SIMULATION OF GROUND-WATER FLOW AT ANCHORAGE, ALASKA, 1955-83

By Leslie D. Patrick, Timothy P. Brabets, and Roy L. Glass

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CONVERSION TABLE

For readers who may prefer to use metric (International System) units rather than inch-pound units, the conversion factors for the terms used in this report are listed below:

Multiply inch-pound unit	**by**	**To obtain metric unit**
inch (in.)	25.4	millimeters (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
foot squared (ft^2)	0.09294	meter squared (m^2)
square mile (mi^2)	2.590	square kilometer (km^2)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
foot per day (ft/d)	0.3048	meter per day (m/d)
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

Sea level:

In this report "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929) -- a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called "Sea Level Datum of 1929."

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ABSTRACT

The ground-water system at Anchorage, Alaska was analyzed by using a two-layer three-dimensional mathematical model. By use of existing data, both nonpumping and pumping steady-state conditions and transient conditions were simulated. Under steady-state conditions, calculated directions of ground-water flow were similar to observed flow patterns, and calculated stream discharges were generally within 10 percent of observed values. However, in many parts of the modeled area computed head values were more than 20 feet higher or lower than observed values. Hydraulic conductivity and transmissivity are the most sensitive hydraulic parameters under steady-state conditions. A pumping rate of 18.8 million gallons per day for steady-state conditions lowers heads in the confined aquifer by as much as 30 feet but reduces streamflow by less than 5 percent. Transient simulations show that drawdowns due to production wells follow similar patterns of nearby observation wells. On the basis of analytical techniques, the confining layer does not appear to contribute significant quantities of water.

INTRODUCTION

The Municipality of Anchorage encompasses approximately 1,200 mi^2 in southcentral Alaska. The area commonly referred to as the Anchorage Bowl encompasses approximately 180 mi^2 and includes most of the urban area of Anchorage (fig. 1). This area is bounded on the north, west, and south by two estuaries, Knik and Turnagain Arms of Cook Inlet, and on the east by the Chugach Mountains.

In 1971 ground water supplied by public utilities totaled 3.7 billion gallons. More than 4.7 billion gallons were supplied in 1981. (Ground-water data in this report were compiled from records supplied by Anchorage Water and Wastewater Utility, Central Alaska Utilities, Elmendorf Air Force Base, and the U.S. Army at Fort Richardson.) Additional ground water is used by homeowners in the southern third of the Anchorage Bowl where public water and sewer facilities are not available.

Increased pumpage in the Anchorage Bowl has led to a gradual decline in ground-water levels (fig. 2); however, levels recovered during 1980-82 owing to above average precipitation. Many homeowners served by individual wells have filed for water rights that could lead to conflicts with water-rights applications by the Anchorage public water utility. There is concern that water from on-site septic systems could contaminate the ground-water supply and that the amount of ground water used could decrease the flows in Anchorage-area streams. Thus, in 1983, the U.S. Geological Survey, in cooperation with the Municipality of Anchorage, began a study of the ground-water system of the Anchorage Bowl.

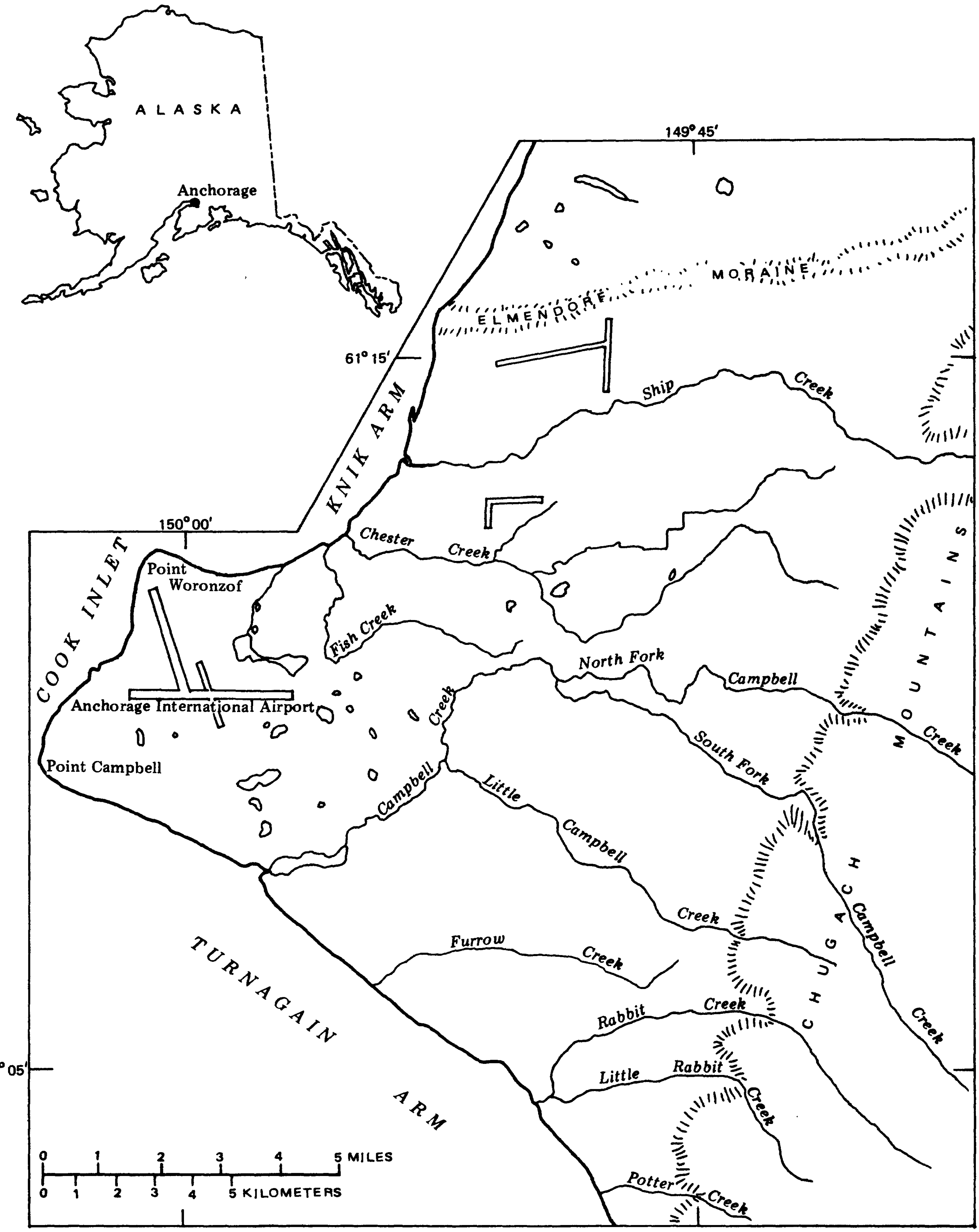


Figure 1.--Physical features of the Anchorage area.

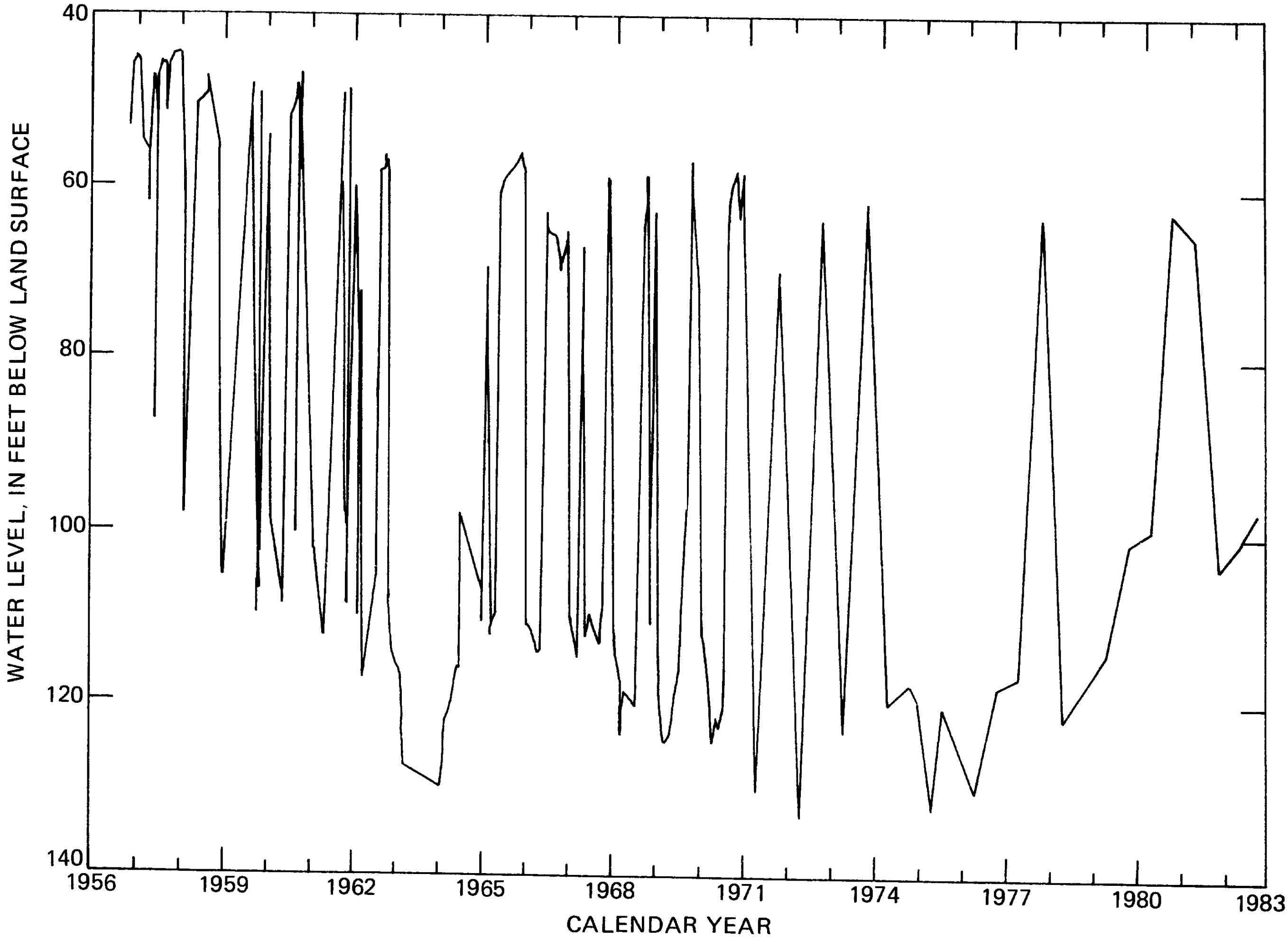


Figure 2.--Long-term water levels from observation well (SB013003009 CDCD2) in the Anchorage Bowl, 1957-83. (See figure 4 for well location.)

This report describes the results of a study to develop a better understanding of the ground-water system of the Anchorage Bowl and how ground-water levels and streamflow may be influenced by present and future pumping rates. The approach was to use data collected from 1955 to 1983 and to use improved techniques for analyzing these data.

Previous studies have investigated the ground-water system of the Anchorage Bowl. From 1967 to 1976, an electric analog model was used to simulate movement of water through the main confined aquifer system beneath the Anchorage Bowl. From 1977 to 1982, a digital model (Trescott and others, 1976) was applied to the Anchorage Bowl (G.W. Freethey, U.S. Geological Survey, written commun., 1982) to improve understanding of the interaction between the unconfined and confined aquifers. However, that model assumed flow to be horizontal (two dimensional), which is not accurate in recharge and discharge areas where a vertical component of flow is present. For the current study, a three-dimensional mathematical model was used to represent the ground-water system more accurately than could the previous models.

This report includes a description of the sources and types of data used in developing the model and a discussion of model calibration, model results, and possible sources of inaccuracies in these results. Detailed analysis of specific sites or local areas were beyond the scope of this study. However, the principles used to construct the model can be used to develop more detailed models of smaller areas where sufficient data are available.

HYDROGEOLOGY

Ground water in the Anchorage Bowl flows in sediments that extend from the foot of the Chugach Mountains westward to and below Cook Inlet (fig. 3). These sediments thicken progressively away from the mountains and are as much as 1,300 ft thick beneath Point Campbell. The sediments overlie metamorphic bedrock near the mountains and low-permeability sedimentary bedrock of the Tertiary Kenai Formation throughout the rest of the Bowl.

Well logs indicate that the sediments form a geohydrologic system composed of an unconfined aquifer (fig. 4), a confining layer (fig. 5) that is nearly continuous except near the mountains, and a series of hydrologically connected units that are collectively termed the confined aquifer (fig. 6). The confined and unconfined aquifers consist of layers and lenses of interbedded sand and gravel, till, silty sand, and silty clay deposits. In the western part of the study area, except for possible breaks near Point Woronzof, the confining layer is a continuous layer of clay and silt -- the Pleistocene Bootlegger Cove Formation (Ulery and Updike, 1983), commonly referred to as "Bootlegger Cove clay." In the eastern part of the modeled area, the confining layer consists predominantly of till or till-like deposits. Continuity of the confining layer in this part of the study area is not well documented. The confining layer ranges from 0 to 270 ft thick and generally thickens with increasing distance from the mountain front.

Water enters (recharges) the ground-water system of the Anchorage Bowl in different ways. Along the mountain front, ground water seeps from

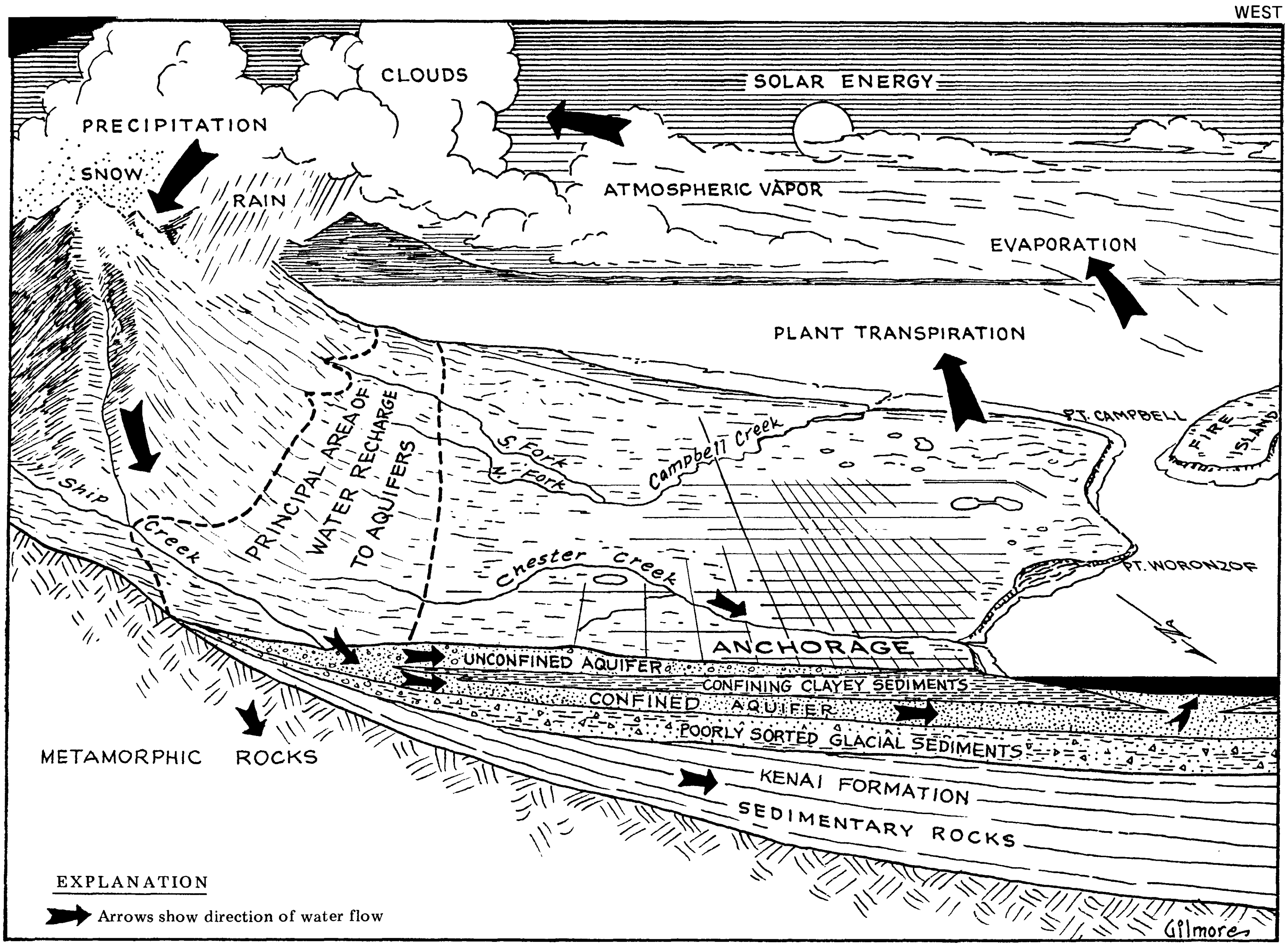
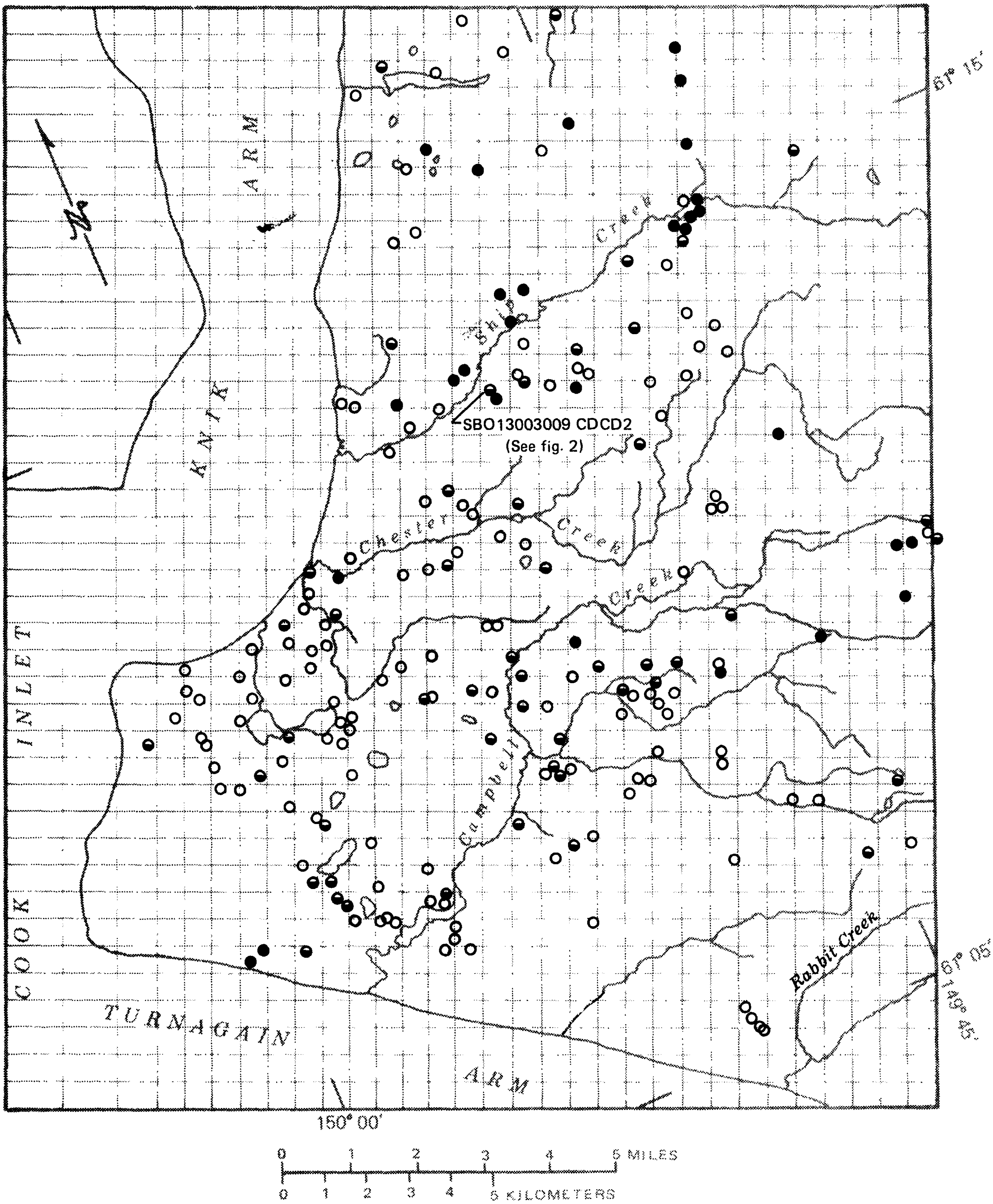


Figure 3. -- Generalized ground-water flow in the Anchorage Bowl (Barnwell and others, 1972).

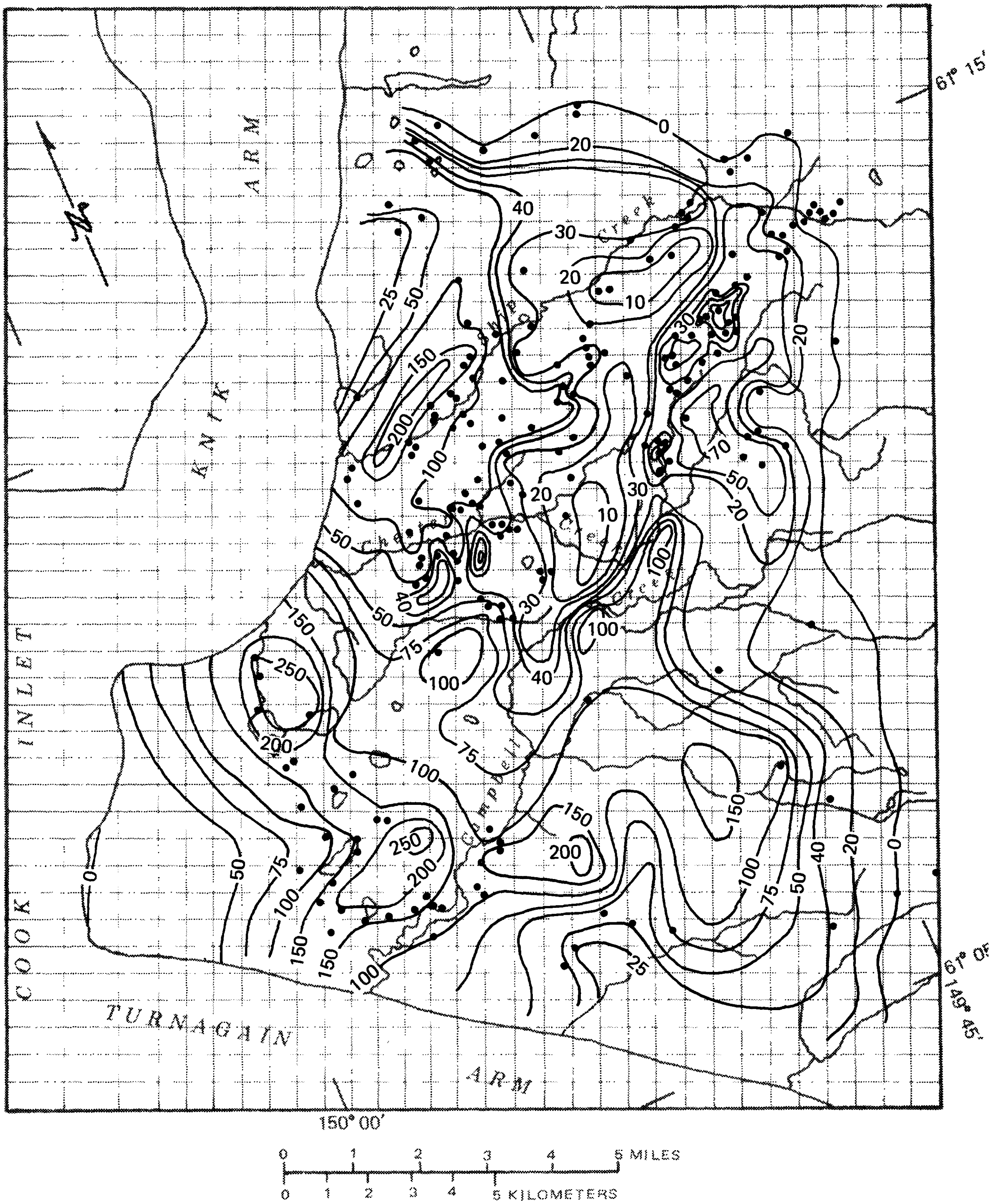


EXPLANATION

- 0-25 ft.
- 26-50 ft.
- >50 ft.

Estimated thickness of the unconfined layer

Figure 4.--Estimated thickness of the unconfined aquifer.

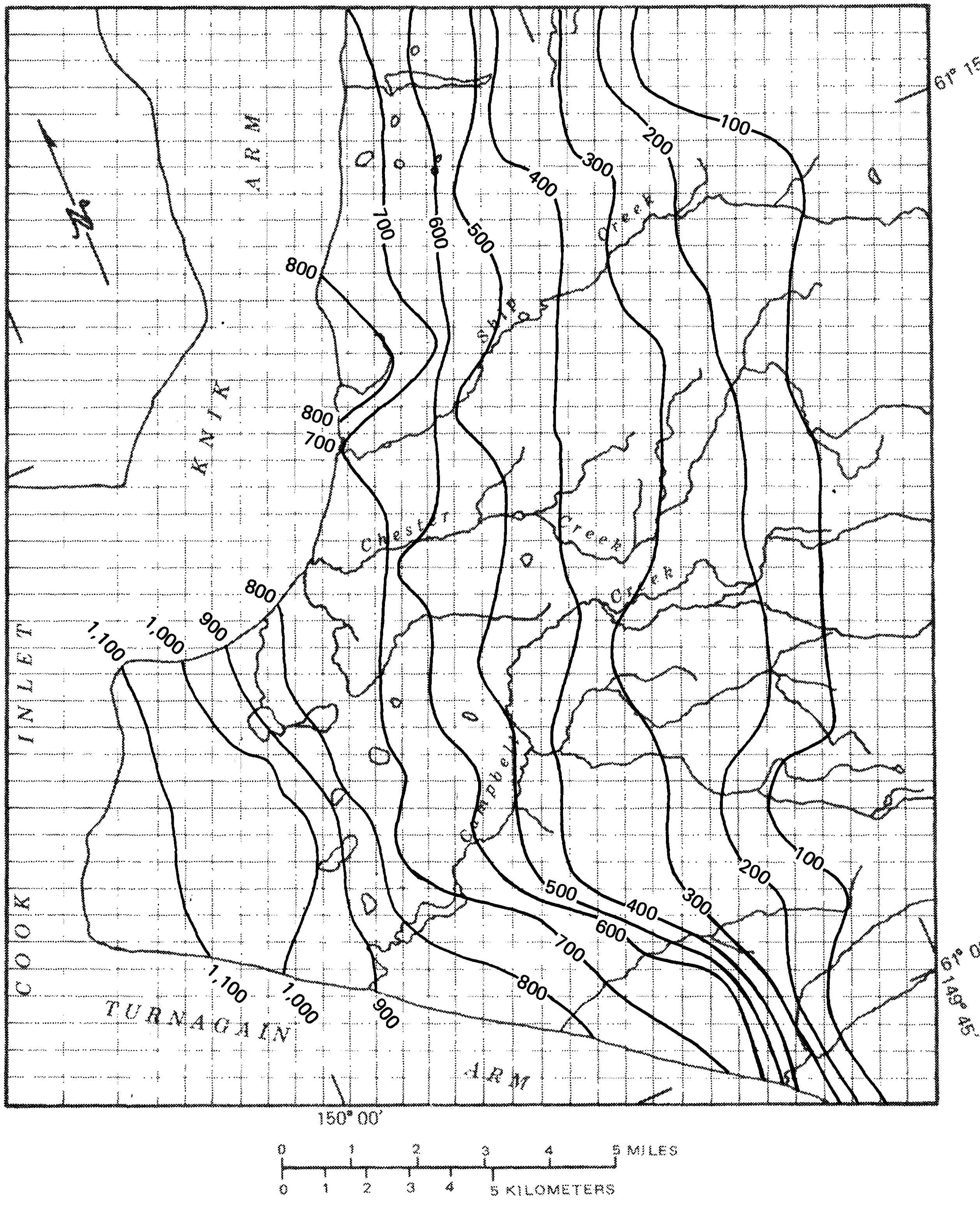


EXPLANATION

—100— Line of equal thickness of confining layer, in feet
Contour interval variable

• Data control point (well)

Figure 5.--Estimated thickness of confining layer.



EXPLANATION

Line of equal thickness of the confined aquifer. Interval 100 feet.

Figure 6.--Estimated thickness of the confined aquifer.

bedrock fractures into the sediments. Snowmelt and rainfall also infiltrate to the aquifer. In the lowlands, water infiltrates into the ground along reaches of streams that flow over areas underlain by permeable sand and gravel where the water table is below the altitude of the streambed. Water discharges from the ground-water system either through the streams and their tributaries where the water table intersects the streambed, or into Cook Inlet.

SIMULATION OF GROUND-WATER FLOW

Model Specifications

The primary function of a mathematical ground-water model is to simulate directions and rates of water movement and distribution of hydraulic head throughout an aquifer system. The model solves the partial-differential equations of ground-water flow and requires that the hydraulic properties and boundaries be defined for the modeled area (McDonald and Harbaugh, 1984).

The aquifer system at Anchorage was overlain by a grid, which was extended in the third dimension to form blocks or "cells" (fig. 7). The cells form 41 rows, 31 columns, and 2 layers. Each cell in the model grid represents a block of earth material within the aquifer system and contains one or more types of materials within which streams and hydraulic properties of aquifers are assumed to be uniform. Any specific cell may be referenced by citing its row, column, and layer location. Limits of the modeled area were selected to include or nearly coincide with natural flow boundaries.

The upper cells (layer 1) represent an unconfined aquifer whereas the lower cells (layer 2) represent a confined aquifer. The intervening confining layer was simulated not as an active layer, but as an impedance to flow (low vertical conductance) between the two active layers. Hydraulic properties of the geologic materials which make up this "inactive" layer were used to estimate its vertical conductance. Heads in the confining layer were not calculated during each run of the model.

Boundary and Initial Conditions

The conceptual model of the Anchorage Bowl consists of a region containing ground water that is surrounded by a closed surface. This surface is generally referred to as the "boundary surface" (Franke and others, 1987) of the flow region and corresponds to identifiable hydrogeologic features at which some characteristic of ground-water flow can be described. For the conceptual model of the Anchorage Bowl, these features were Knik and Turnagain Arms, the Chugach Mountains and the Elmendorf Moraine. After the boundary surface was chosen, boundary types (figs. 8 and 9) were assigned to each node on the boundary surface.

A no-flow boundary, across which flow is assumed to be zero, is used to represent nearly impermeable rocks. Although natural earth materials are never completely impermeable, they are sometimes regarded as effectively impermeable for modeling purposes if the hydraulic conductance of the adjacent materials differs by several orders of magnitude.

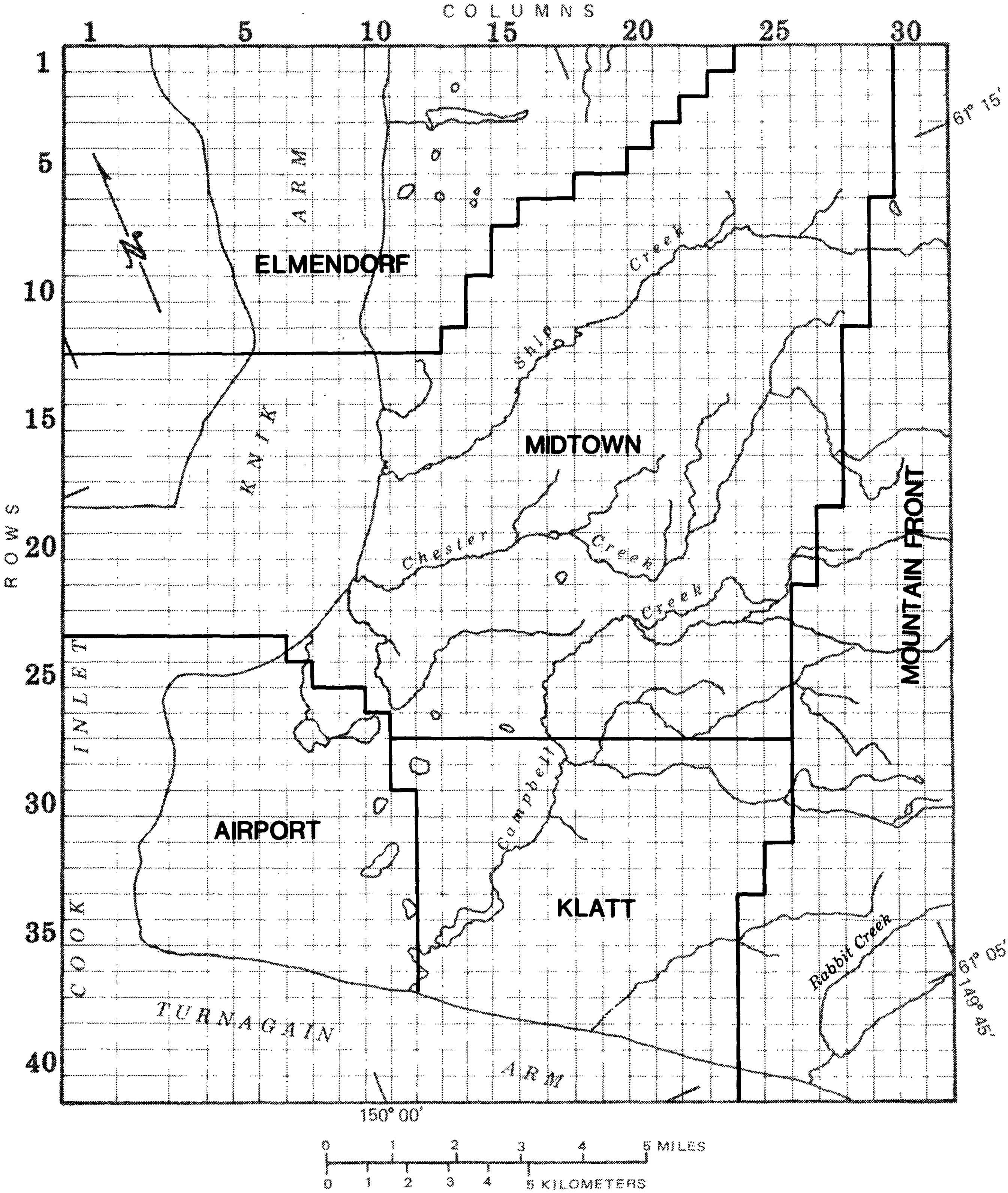


Figure 7.--Grid used to model the ground-water system and the five sub-areas, based on generalized geology.

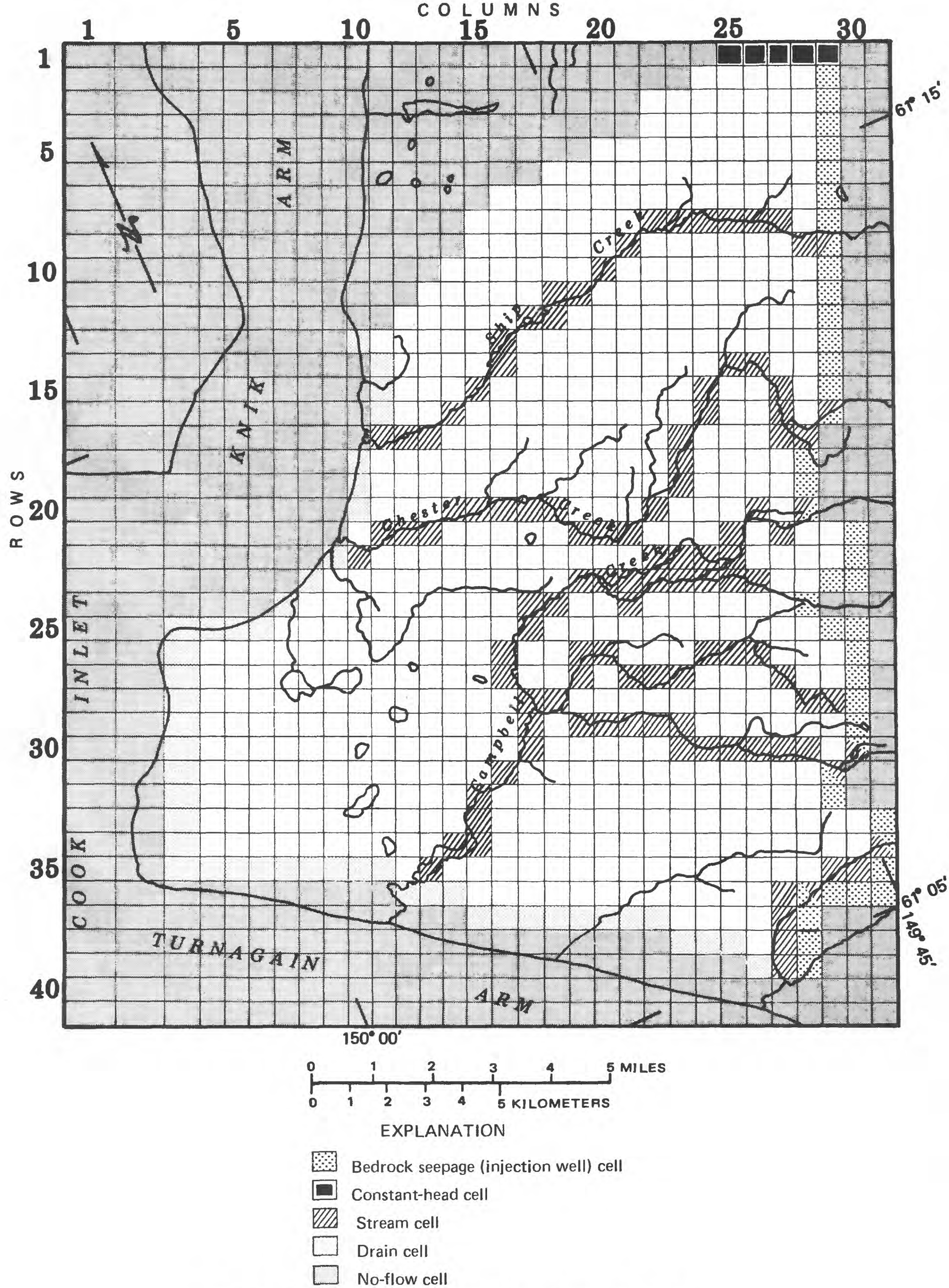


Figure 8.--Model grid for the unconfined aquifer (layer 1).

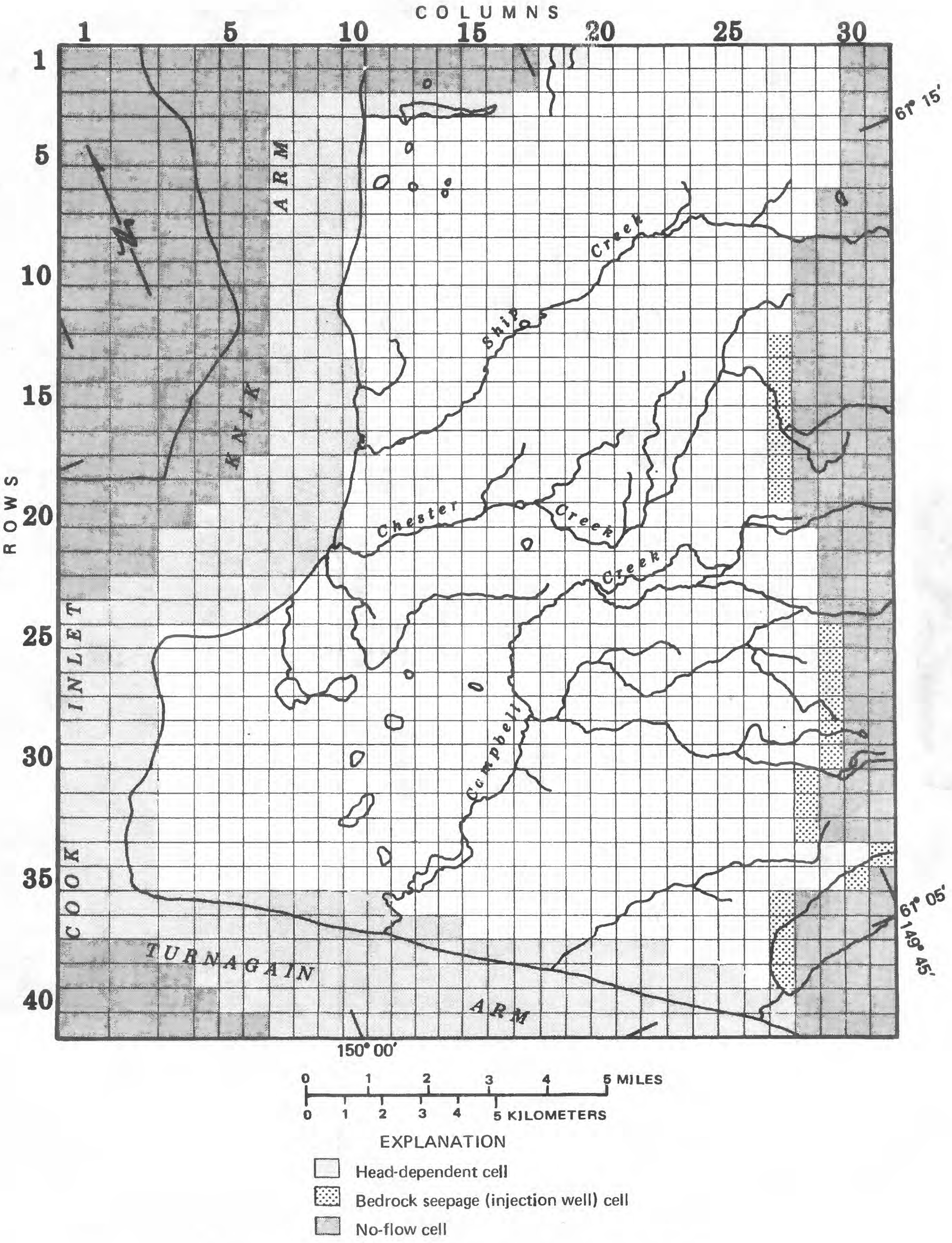


Figure 9.--Model grid for the confined aquifer (layer 2).

Cells in layer 1 that were set as "no-flow boundaries" consisted of nodes outside the modeled area and cells at the Elmendorf Moraine. The Elmendorf Moraine was assumed to be a no-flow boundary because of its relatively low transmissivity. For layer 2, cells were set as no-flow along the mountain front where unconsolidated material is less than 150 ft thick.

A constant-head boundary represents a part of an aquifer system in which head does not vary appreciably with time. Five cells (spanning a distance of approximately 1 mi) in layer 1 at the north edge of the modeled area were assumed to be constant-head cells. This area consists of outwash gravels adjacent to the Elmendorf Moraine. Heads on this part of the moraine were assumed to be nearly constant because of the presence of several small lakes and ponds, a small perennial stream, and wetlands.

Flow into or out of some cells is dependent upon changes in head in the cell itself or in changes in head in adjacent cells. These cells are referred to as head-dependent cells. No cells in layer 1 were considered to be head-dependent cells. In layer 2, cells at the north end of the modeled area were considered head-dependent since layer 2 extends beyond the upper layer. Also, the cells in layer 2 in Knik and Turnagain Arm were considered head-dependent. The elevations of these heads were corrected from salt-water heads to fresh-water heads.

Stream cells allow water to flow between an aquifer and a stream, at a rate based on the relative elevations of their water levels. Flow into and out of layer 1 from significant streams was simulated by identifying those cells that contained streams and assigning each cell a conductance value. The conductance value, C, equals the vertical hydraulic conductivity, K, divided by the thickness of the streambed, b, and describes the capacity of the streambed to transmit water between the stream and aquifer system. Values of conductance were estimated by visual inspection of bed materials, and in Ship Creek, by measurements of streambed permeability. For our model, the thickness, b, was assumed to equal 1 ft; thus conductance equaled vertical hydraulic conductivity.

Cells containing seepage faces -- another type of head-dependent flow boundary -- along bluffs or in marshes or mudflats (along Turnagain and Knik Arms) were defined as drains. The elevation of each drain was set at the average elevation of the top of the confining layer in that cell. When the elevation of the water level in the cell is above the specified elevation, water flows out of the aquifer at a rate proportional to the elevation of the water above the drain and the conductance of the seepage face. When the aquifer's water level falls below the specified elevation, no water discharges through the drain.

Aquifer Recharge

The three principal means of recharge to the Anchorage Bowl aquifers are bedrock seepage, precipitation, and leakage from streams. Values for these three components are described in the following paragraphs.

The rate at which water enters the unconsolidated material from bedrock is not precisely known. Previous work by U.S. Geological Survey personnel (J.B. Weeks, written commun., 1968) has led to estimates which range from 15 to 49 Mgal/d. For this model, a value of 19 Mgal/d was used to represent

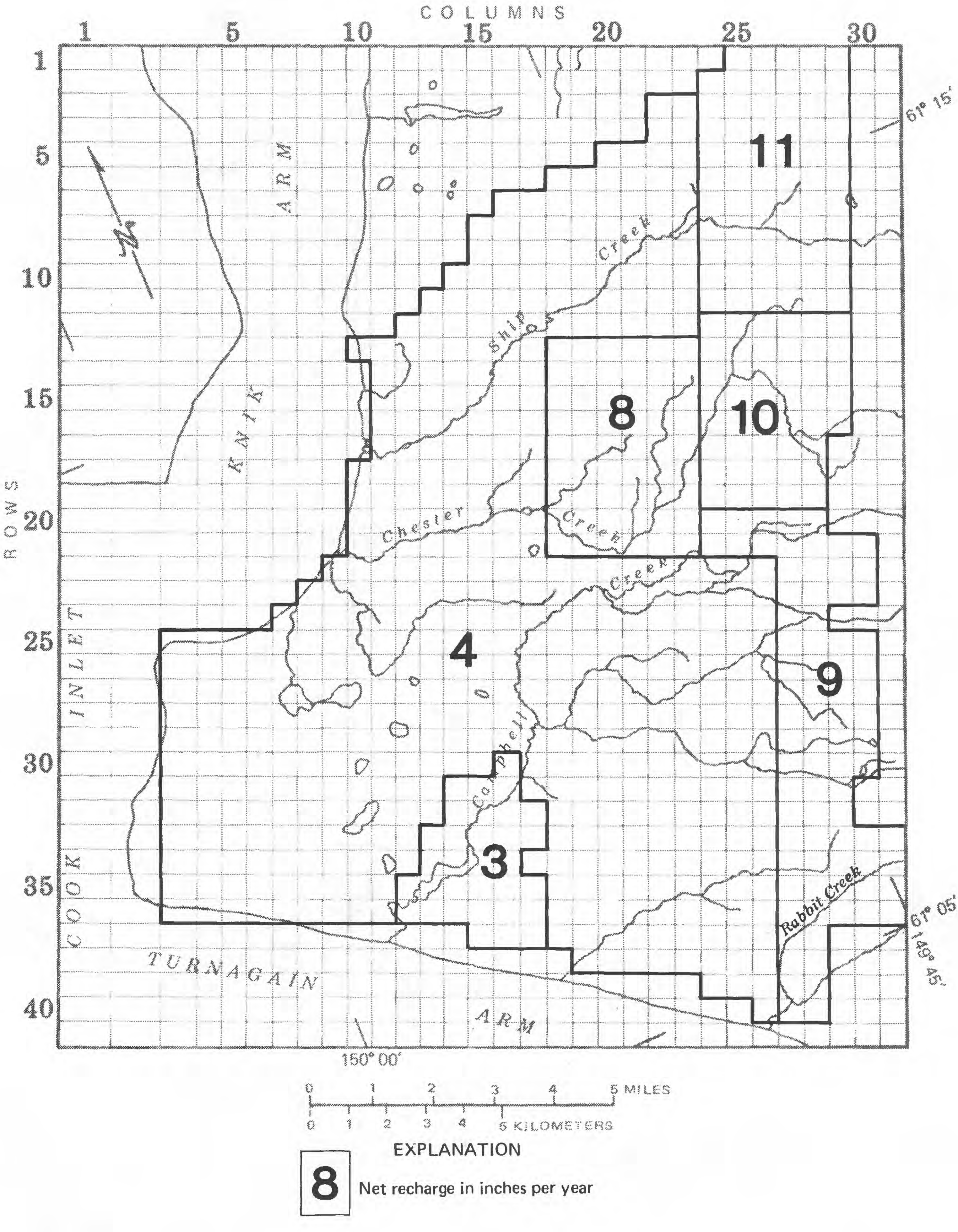


Figure 10.--Net recharge to the unconfined aquifer (layer 1) from precipitation.

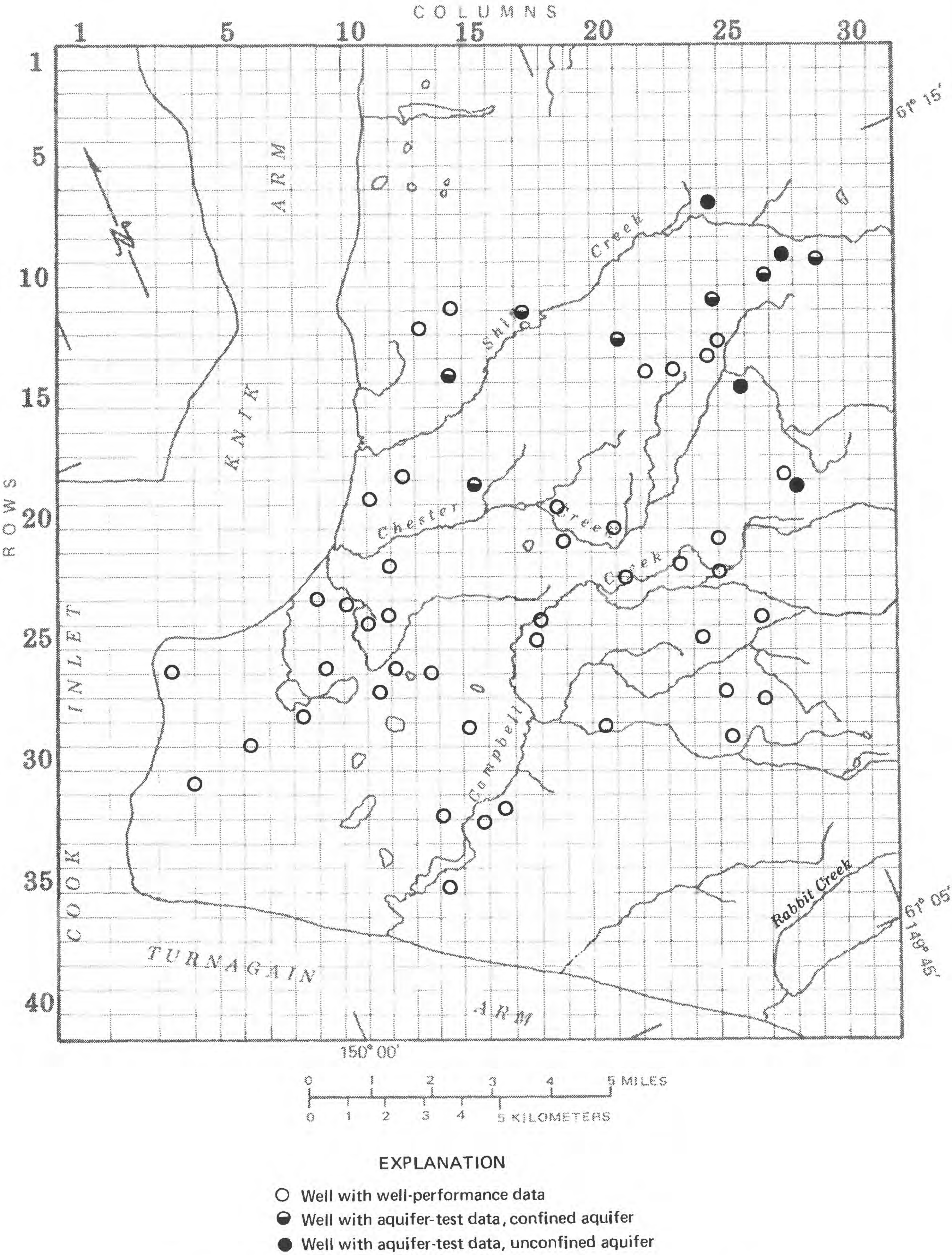


Figure 11.--Locations of wells used to estimate aquifer properties.

the total recharge to layers 1 and 2 as ground-water inflow from metamorphic bedrock. Of this 19 Mgal/d, 15 were apportioned equally among the 42 mountain-front cells in layer 1 to simulate seepage of water from bedrock into unconsolidated materials (fig. 8). The remaining 4 Mgal/d were apportioned among 22 active cells at the eastern boundary of layer 2 (fig. 9). These rates remained constant during all model runs. No additional bedrock seepage was assumed to enter the confined aquifer from underlying sedimentary bedrock.

Annual precipitation averages about 15 in. at Anchorage International Airport (altitude, approximately sea level) but increases with increasing altitude eastward toward the Chugach Mountain front. Although no long-term precipitation data are available for higher altitudes, short-term records were used to apportion precipitation across the various model cells. The quantity of precipitation lost to evapotranspiration is not known, but estimates of evapotranspiration in low-lying areas of Anchorage range from 10 to 20 in. (Zenone, 1976). For this model, net recharge rates were estimated as the difference between precipitation and evapotranspiration and ranged from 3 in/yr in parts of the lowlands to 11 in/yr in the mountains in the northeastern part of the modeled area. Recharge was input at a constant rate to layer 1; no allowance was made for seasonal or yearly variation (fig. 10).

Four major streams flow through the Anchorage Bowl: Ship, Chester, Campbell, and Rabbit Creeks (fig. 1). Data from gaging stations (see fig. 16 later in report) on these streams indicate reaches where a stream may be "gaining" or "losing" water and thus allow estimation of the approximate quantities of water that are gained from or lost to the adjacent aquifer. On the basis of this information, 15.4 Mgal/d was estimated to recharge the ground-water system from these four streams. This recharge total was apportioned among those cells which represented the "losing" part of a particular stream. For those nodes that represented the gaining part of a stream, the quantity of flow assigned to the node was based on interpolation of the mean annual flow of the nearest stream-gaging station. The gains and losses thus estimated were reproduced during calibration of the model by varying streambed conductance and other factors that affect head in the aquifer.

Hydraulic Properties

Hydraulic conductivity, transmissivity, and the storage coefficient are hydraulic properties that describe an aquifer's ability to transmit and store ground water. Data from about 50 aquifer tests or well-performance tests (fig. 11) and lithologic data for several hundred wells were used to estimate these properties throughout the modeled area. It should be noted that well-log data are concentrated in some areas, whereas for other areas no data are available. The hydraulic properties for areas of sparse data were estimated from our knowledge of local geohydrologic conditions and extrapolation of data from the nearest well.

Hydraulic conductivity is the property of a geologic material that describes its capacity to transmit water. For the unconfined aquifer, values of horizontal hydraulic conductivity, estimated from lithologic and specific-capacity data (well pumping rate/observed drawdown) and from the well-performance tests, ranged from 3 to 150 ft/d. A map showing relative permeability (a measure of the ease of fluid flow through rocks) of surficial geologic materials (Freethey, 1976) was also used as a guide where no wells were present. Values are shown on figure 12 and were initially assigned as follows:

Relative permeability of surficial deposits (from Freethey, 1976)	Hydraulic conductivity (feet per day)
High to very high	100
Moderate to high	10
Low to moderate	1
Low to very low	0.1

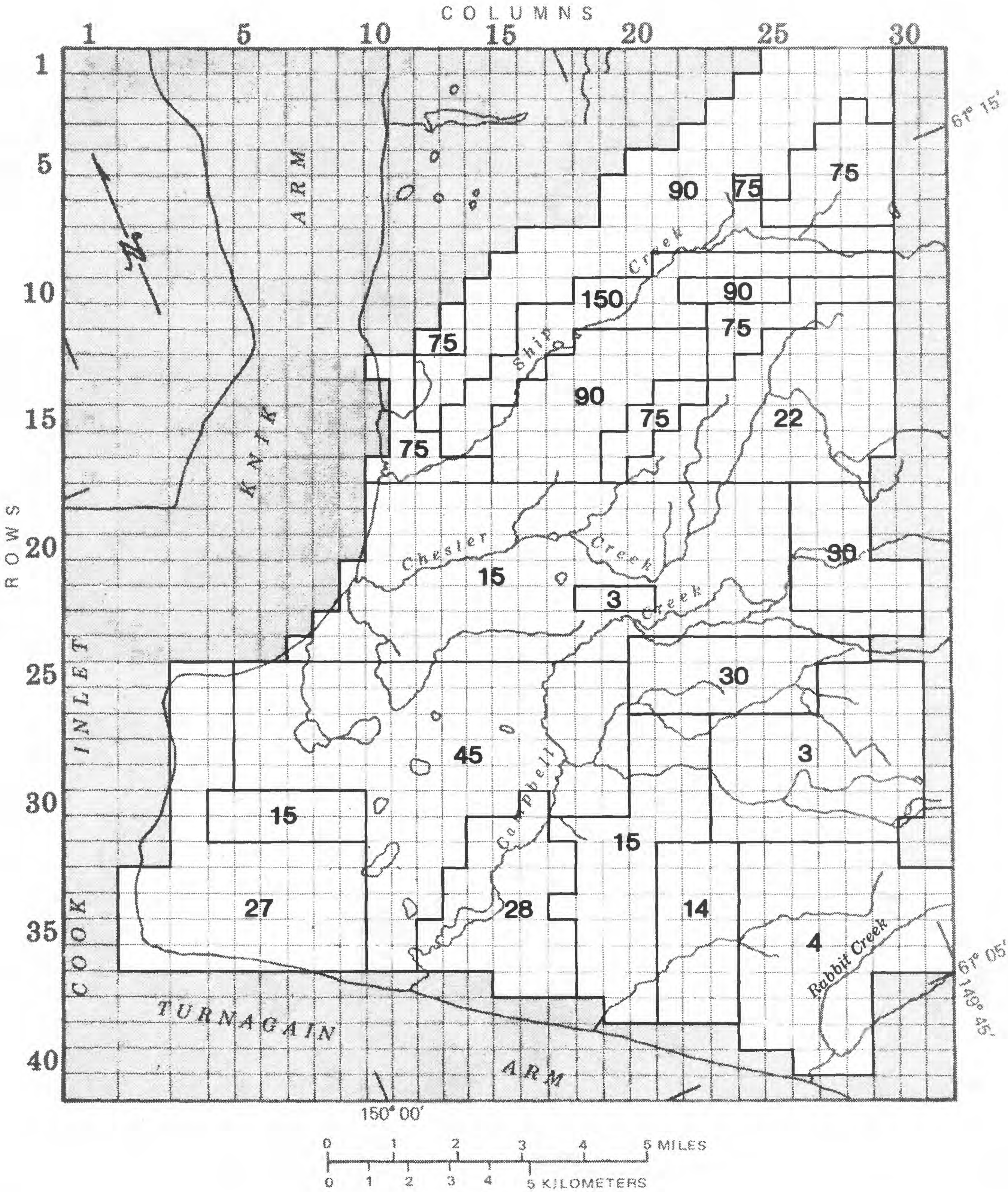
Similarly, values of horizontal hydraulic conductivity for confined cells in layer 2 (fig. 13) were estimated using lithologic and aquifer-test data.

Hydraulic conductivity in the vertical direction also was estimated throughout the modeled area. The average vertical conductivity of the confining layer was estimated from typical values for various types of geologic materials (Morris and Johnson, 1967) to range from 1×10^{-4} ft/d for predominantly clayey areas to 1 ft/d in the mountain-front area where the confining layer is assumed to be either discontinuous or absent (fig. 14). Values of vertical hydraulic conductivity in inactive cells in the northwestern part of the area were set to 1 ft/d.

Transmissivity is an expression of the rate at which water is transmitted through a unit-wide section of an aquifer under a unit hydraulic gradient. It equals the product of an aquifer's hydraulic conductivity multiplied by its saturated thickness. Cederstrom and others (1964, p. 62-63), used aquifer-test data from the Anchorage Bowl to calculate transmissivity values for individual strata in the confined-aquifer system ranging from approximately 5,300 to 13,000 ft^2/d. They estimated the average transmissivity for the entire thickness of the confined aquifer to be approximately 27,000 ft^2/d.

Results of aquifer tests conducted since 1964 and re-evaluation of earlier tests tend to support transmissivity values of less than 10,000 ft^2/d. For this model, transmissivity values for each cell in layer 2 (fig. 15) were calculated from results of aquifer tests or by multiplying each cell's hydraulic conductivity by its thickness. Values of transmissivity ranged from 0 to 26,000 ft^2/d.

The storage coefficient of a saturated aquifer is defined as the volume of water that an aquifer releases from or takes into storage per unit area of the aquifer per unit change in hydraulic head. Values of storage coefficient are much larger for unconfined aquifers than for confined aquifers. This is because releases from storage in unconfined aquifers represent an actual dewatering of pores, whereas releases from storage in confined aquifers result from the expansion of water and compaction of the



EXPLANATION

15	Hydraulic conductivity in feet per day
	Inactive cell

Figure 12.--Estimated hydraulic conductivity of the unconfined aquifer (layer 1).

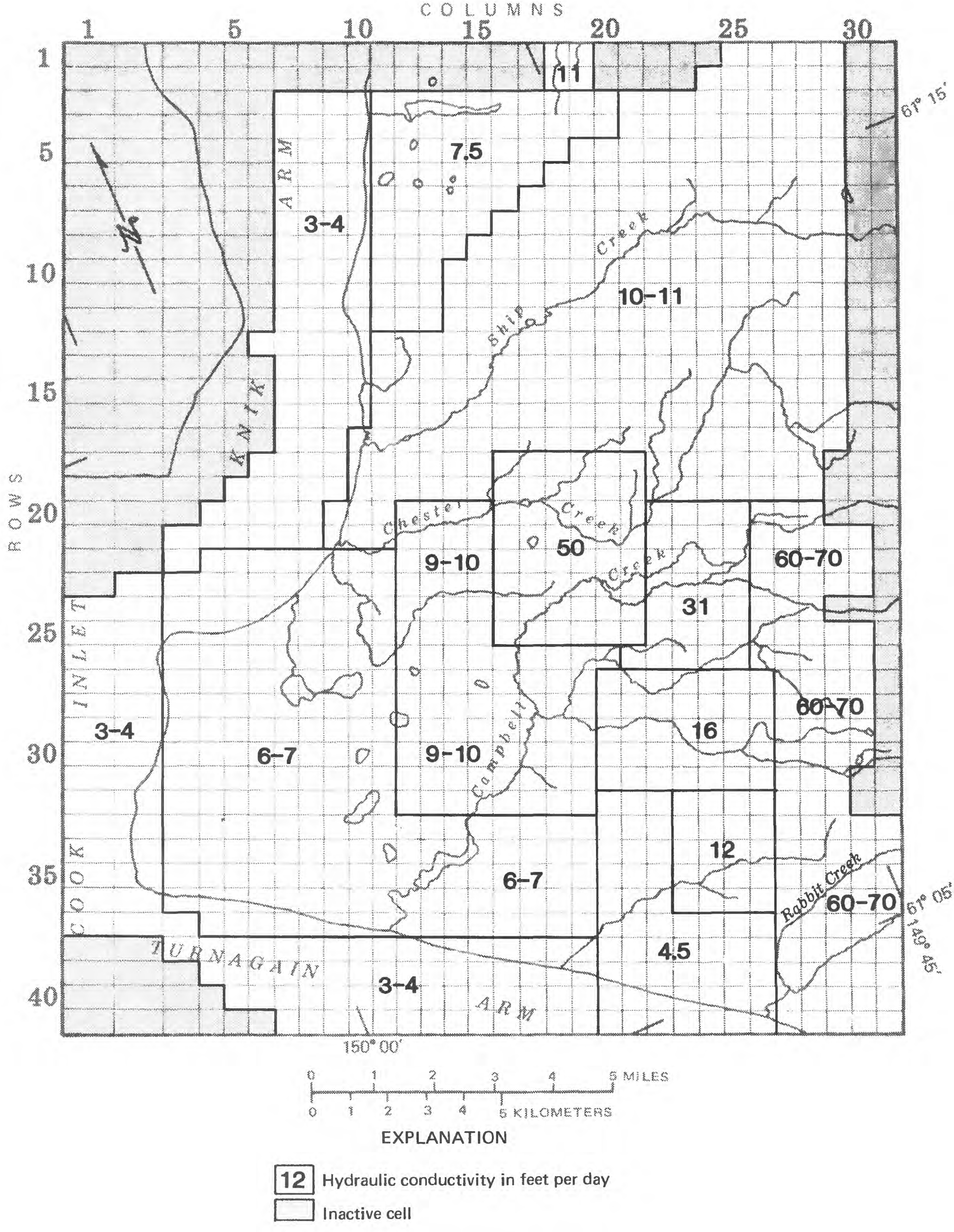
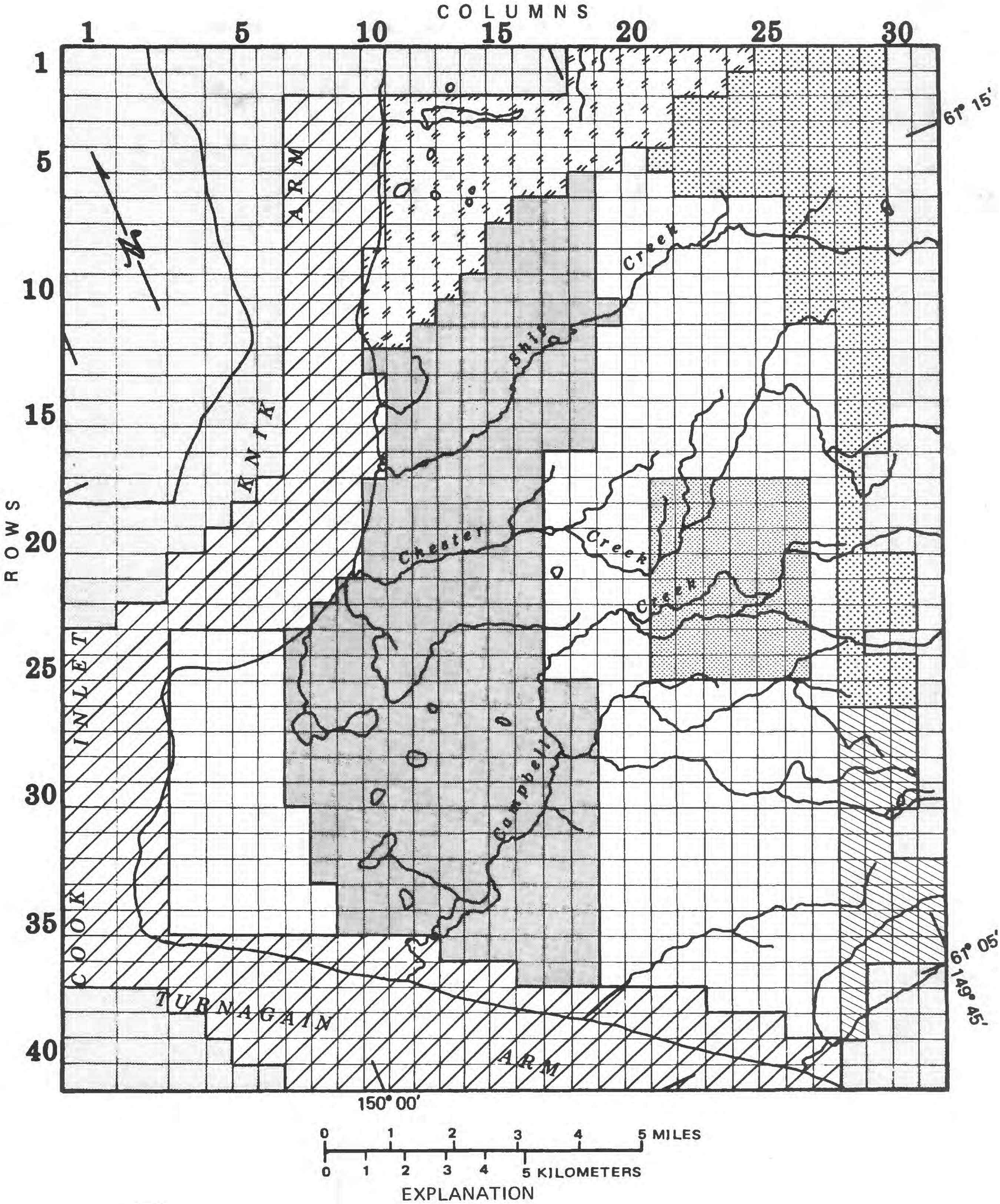


Figure 13.--Estimated hydraulic conductivity of the confined aquifer (layer 2).



EXPLANATION

- Inactive cell
- Layer 1-Inactive
- Area where leakage to layer 2 occurs through Knik and Turnagain Arms
- Region 1-Conductivity = 0.0001
- Region 2-Conductivity = 0.005
- Region 3-Conductivity = 1.0
- Region 4-Conductivity = 0.8
- Region 5-Conductivity = 0.001

Figure 14.--Estimated vertical hydraulic conductivity of the confining layer and hydraulic conductivity regions.

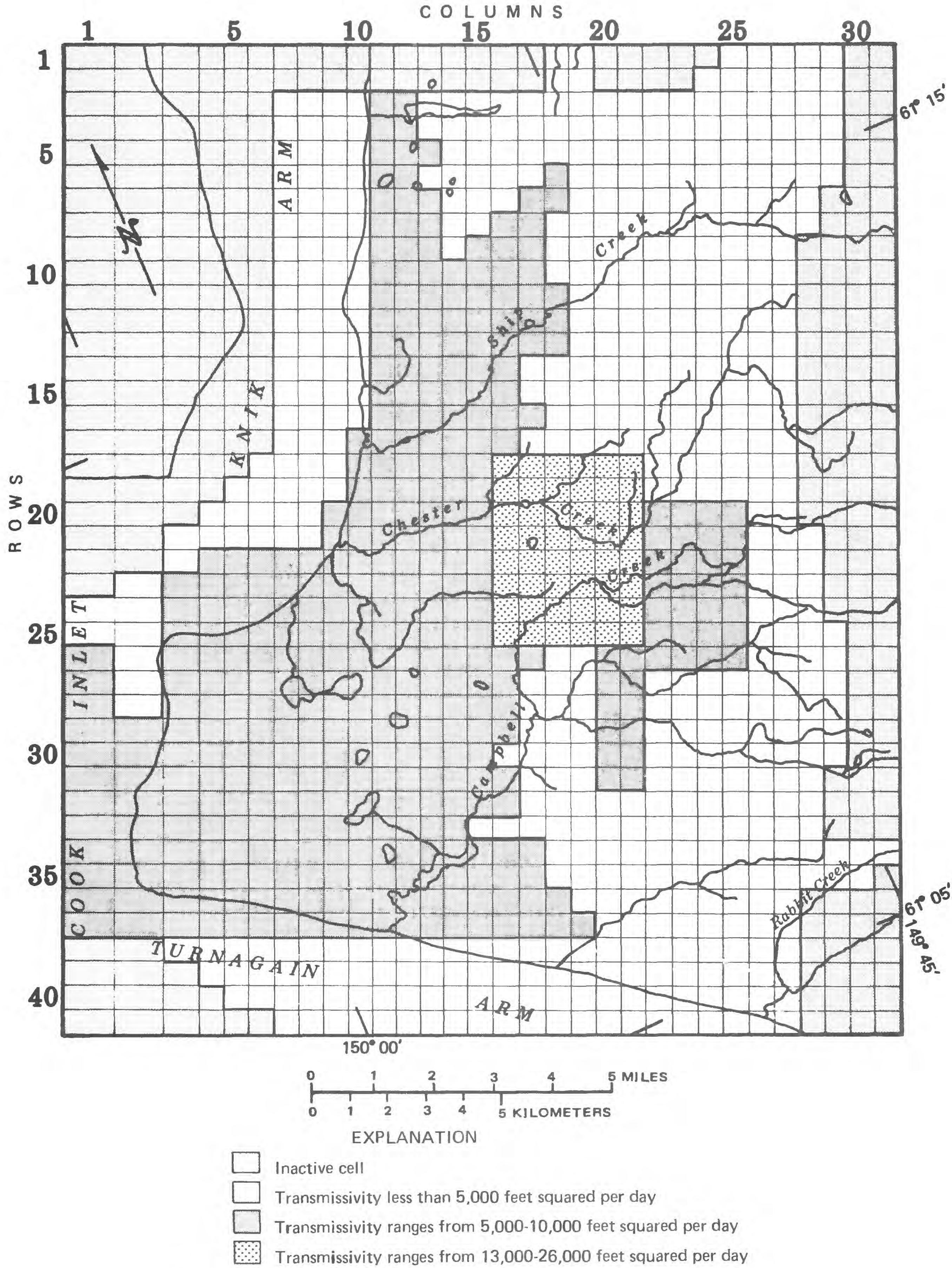


Figure 15.--Estimated transmissivity of the confined aquifer (layer 2).

aquifer caused by changes in fluid pressure. The usual range of storage coefficient for unconfined aquifers is 0.01 to 0.3 (Freeze and Cherry, 1979, p. 61) and a value of 0.15 was assumed for each cell in layer 1. Values of storage coefficient for confined aquifers range from 0.005 to 0.00005 (Freeze and Cherry, 1979, p. 60). Analysis of aquifer test data for wells completed in the confined aquifer indicates that storage coefficient values typically range from 1×10^{-4} to 1×10^{-5} throughout much of the Anchorage Bowl. A value of 10^{-4} was used for all but six active cells in layer 2.

Calibration of the Model

Calibration is a procedure by which differences between observed and simulated values of hydrologic factors are mathematically minimized so that the model will replicate the behavior of aquifers during both steady-state and transient conditions. This is accomplished by adjusting the input data such as aquifer properties, boundary conditions, and hydraulic stresses. Because of the many interrelated factors affecting ground-water flow, calibration is a subjective procedure. The degree of allowable adjustment of any parameter, however, is generally directly related to the uncertainty of its values. For example, because withdrawal rates are well defined, those values were not adjusted. Hydraulic conductivity and transmissivity, however, generally are imprecisely known because lithologic variation is usually not well defined and because the methods by which they are determined are subject to many limitations. In this study, the calibration goal was to have the model calculate heads (or drawdowns, the changes in head with time) that were within 20 ft of observed heads (or changes in head) for each cell and calculate stream discharges that were within 10 percent of observed stream discharge.

An additional part of the calibration process included the computation of water budgets for 17 regions of the modeled area (fig. 16). These budgets provided a measure of the relative contribution of each element of the recharge or discharge within areas of similar hydrologic characteristics and aquifer properties. For example, the recharge components for a particular region may consist of direct infiltration of precipitation, stream leakage, and flow from bedrock seepage, whereas the discharge components may consist of ground-water pumpage, stream discharge, or leakage to other regions. Calibration was aided by this ability to evaluate areas smaller than the entire model.

The model was first calibrated to steady-state conditions, under which recharge to the system equals discharge from the system, no water is derived from storage, and there is no change in head with time. Nonpumping and pumping situations were examined under steady-state conditions. The nonpumping model runs simulated conditions in the 1950's, before significant pumping began. The model runs for steady-state conditions with pumping simulated the effects of high rates of pumping for an infinitely long time. From inspection of the pumpage data from 1956-77, the highest value known to have been "pumped," 18.8 Mgal/d, was chosen for the simulation. Although no data are available to calibrate the model under conditions of prolonged pumping at 18.8 Mgal/d, it was desirable to ascertain that the model did not "self-destruct," or violate boundary conditions, at this pumping rate.

After the "best fit" of the model was obtained under steady-state, nonpumping conditions, a number of transient model runs were made to

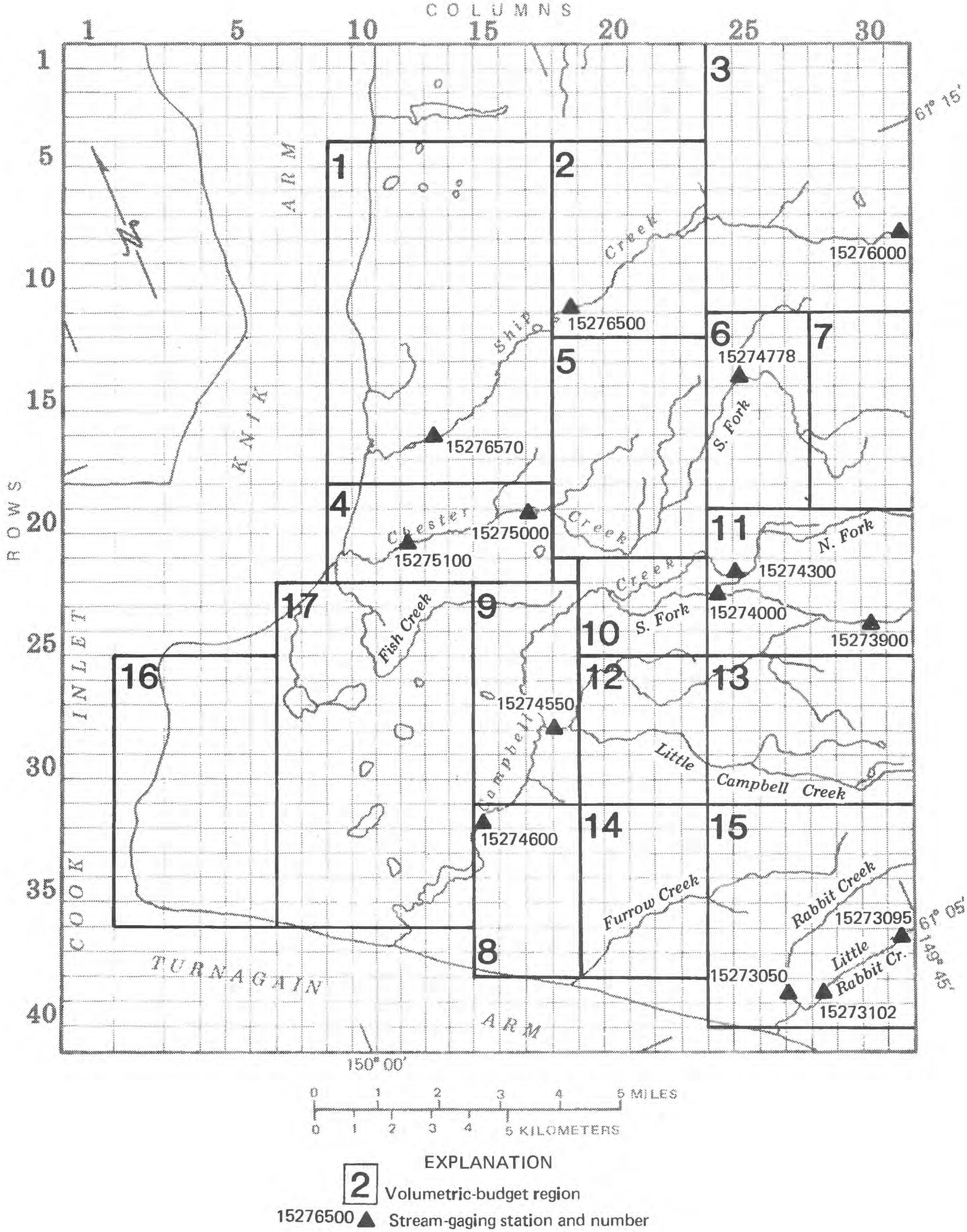


Figure 16.--Location of stream-gaging stations and volumetric-budget regions.

determine whether the model would produce a reasonable simulation of pumping conditions. However, a full calibration of the transient model was not done.

Steady State

Model simulations that used the hydraulic inputs, properties, and boundaries described above produced flow patterns that resembled pre-pumping (pre-1956) conditions (figs. 17 and 18). In some areas, however, calculated heads were more than 20 ft higher or lower than observed heads (figs. 19 and 20). These areas include the mountain front, downtown Anchorage, and north of Ship Creek. One factor that contributes to the poor calibration is the sparsity of water-level and hydraulic property data in these areas. Another reason for disagreement is that an observed water level is the head at a point (observation well) whereas a model-calculated water level is the average head for a model cell and is referenced to the center of the cell that contains the well. This discrepancy between calculated and observed water levels is greatest in cells where the hydraulic gradient is steepest.

Computed stream discharges compared favorably with observed streamflows. The 10 percent criterion was met for all sites except for two sites on Chester Creek and a site on the middle reach of Ship Creek (table 1) where hydraulic properties of the unconfined aquifer and streambed are not known with certainty.

Because the calibration procedure did not produce a precise correspondence between model heads and field-measured heads, and in order to assess our model formulation and the values of the input parameters used, additional model runs were made. The purpose of these runs was to test the model's sensitivity to changes in aquifer properties and boundary conditions. For each run a single aquifer property or condition was changed (tables 2 and 3) and the effects of this change on heads and volumetric budgets were observed. A "sensitivity analysis" was made only on aquifer properties for which actual data were available. For example, the hydraulic conductivity of layer 1 and transmissivity of layer 2 were based on actual data. Because the transmissivity of layer 1 and hydraulic conductivity of layer 2 were "computed" values, no sensitivity analyses were made on these properties.

Changes in heads were analyzed using the mean square error statistical parameter. The mean square error is computed as follows:
1) Compute the mean:

$$\text{Mean} = \sum_{h-1}^{N} (h_o - h_c) / N ,$$

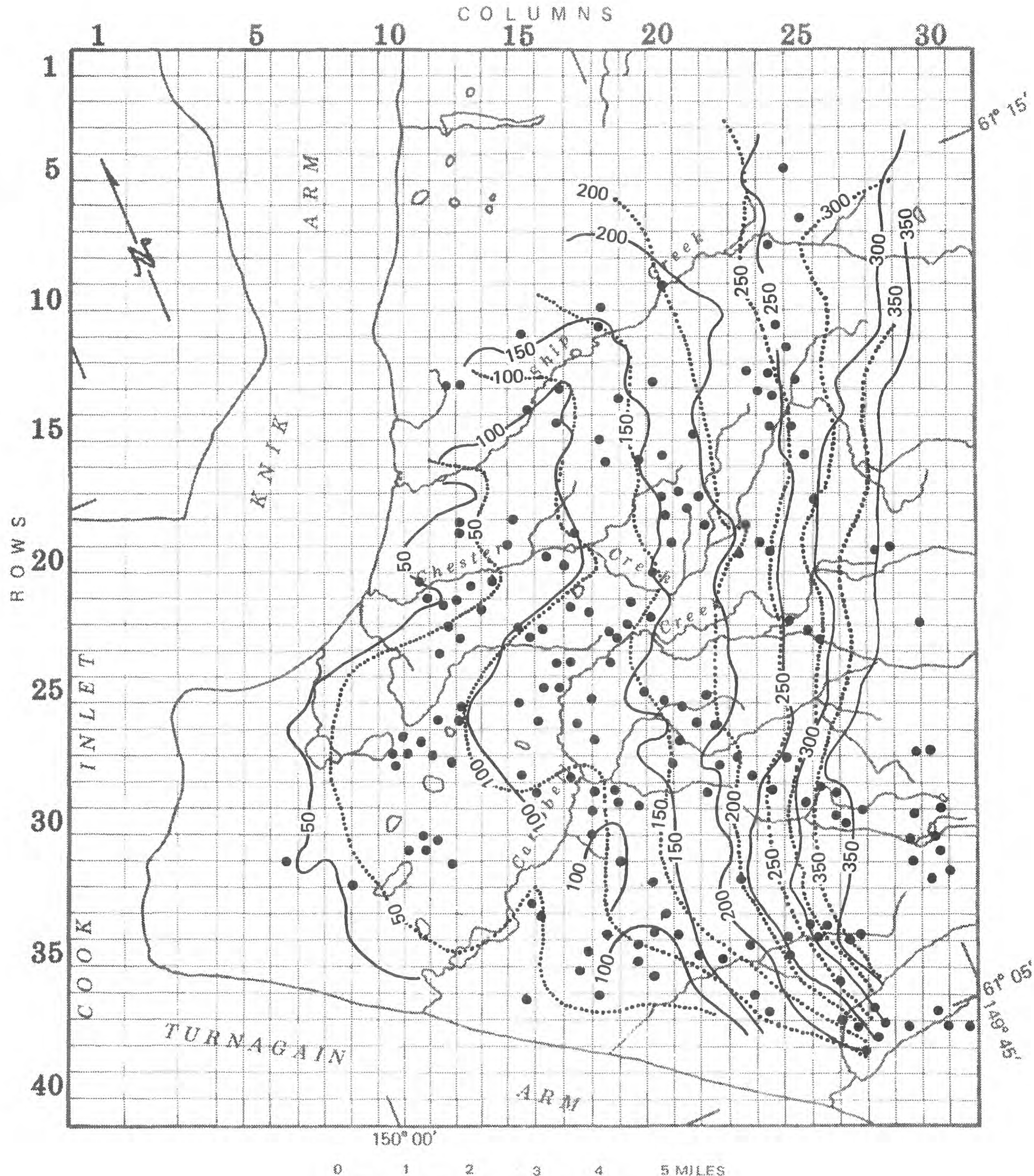
where h_o and h_c are observed and computed heads, respectively, and N is the number of observations.

2) Compute the variance (VAR):

$$\text{VAR} = \sum_{h-1}^{N} [(h_o - h_c) - \text{mean}]^2 / N\text{-}1.$$

3) Compute the mean square error (MSE):

$$\text{MSE} = (\text{mean})^2 + \text{VAR}.$$



EXPLANATION

100 MEASURED WATER-TABLE CONTOUR--Shows altitude of water table. Contour interval 50 feet. Datum is sea level.

100 SIMULATED WATER-TABLE CONTOUR--Shows altitude of water table Contour interval 50 feet. Datum is sea level.

Figure 17.--Observed and simulated elevations of the pre-1956 water table.

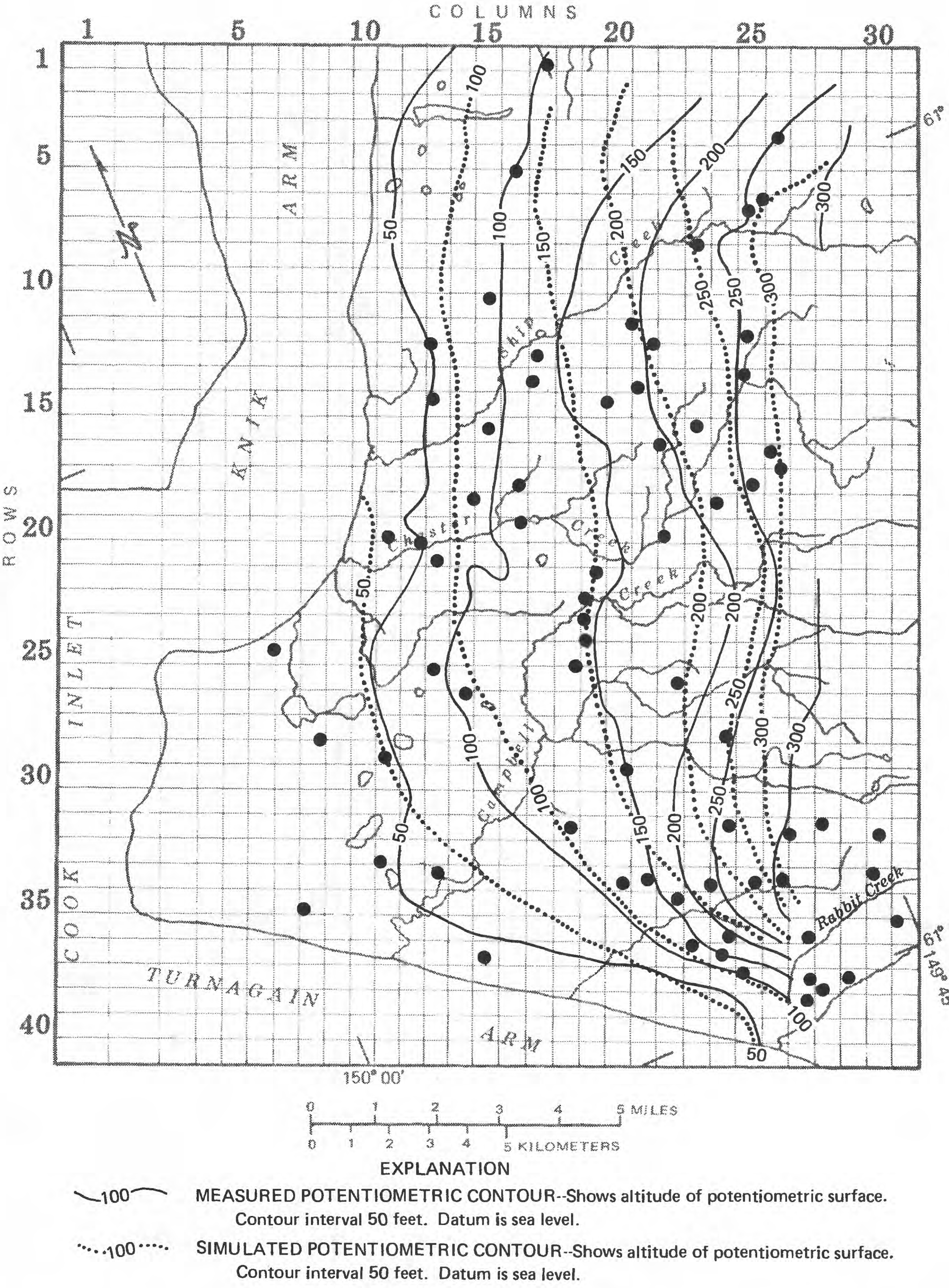


Figure 18.--Observed and simulated elevations of the pre-1956 potentiometric surface.

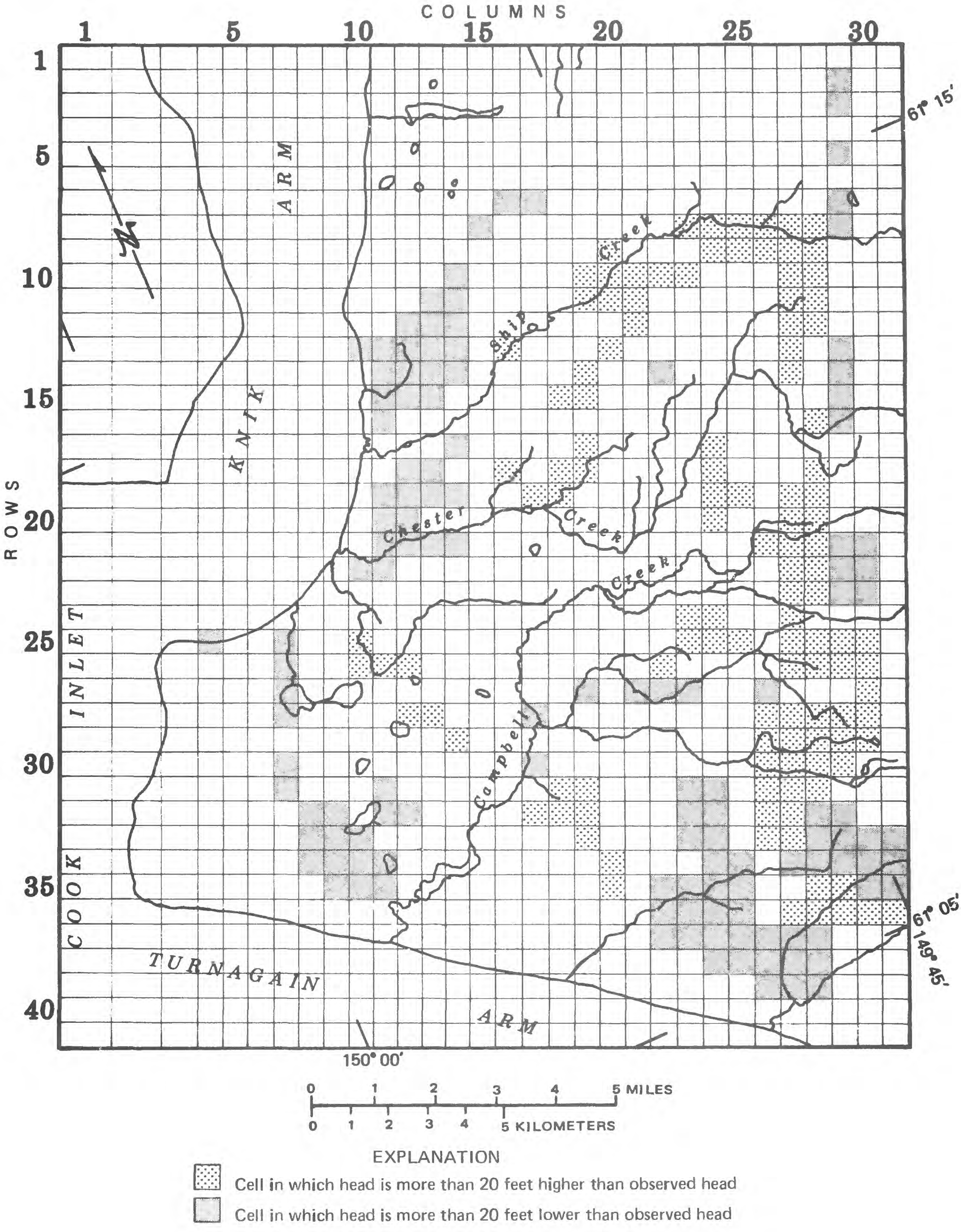


Figure 19.--Areas of the unconfined aquifer (layer 1) where calculated heads were more than 20 feet higher or lower than observed head.

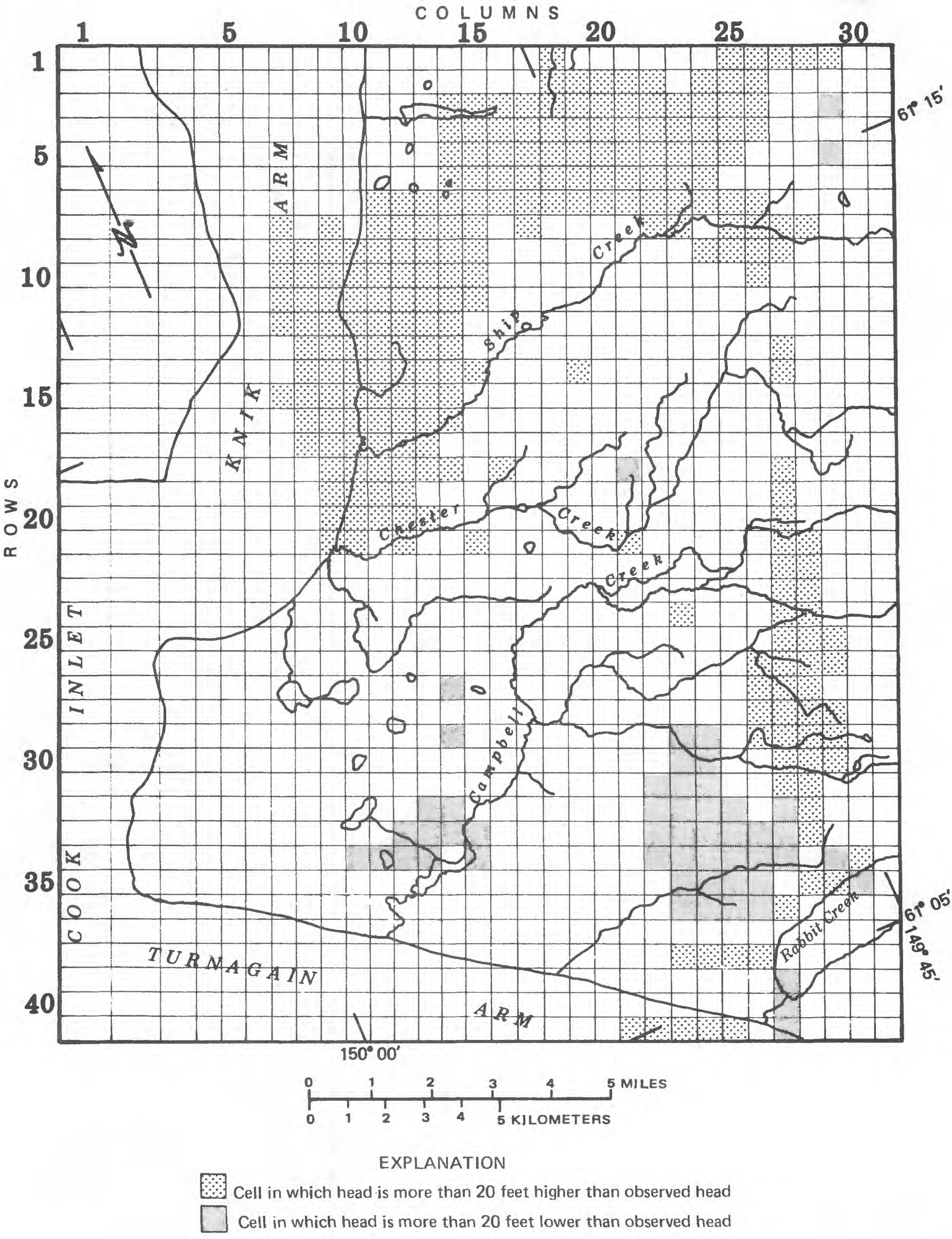


Figure 20.--Areas of the confined aquifer (layer 2) where calculated heads were more than 20 feet higher or lower than observed head.

Table 1.--Comparison of computed streamflow and observed streamflow in layer 1

[ft^3/s, cubic feet per second]

Station number (fig. 16)	Station name	Period of record	Cell location Row	Column	Total mean annual flow (ft^3/s) Observed	Computed	Percent difference
15273050	Rabbit Creek at Anchorage	1979-80	39	27	28.6a	13.54	(b)
15273900	South Fork Campbell Creek at Canyon Mouth	1966-79 1981	24	28	40.7	41.00	+0.07
15274000	South Fork Campbell Creek near Anchorage	1947-71	23	24	38.3	40.29	+5.2
15274300	North Fork Campbell Creek near Anchorage	1974-83	22	25	18.2	19.61	+7.7
15274600	Campbell Creek near Spenard	1966-83	30	17	65.5	66.03	+0.8
15274798	South Branch South Fork Chester Creek at Anchorage	1981-83	14	25	7.5	7.03	-6.3
15275000	Chester Creek at Anchorage	1958-76	20	17	18.2	11.79	-35.2
15275100	Chester Creek at Arctic Boulevard	1966-83	21	12	18.1	13.97	-22.8
15277600	Ship Creek at Anchorage	1946-83	9	29	145	145	0.0
15276500	Ship Creek at Elmendorf Air Force Base	1963-71	11	18	113	131	+15.9
15276570	Ship Creek below powerplant	1970-81	17	13	144	146	+1.4

[a] A more accurate long-term mean is expected to be about 18 ft^3/s

[b] Not calculated due to short period of record

Table 2.--Changes in aquifer properties and boundary conditions used in sensitivity analyses

Run number	Property	Multiplication factor
1	Recharge	4
2		2
3		0.5
4		0.25
5	Transmissivity-layer 2	10
6		2
7		0.5
8		0.1
9	Hydraulic conductivity-layer 1	10
10		2
11		0.5
12		0.1
13	Head-dependent boundary-conductance	10
14		2
15		0.5
16		0.1
17	Drain conductance	10
18		2
19		0.5
20		0.1
21	Streambed conductance	10
22		2
23		0.5
24		0.1

Table 3.--Changes in vertical conductance of the confining layer used in sensitivity analyses

Run number	Hydraulic conductivity regions (fig. 14)	Multiplication factor
25	3	100
26	1,2,5	100
27	3	10
28	1,2,5	10
29	3	2
30	1,2,5	2
31	3	0.5
32	1,2,5	0.5
33	3	0.1
34	1,2,5	0.1
35	3	0.01
36	1,2,5	0.01
37	2	10
38	2	0.1
39	5	10
40	5	0.1
41	1	10
42	1	0.1
43	3,4	10
44	3,4	0.1
45	3	10
	4	0.01
46	3	0.1
	4	10
47	4	10
48	4	0.1
49	2	10
	1	0.1
50	2	0.1
	1	10

The mean square error was chosen because a small error might indicate nearly equal amounts of positive and negative error, whereas a small variance might indicate a small scatter of errors about a large mean error.

For the model as a whole, the largest increases in mean square errors resulted when the hydraulic conductivity of layer 1 was divided by 10 (fig. 21). The next largest MSE was noted when the transmissivity of layer 2 was multiplied by 10. Analysis of the individual layers (figs. 22 and 23) showed that for layer 1 the computed head values were most sensitive to a decrease in hydraulic conductivity or an increase in the transmissivity of layer 2 whereas the computed head values for layer 2 were most affected by an increase in transmissivity.

In order to evaluate sensitivity of the sub-areas of the model (fig. 7) to changes in aquifer properties and boundary conditions, 50 model runs were made with the properties varied (tables 2 and 3). Generally, computed heads in the mountain-front area were most sensitive to changes in aquifer properties for both layers. Heads in layer 1 in the Klatt area were affected by an increase in recharge and heads in layer 2 in the Elmendorf area were most affected by a decrease in the elevation of the general head boundary.

The steady-state model was less sensitive to changes in vertical conductance of the confining layer (table 3) than was anticipated. Except when the vertical conductance was reduced by a factor of 100, the differences from initial heads due to changes in vertical conductance (as observed through the mean square error technique) were small. Almost all changes in vertical conductance had a significant impact in the mountain-front area of both aquifers.

Steady-state model runs were next made using a pumping figure of 18.8 Mgal/d. A comparison was made between the calculated water budgets for steady-state conditions without pumping and for steady-state conditions with pumping at 18.8 Mgal/d. For this run, the unconfined aquifer showed an increase in recharge from streams and a corresponding decrease in discharge to the streams. The decrease in stream discharge is about 5 percent of the total discharge (layer 1 plus layer 2) from the modeled area. The simulated pumping causes considerably more leakage from the unconfined aquifer into the confined aquifer and a decrease in flow to Cook Inlet from the confined aquifer. It should be understood that these observations are only qualitatively valid and that additional calibration work, including data collection, will be necessary to create a predictive model.

The possible effects of ceasing all ground-water pumping also were evaluated. Model results show that heads in the confined aquifer recover more quickly than do heads in the unconfined aquifer. Within 1 month of the cessation of pumping (as simulated in the model), the node in row 22, column 18 layer 2, which had a drawdown of approximately 30 ft, recovered 7 ft (fig. 24). After 6 months, the node had recovered to within 7 ft of pre-pumping levels. Recovery in layer 1 was much slower; in 1 month there was

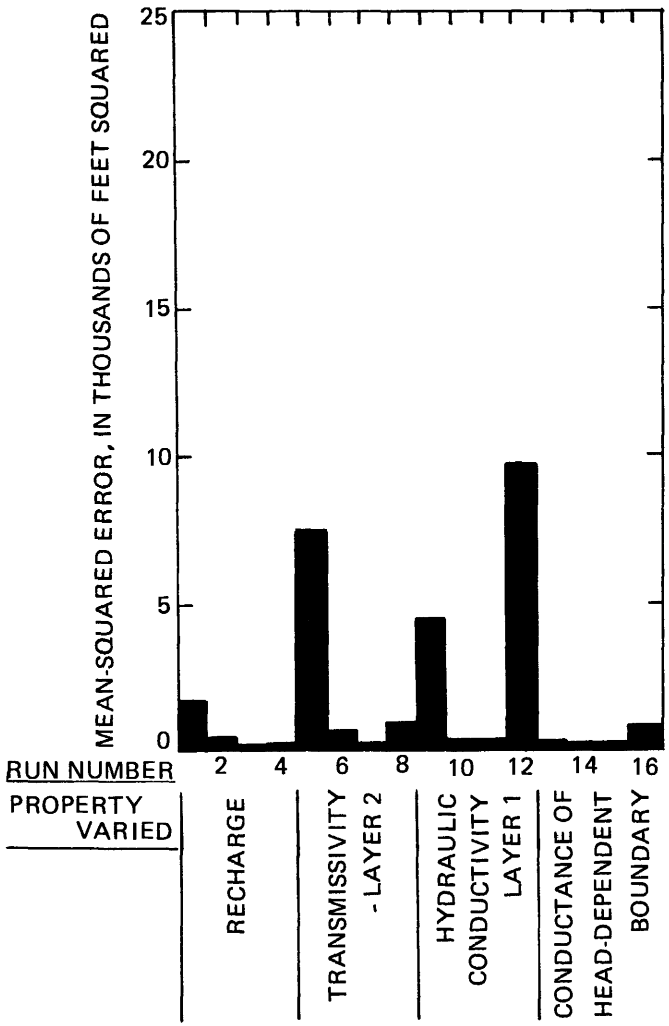


Figure 21.--Results of sensitivity analysis for steady-state conditions, for the model as a whole.

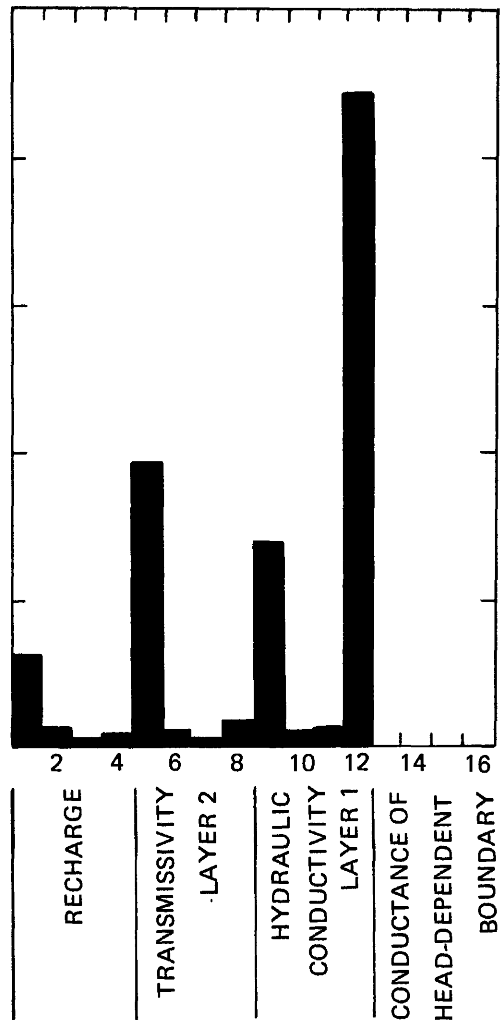


Figure 22.--Results of sensitivity analysis for the unconfined aquifer (layer 1), steady-state conditions.

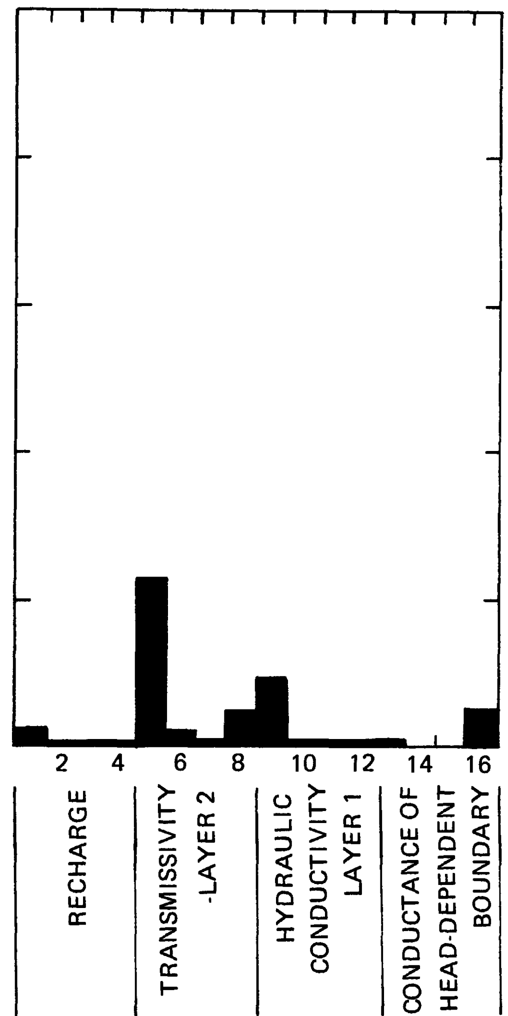


Figure 23.--Results of sensitivity analysis for the confined aquifer (layer 2), steady-state conditions.

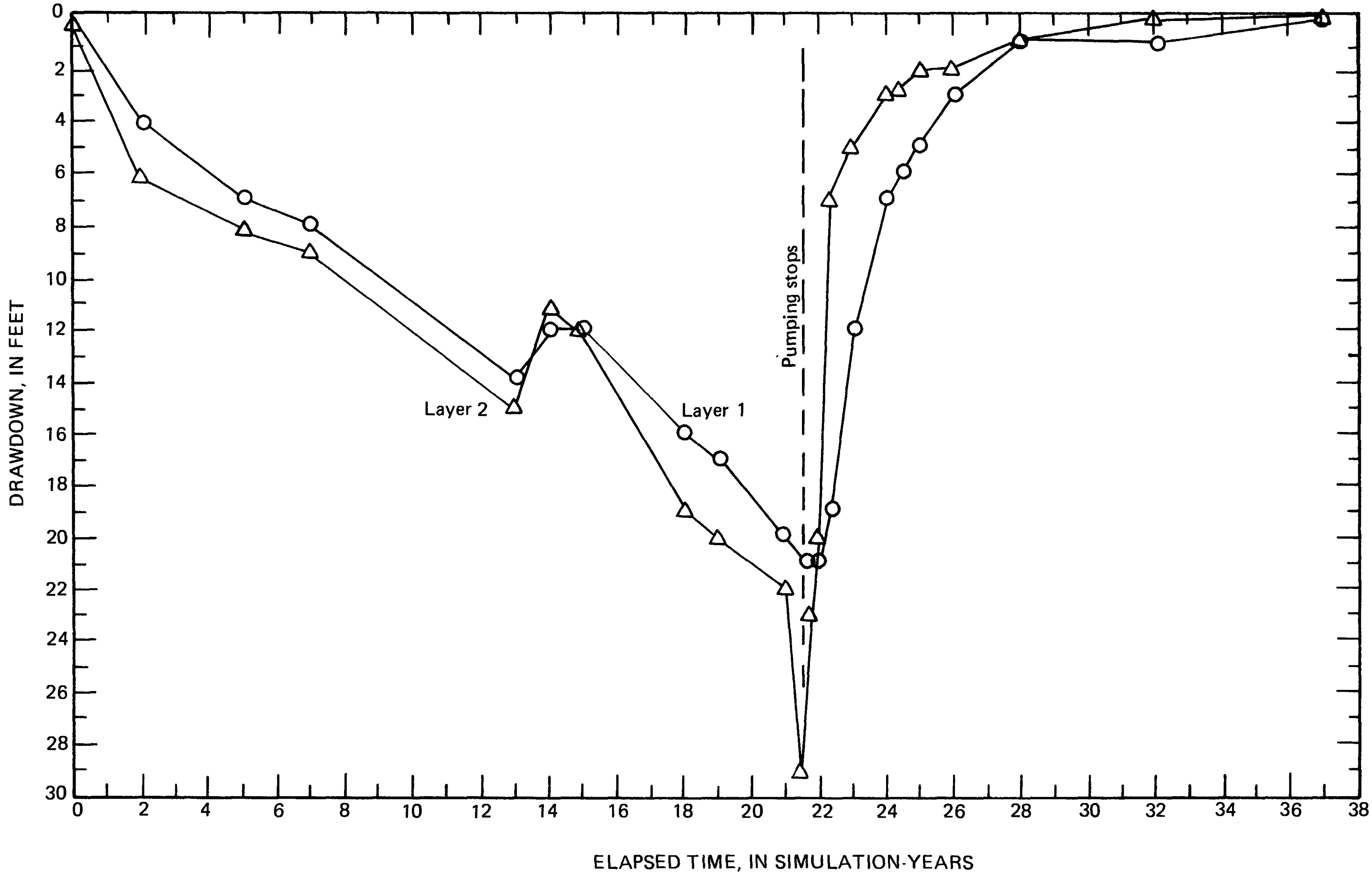


Figure 24.--Calculated drawdown and recovery as simulated by the model at node 22, 18.

about 1 ft of recovery for the model as a whole. After 20 years, three nodes in layer 1 showed 2 ft of drawdown from pre-pumping levels whereas the heads in all nodes in layer 2 had returned to within 1 ft of pre-pumping water levels.

Transient

Pumping records, based on data provided by the several public water suppliers in the Anchorage Bowl and on estimates of industrial use and per-capita consumption, were available from 1956 through 1977. This period was divided into 21 pumping periods (fig. 25). The first 9 periods were based on groups of consecutive years within which pumpage totals were similar, and the last 12 were based on monthly pumpage. Ground-water pumpage was simulated by assigning particular cells in layer 2 a pumping rate based on actual or estimated rates. For example, the pumpage for June 1977 (about 18 Mgal/d) was distributed among the cells shown in figure 26. The final model heads generated from the steady-state runs were used as the initial heads for the transient runs. This insures that the initial head data and the model hydrologic inputs and parameters are consistent.

For each production well, comparisons were made between the model-simulated water levels, water levels from nearby observation wells, and the quantity of water pumped from the production well. Because it was believed that there were not enough observation wells in the unconfined aquifer, results are presented only for the confined aquifer, in which the production wells are completed. Figures 27, 28, and 29 show the comparisons from three production wells that were selected to represent various locations in the Anchorage Bowl.

In general, the water levels simulated by the model respond to the different pumping rates and follow the same trends as the observed water levels. By comparing the net change in water level (highest water level minus lowest water level) between the model-simulated and observed values, there appears to be closer agreement with wells that had a relatively low production rate (less than 0.5 Mgal/d) than those with a higher production rate (greater than 1.0 Mgal/d).

Two important factors should be noted in these transient simulations. First, the model gives the results for the center of a particular node and neither the particular production well nor the observation well are located exactly in the center of the node. Second, no additional runs were made using different values of the storage coefficient or other hydraulic properties. Further study of these factors is necessary before the model can be used for predictive purposes.

Consideration of the Confining Layer

For this study, the model was configured to represent the ground-water system in the Anchorage Bowl as a two-layer system -- a confined aquifer and an unconfined aquifer separated by an "implicit" confining unit. Under this condition, all flow in the confining layer is assumed to be vertical and the vertical conductance represents the "resistance" to vertical flow. For model simplification it was assumed that no water was derived from storage in the confining unit. To test the validity of the "no-storage assumption,"

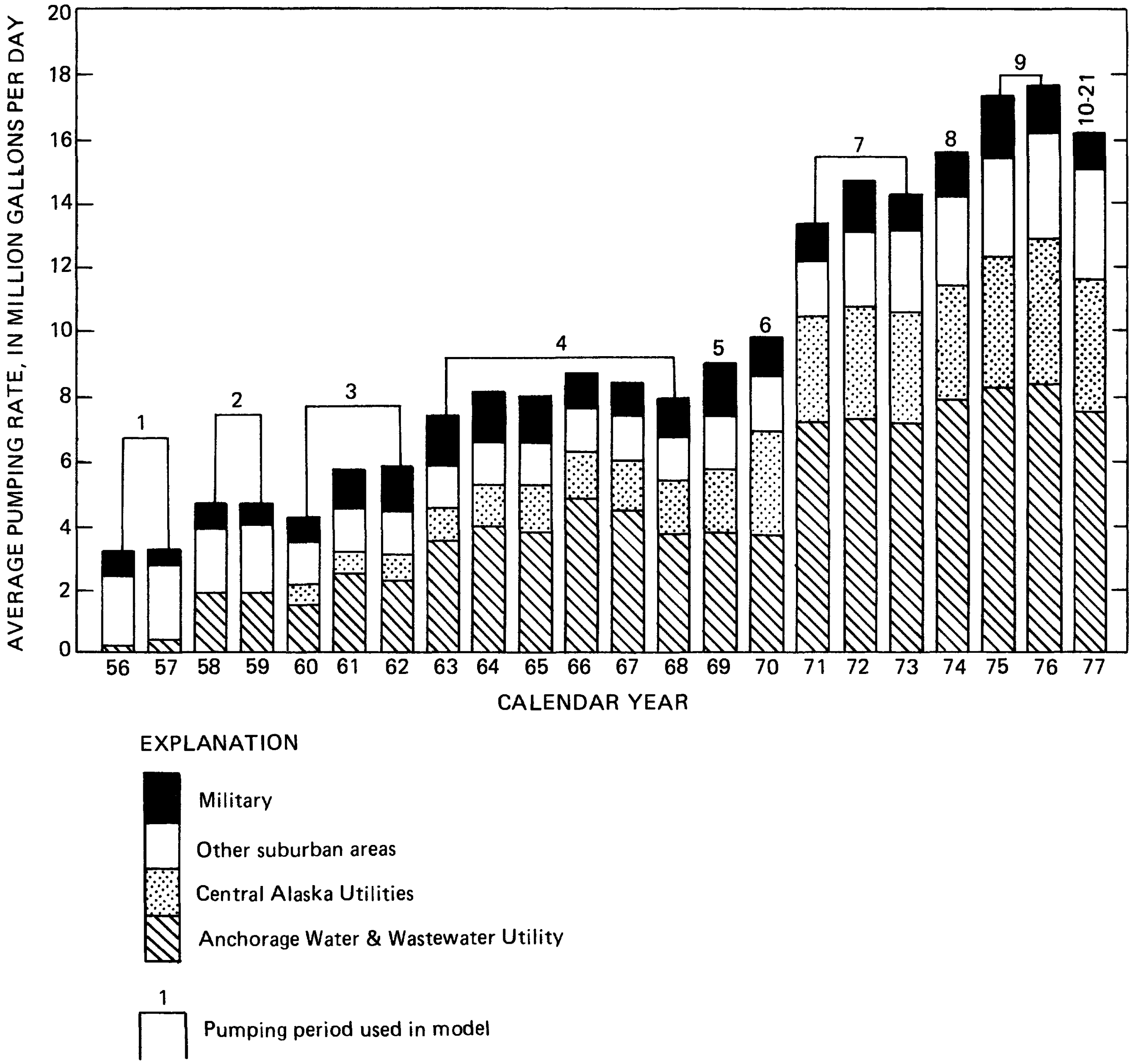
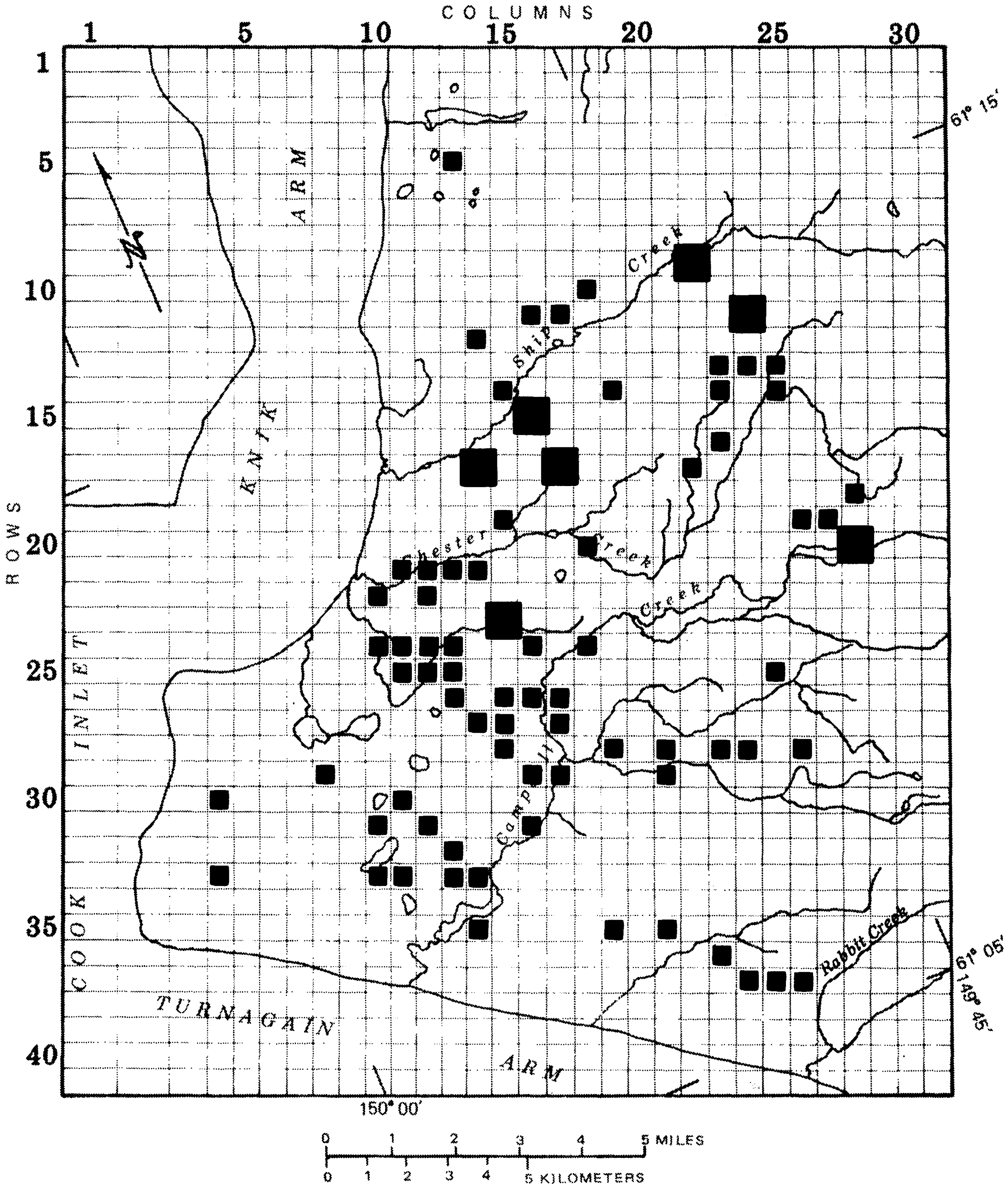


Figure 25.--Estimated average yearly pumping rate from the confined ground-water system, 1956-77.



EXPLANATION

Pumping rate less than 748,000 gallons per day

Pumping rate greater than 748,000 gallons per day

Figure 26.--Distribution of simulated pumping, June 1977.

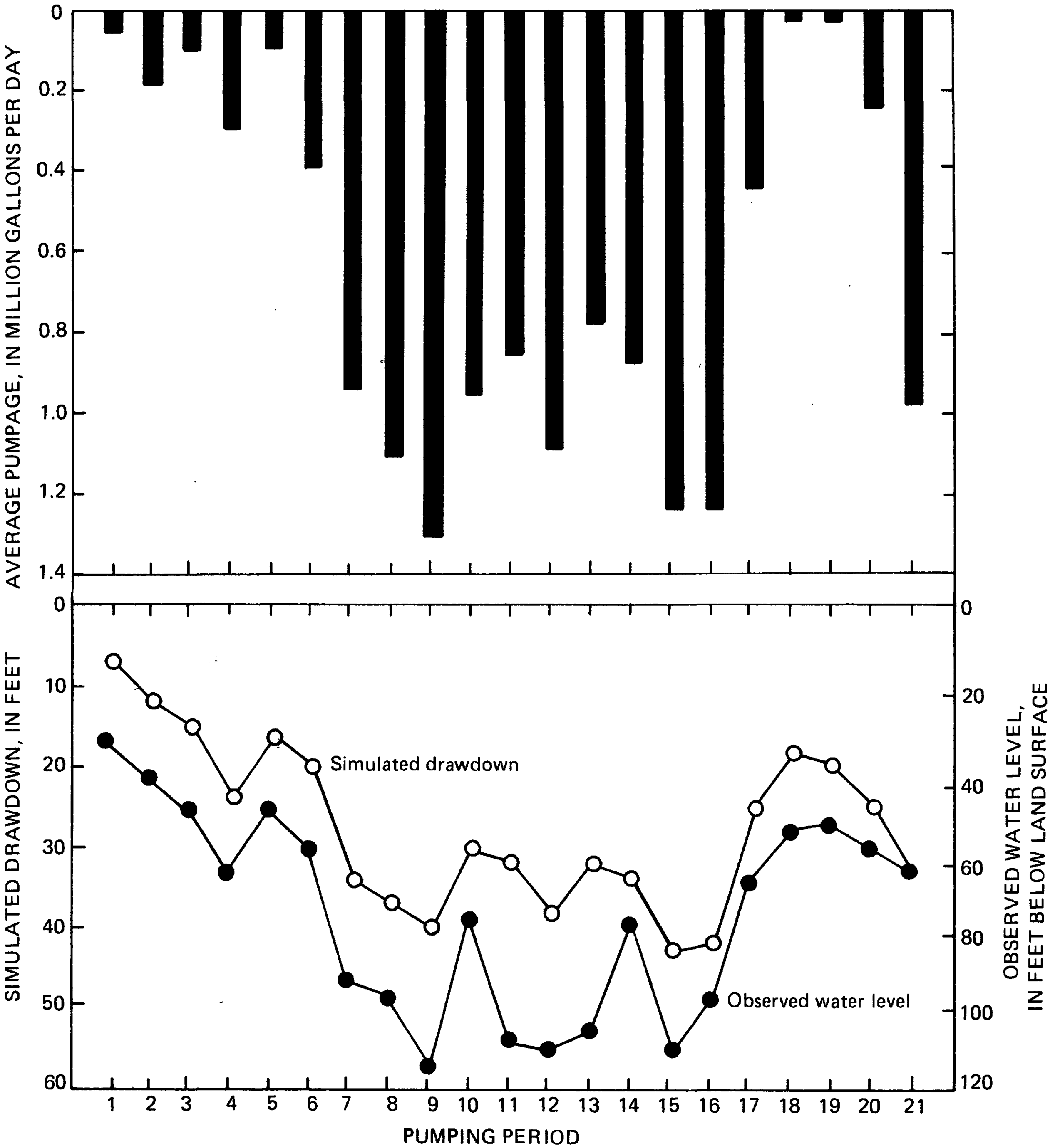


Figure 27.--Comparison of pumpage, observed water level, and simulated drawdown for node 17, 14, 2. (See fig. 25 for details of pumping periods.)

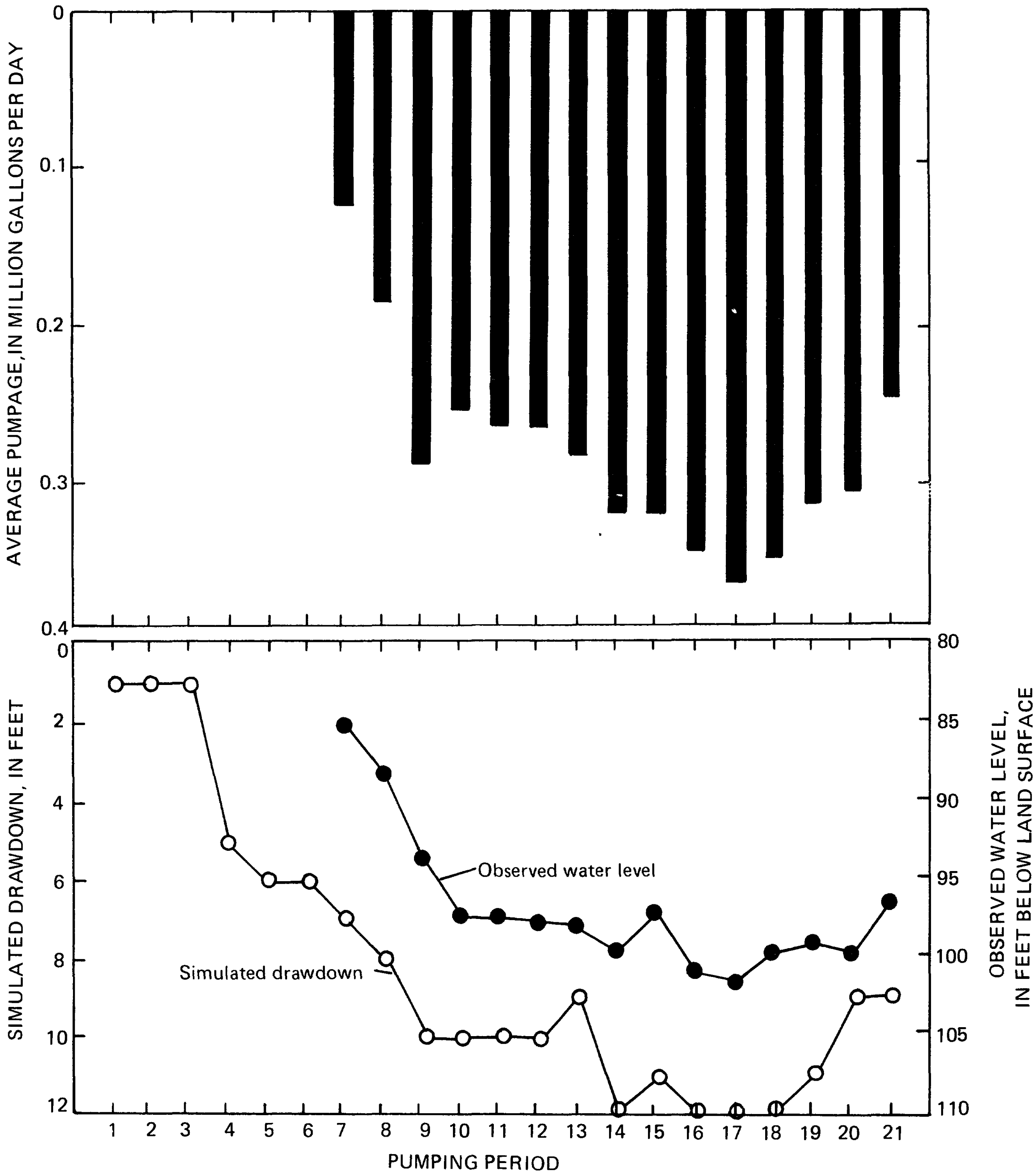


Figure 28.--Comparison of pumpage, observed water level, and simulated drawdown for node 23, 15, 2. (See fig. 25 for details of pumping periods.)

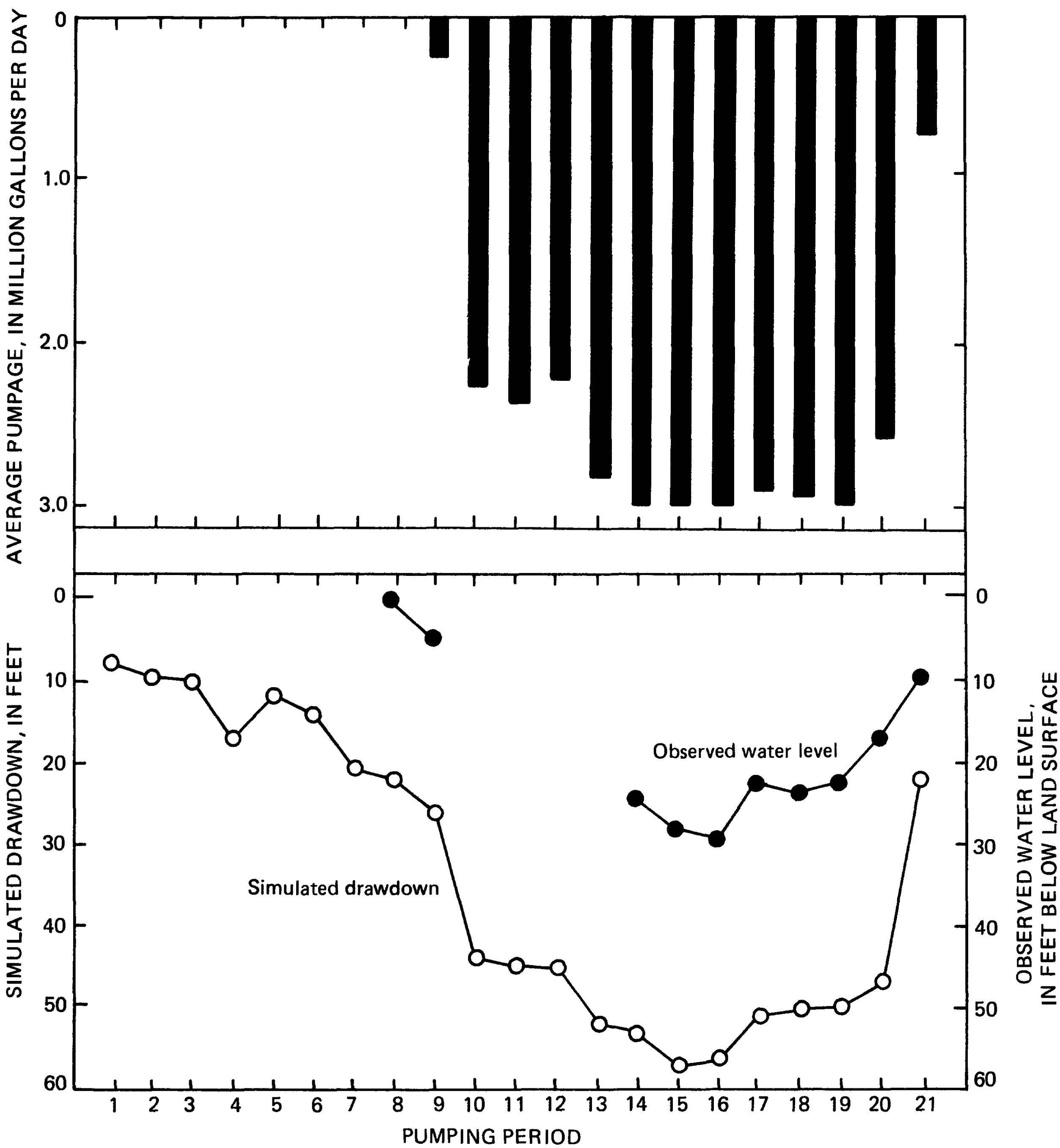


Figure 29.--Comparison of pumpage, observed water level, and simulated drawdown for node 36, 23, 2. (See fig. 25 for details of pumping periods.)

estimates of the time needed for the effects of a head change in the unconfined aquifer to propagate through the confining layer and for steady-flow conditions to be established in the confining layer were determined using a technique described by Nichols (1977).

According to Nichols, dimensionless time, T_D is given by :

$$T_D = \frac{K't}{S_s'\ell^2},$$

where K' is the hydraulic conductivity of the confining layer,

S_s' is the specific storage of the confining layer,

ℓ is the thickness of the confining layer, and

t is the real time.

When dimensionless time equals about 1×10^{-1}, the effects of a head decrease in the aquifer will have extended through the confining layer and the overlying or underlying aquifers will be affected. Real time is then estimated using the equation:

$$t = \frac{(1 \times 10^{-1})S_s'\ell^2}{K'}$$

A value of 2.3×10^{-6} ft^{-1} was used as a value for S_s'. This value was based on previous work by Nelson (1982, p. 10). Values of ℓ and K' were the same as used in the modeling process. Determining the values of t for the entire grid indicated that the maximum amount of time was 60 days. Assuming that the values of S_s', ℓ, and K' are accurate, would thus indicate that leakage is not significant in the confining layer.

Data Needs to Refine Model

Although a precise mathematical duplication of the ground-water system of the Anchorage Bowl would be impractical owing to the complex geology of the area, the ability of the model to represent the system would be improved if efforts were made to:

1) Conduct aquifer tests that could provide hydraulic property information where it is lacking. These tests would require strict control of pumping conditions, a number of properly positioned observation wells, and long pumping periods.

2) Determine precipitation and evapotranspiration rates throughout the area.

3) Determine in detail the rates and locations of water withdrawals by domestic, municipal, and self-supplied industrial users.

4) Define more accurately the rate at which water seeps into unconsolidated materials from the bedrock of the Chugach Mountains.

5) Accurately determine the head conditions near the aquifer boundaries, and improve definition of streambed characteristics and stream losses and gains.

6) Compare drillers' logs and borehole geophysical logs to improve the accuracy of geologic interpretations

7) Continue long-term monitoring of water levels in observation wells in both the confined and unconfined aquifers.

8) Measure water levels in additional wells periodically to produce accurate potentiometric contour maps that can be used to improve model calibrations.

SUMMARY AND CONCLUSIONS

The ground-water system of Anchorage, Alaska was analyzed by constructing a two-layer, three-dimensional, mathematical model. The model was configured such that layer 1 represented an upper unconfined aquifer and layer 2 represented a deeper confined aquifer. A layer of poorly permeable materials that separates the aquifers was not simulated (that is, hydraulic heads in this layer were not calculated) but the hydraulic properties of the geologic material were used to calculate hydraulic connection and movement of water between the two active layers.

Under steady-state conditions, directions of ground-water flow calculated by the model were similar to observed directions of flow, and computed stream discharges were generally within 10 percent of observed values. However, in many parts of the modeled area, calculated heads were more than 20 ft higher or lower than observed heads.

For steady-state simulations, model results are most sensitive to changes in the hydraulic conductivity of layer 1 and the transmissivity of layer 2. The area most sensitive to changes in aquifer properties is the mountain front, and additional hydraulic data is needed in this area.

Pumping conditions under steady state were simulated to determine the general effects on ground-water levels and streamflows. Model results indicate that when 18.8 Mgal/d is pumped, ground-water levels were significantly lower, streamflows decreased, and leakage from the unconfined aquifer into the confined aquifer increased. Flow from the confined aquifer into Cook Inlet also decreased.

Transient simulations were made for 21 pumping periods. These initial runs indicate that drawdowns of production wells follow the trends of observation wells. However, more detailed calibration and sensitivity analyses are needed in order to use the model as a predictive tool. By use of analytical techniques the amount of leakage from the confining layer appears to be minor in most areas.

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