the nearest exponent (1×10^{-4}; 1×10^{-3}; 1×10^{-2}; and so forth) without affecting the simulation results. The rounding of both transmissivities and vertical leakances is reasonable because of the lack of information on the extent and distribution of aquifers, their hydraulic properties, and the lack of ground-water levels in many areas. Such groupings also simplified the final calibration while reasonably duplicating regional ground-water levels, and the distribution and quantity of discharge. The best match with estimated water levels and discharge was simulated when the grouped transmissivities were multiplied by a factor of 2.2 in the upper layer and when the values were multiplied by a factor of 3.3 in the lower layer. In a few areas, transmissivities were further multiplied by a factor ranging from 2 to 5. Even though transmissivities are generally grouped by a factor of 10, the range in simulated transmissivities did not change greatly from the initial calibration. In the upper layer, transmissivities following final calibration ranged from 2.2×10^{-5} to 2.2×10^{-1} ft^2/s; both the minimum and maximum values are about 10 times less than the initially calibrated values. In the lower layer, transmissivities following model calibration ranged from 3.3×10^{-5} to 6.6×10^{-1} ft^2/s.

Vertical leakances following initial calibration ranged from 1×10^{-16} to 3×10^{-9} per second. During final calibration, increasing vertical leakances of less than 1×10^{-13} to that value produced little difference in simulated water levels and discharge. Similarly, decreasing values greater than 1×10^{-11} to that value also produced little differences. Finally, all other leakance values were rounded to values of 1×10^{-11}, 1×10^{-12}, or 1×10^{-13} per second. The distribution of vertical leakances is shown in figure 16.

The average vertical leakance for all model cells is 4×10^{-12} per second. Overall, 62 percent of cells (1,517 of 2,456) have a value of 1×10^{-12} per second, 34 percent (833 cells) have a value of 1×10^{-11} per second, and only 4 percent (106 cells) have a value of 1×10^{-13} per second. Most of the cells (95 out of 106) having the lowest vertical leakances are in or adjacent to the Great Salt Lake Desert. More than half of the cells having the highest leakances (455 out of 833) are in the central third of the modeled area (rows 21 to 40). In contrast, only 17 percent of the cells having the highest leakances (140 out of 833) are in the southern third of the modeled area (rows 41 to 61). In the central part, about half of the highest leakances correspond to mountain ranges, whereas in the southern third, 60 percent correspond to mountain ranges.

The magnitudes of the computed transmissivities and vertical leakances are dependent on the quantity of assigned recharge. Increasing recharge results in a corresponding increase in discharge and requires a proportional increase in transmissivities and vertical leakances to maintain the same head gradients. The estimates of recharge are only approximations; thus, recharge was increased by a factor of 2 and decreased by a factor of 2 during model calibration to evaluate its effect on transmissivities and vertical leakances.

Conductances used for the head-dependent flow boundaries range from 0.005 to 0.5 ft^2/s and average 0.13 ft^2/s for the 94 cells. Only one cell has a value of 0.005, and three have a value of 0.5. Conductances are slightly different between the different areas. Conductances for the Humboldt River range from 0.1 to 0.5 ft^2/s and average 0.24 ft^2/s; more than half of the cells (11 of 20) have a value of 0.3 ft^2/s. Conductances for the Great Salt Lake and Utah Lake are 0.1 ft^2/s, except for four cells along the Great Salt Lake, which

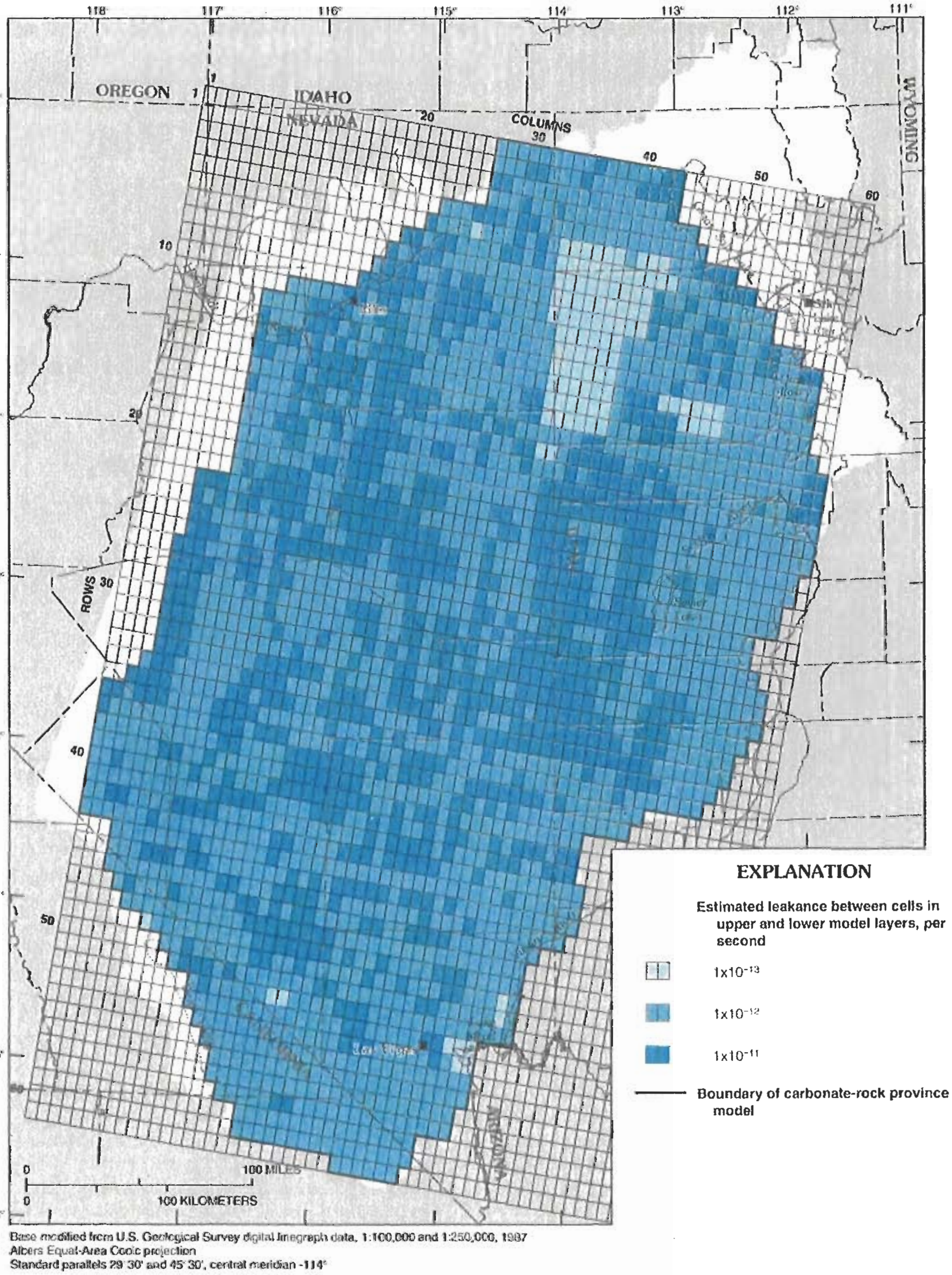


FIGURE 16.—Estimated vertical leakance between cells in upper and lower model layers.

Excerpt from Rush and Kazmi (1965), page 24 and figure 5

Subsurface outflow.--Subsurface, or ground-water, outflow occurs from the southeastern part of Spring Valley principally through the carbonate rocks of the Snake Range to Hamlin Valley. (See discussions of occurrence, movement, and recharge.) Based on an average water-level gradient in the alluvium east of well 8/68-14a1 of about 20 feet per mile (fig. 5), an approximate flow width of 4 miles, and an assumed coefficient of transmissibility of the alluvium of 50,000 gpd per foot, the estimated outflow is roughly 4,000 acre-feet per year. This quantity agrees reasonably well with the estimated recharge of 3,500 acre-feet per year for the area south of the ground-water divide in Spring Valley (pl. 1 and fig. 5).

Eastward movement of ground water from other parts of Spring Valley has not been identified, although carbonate rocks, which are moderately permeable, occur throughout most of the Snake Range.

Discharge from wells.--A few wells are pumped in Spring Valley but only a small amount of the available ground water is utilized. Though stock and domestic wells are numerous, their combined discharge is small, probably not exceeding 200 acre-feet per year. About 10 irrigation wells are used in the valley; their use is limited to years when streamflow is insufficient to satisfy the needs for irrigation. In 1963 and 1964 the wells generally were not used because of adequate snowmelt feeding the creeks. At the time the field work for this report was being done, in July and August 1964, only one irrigation well (13/67-31a1) was being pumped to irrigate about 130 acres of grain. The pumpage estimate for the season was 300 acre-feet. The irrigation of this acreage is entirely dependent on the well because no surface-water supply is available. In 1963, well 12/67-12d3 at the Kirkeby Ranch reportedly pumped about 180 acre-feet of water. The two irrigation wells on the Robison Ranch (T. 18 N., R. 66 E.) have not been used since 1962. No pumpage data are available for irrigation wells in the valley prior to 1962.

Flowing wells discharge an estimated 700 acre-feet of ground water per year. Some of this discharge is used for domestic and stock-watering purposes; however, most of it supports meadow grass and rabbitbrush or percolates back to the water table. The discharge of these wells, like that of the springs, is included in the estimated average annual discharge by the phreatophytes and bare soil.

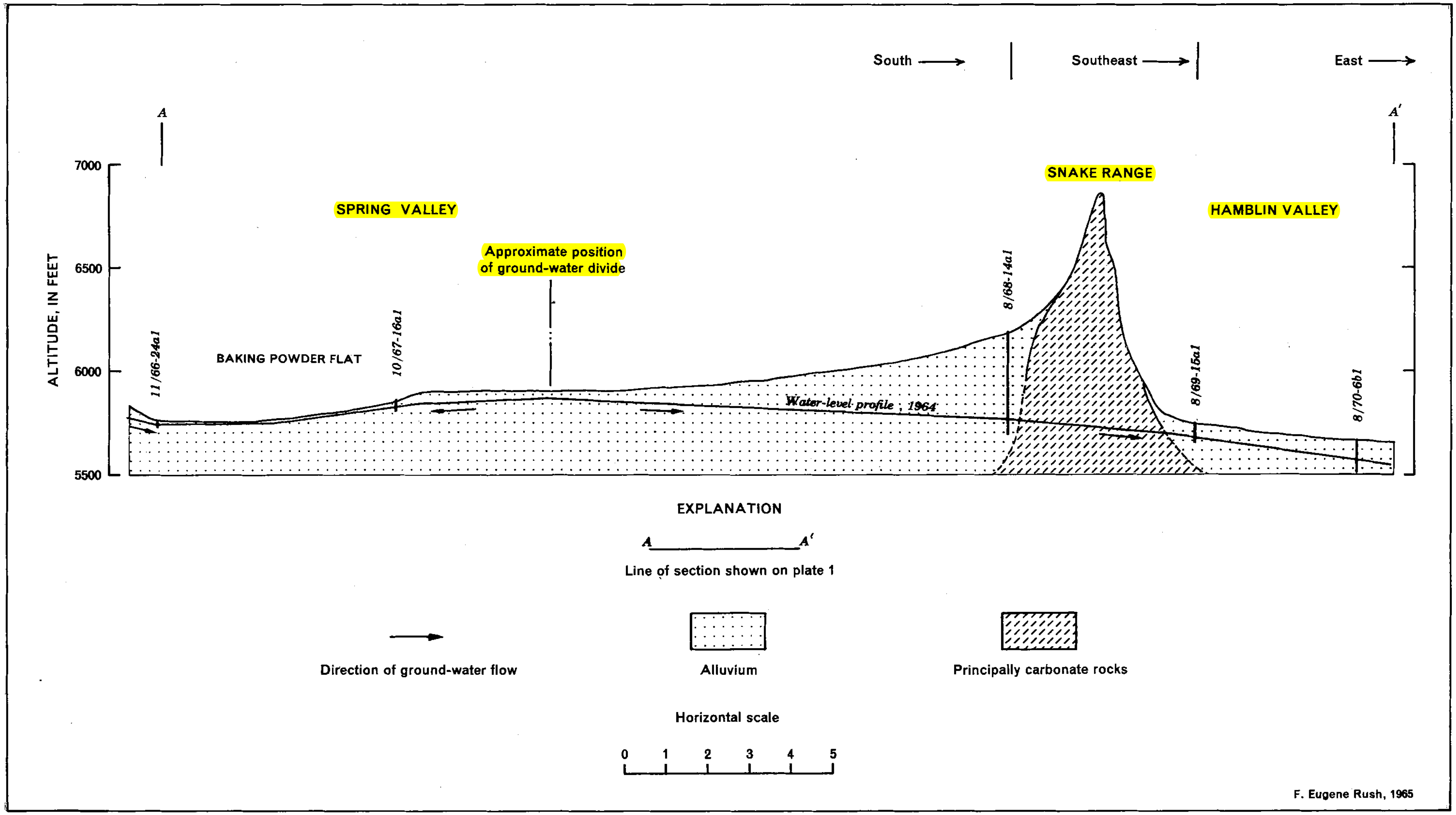


Figure 5.— Cross section of southeastern Spring Valley showing the general topography, water table, and direction of ground-water flow

Excerpt from Elliott, et al (2006), page 44 and plate 1

Large-scale ground-water withdrawals in the valleys likely would affect the discharge of the springs on the southeast and west sides of the southern Snake Range, and streamflow along Big Springs Creek and Lake Creek. These areas, although not studied in detail, probably represent areas that drain ground water as described by Theis (1940). Thus, the spring-discharge areas, Big Springs Creek, Lake Creek, and Pruess Lake were included as areas where surface-water resources likely are susceptible to ground-water withdrawals in Snake and Spring Valleys (pl. 1). Big Springs and the numerous springs at the base of the alluvial slopes on the west side of the southern Snake Range could be affected by ground-water withdrawals similar to springs in Pahrump and Las Vegas Valleys. Large-scale ground-water withdrawals from aquifers in the valleys lowered hydraulic heads and caused springs to stop flowing (Malmberg, 1965, p. 59; Harrill, 1976, p. 43; Harrill, 1986, p. 22).

Summary

Discharge data were continually collected at eight streams and one spring during 2003 and 2004 to quantify discharge and assess the spatial and temporal variability of flow of streams and springs within the Great Basin National Park area. Streamflow gages were installed near the park boundary on Strawberry Creek, Shingle Creek, Lehman Creek, Baker Creek, Snake Creek, South Fork Big Wash, Williams Canyon, Decathon Canyon, and Rowland Spring. Three additional gages were installed along Snake Creek to help characterize streamflow gains and losses within the upper and lower reaches of the drainage. Three of the sites, Snake Creek at the park boundary, South Fork Big Wash, and Decathon Canyon, were intermittent. All other sites were perennial. Mean annual discharge for the perennial streams ranged from 0.53 ft^3/s at South Fork Big Wash to 9.08 ft^3/s at Baker Creek. Seasonal variability of streamflow was climate driven and generally uniform as the minimum and maximum mean monthly discharges occurred in February and June, respectively, at all perennial sites except Strawberry Creek. Decathon Canyon had the lowest annual stream discharge as flow only occurred on 1 day in each of the 2 years of data collection. Maximum mean monthly discharge at Snake Creek at the park boundary and South Fork Big Wash also occurred in June during spring runoff.

Synoptic-discharge and water-property measurements were collected during the spring, summer, and autumn of 2003 along selected reaches on Strawberry, Shingle, Lehman, Baker, and Snake Creeks and Big Wash. Profiles of the selected reaches were developed to relate stream characteristics to geology, and show areas where streams are effluent and likely or potentially susceptible to ground-water withdrawals in adjacent valleys. Streams in contact with permeable rocks or sediments, and areas where streams receive either spring discharge or ground-water inflow comprise areas where streams are most susceptible.

The areas where surface-water resources likely are susceptible to ground-water withdrawals in adjacent valleys that were part of this study include (1) the lower half of Strawberry Creek downstream of the fault contact of the intrusive rocks and Tertiary rocks, including the springs and seeps; (2) Shingle Creek downstream of the intrusive rocks and upstream of the pipeline; (3) Lehman Creek from the lower Lehman Creek campground to the terminus of the stream in Snake Valley, including Rowland Spring and Cave Springs; (4) Baker Creek upstream of the confluence with Pole Canyon tributary to the terminus of the stream in Snake Valley; (5) Snake Creek from just upstream of the park boundary to the terminus of the stream, including Spring Creek.

Areas within the park where surface-water resources potentially are susceptible to ground-water withdrawals include that part of Snake Creek that crosses over the younger undifferentiated rocks (D€r) and its tributaries on the upper plate of the SSRD, and the upper part of Snake Creek that crosses over undifferentiated sedimentary rocks (€r) on the lower plate of the SSRD.

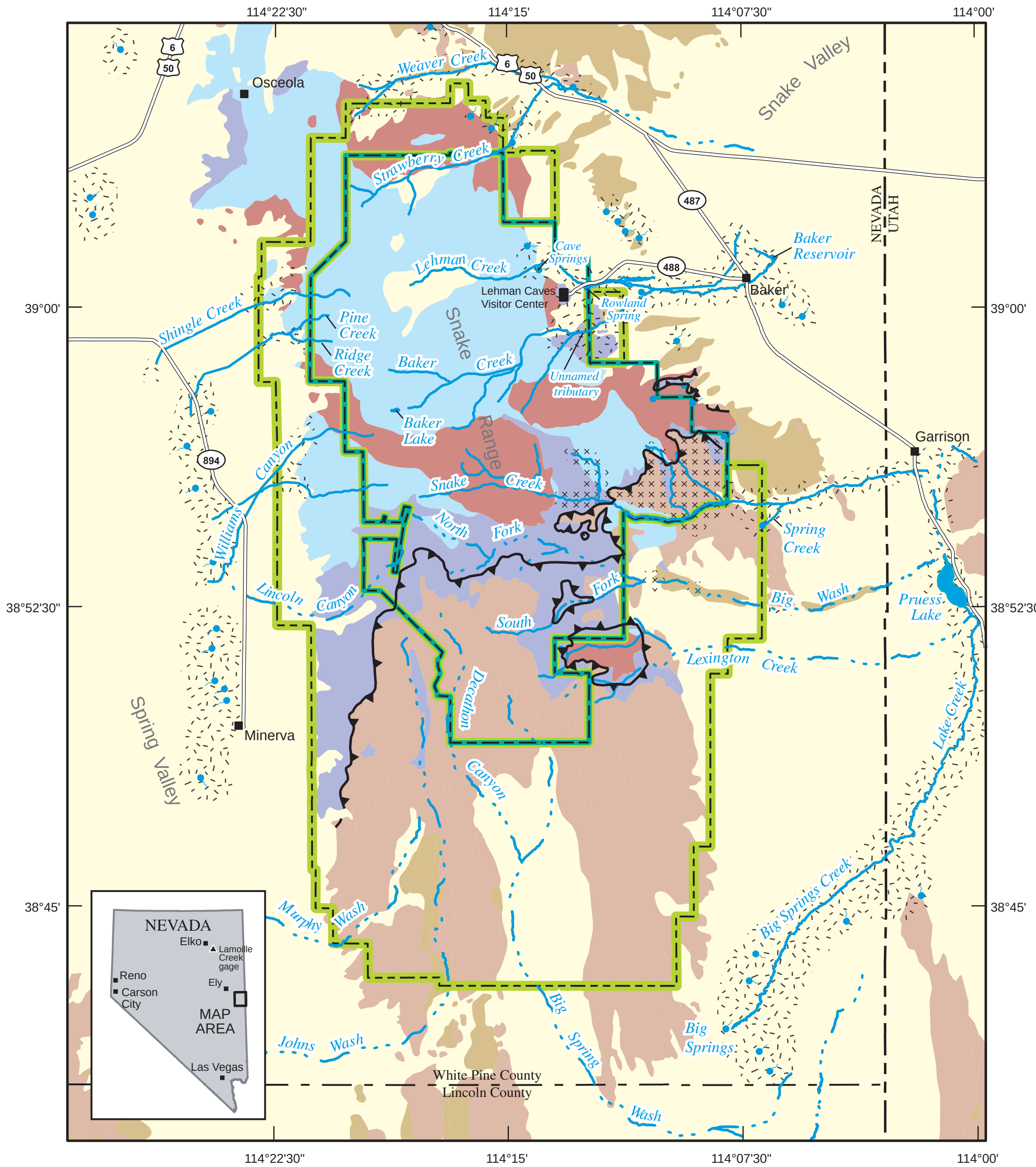
Surface-water resources in other areas adjacent to the park that likely are susceptible to ground-water withdrawals in Spring and Snake Valleys are (1) Williams Canyon upstream of the pipeline, and the following areas that were not gaged, (2) Weaver Creek along the alluvial slope on the northeast end of the southern Snake Range, (3) Pine and Ridge Creeks on the west side of the southern Snake Range between the mountain front and where streams are diverted into pipelines, (4) the numerous springs at the change in slope between the valley floor of Spring Valley and the alluvial slope on the west side of the southern Snake Range, and (5) Big Springs, Big Springs Creek, Lake Creek, Big Wash near Hidden Canyon Ranch, and Pruess Lake in southern Snake Valley.

Areas within the park where surface-water resources probably are not susceptible to ground-water withdrawals in adjacent Spring and Snake Valleys include Big Wash, Lexington Creek, Decathon Canyon, Big Spring Wash, and Lincoln Canyon. Johns Wash and Murphy Wash, adjacent to the park, also would not be susceptible to ground-water withdrawals.

U.S. Department of the Interior
U.S. Geological Survey

Prepared in cooperation with the
NATIONAL PARK SERVICE

SCIENTIFIC INVESTIGATIONS REPORT 2006–5099
Plate 1
Elliott, P.E., Beck, D.A., and Prudic, D.E., 2006
Characterization of Surface-Water Resources in the Great Basin
National Park Area and Their Susceptibility to Ground-Water
Withdrawals in Adjacent Valleys, White Pine County, Nevada



Base from U.S. Geological Survey digital data, 1:100,000, 1979–1988. Universal Transverse Mercator projection, North American Datum of 1983, Zone 11

5 0 5 MILES
5 0 5 KILOMETERS

Geology modified from Hose and Blake (1976)

EXPLANATION

Area where surface-water resources likely are susceptible to ground-water withdrawals

Area where surface-water resources potentially are susceptible to ground-water withdrawals

Geology

- QTs Alluvial and glacial deposits
- Tr Tertiary rocks
- TJi Intrusive rocks
- D€r Younger undifferentiated rocks
- €r Undifferentiated sedimentary rocks
- €Zr Older undifferentiated rocks

Southern Snake Range décollement

Humboldt National Forest boundary

Great Basin National Park boundary

Spring

GENERALIZED AREAS WHERE SURFACE-WATER RESOURCES LIKELY OR POTENTIALLY ARE SUSCEPTIBLE TO GROUND-WATER WITHDRAWALS IN ADJACENT VALLEYS, GREAT BASIN NATIONAL PARK AREA, NEVADA

By

Peggy E. Elliott, David A. Beck, and David E. Prudic