Delineation of Areas Contributing Recharge to Municipal Wells in Three Selected Confined-Glacial Aquifers in Erie County, New York

By Richard M. Yager, Todd S. Miller, and John Thayer

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To Obtain
	Length	
inch (in.)	2.54	centimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Area	
square foot (ft^2)	0.09290	square meter
square mile (mi^2)	2.590	square kilometer
acre	0.40483	hectare
	Volume	
cubic feet (ft^3)	0.02832cubic meter	
	Flow	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
million gallons per day (Mgal/d)	3.785	cubic meters per day
gallons per minute (gal/min)	0.06309	liter per second
gallons per second (gal/s)	0.001052	liter per second
	Hydraulic Conductivity	
foot per day(ft/d)	0.3048	meter per day
	Transmissivity	
square foot per day (ft^2/d)	0.09290	square meter per day

National Geodetic Vertical Datum of 1929 (NGVD) is 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

Areas that contribute water to municipal wells screened in three glacial aquifers that supply water to the Village of Alden and the Towns of Collins and Holland in Erie County were delineated through conceptual ground-water-flow models that were based on hydrogeologic information from previous studies. Seismic refraction profiles, stream-discharge measurements, and streambed inspections were conducted to obtain additional information on depth to bedrock and the degree of hydraulic connection between aquifers and streams. Uncertainty in the location and extent of the areas contributing recharge to the municipal wells was examined through a comparison of areas delineated with conceptual models representing alternative hypotheses concerning aquifer boundaries and hydraulic properties; this comparison indicated what additional information would be needed to decrease the uncertainty in the contributing-area delineations.

The Village of Alden's municipal wells are screened in a shallow, confined aquifer overlain by fine-grained sediments deposited during a readvance of glacial ice. The ground-water system was simulated with a two-layer model representing the confined aquifer and the underlying weathered-shale bedrock, and recharge to the confined aquifer was derived from downward leakage through the confining layer and underflow through the weathered bedrock from upgradient areas. MODFLOWP, a parameter-estimation method based on nonlinear regression, was used to calibrate the model to ground-water levels measured in 18 wells and ground-water discharge to Spring Creek. Areas contributing recharge to four municipal wells ranged from 0.045 to 0.07 square miles, and from 7 to 28 percent of this recharge originated as runoff from upland areas upgradient to the south; the remainder was derived from precipitation on areas overlying the aquifer. Decreasing the vertical-leakage rate from 24 to 10 inches per year in an alternative model increased the size of the contributing areas by 40 to 90 percent and increased the percentage of recharge that entered the aquifer from upland areas as underflow and unchanneled runoff by 10 to 40 percent.

The Town of Collins municipal well no. 1 is screened in a deep, confined sand and gravel aquifer in a buried bedrock valley eroded by a tributary to the ancestral Allegheny River valley. The confined aquifer is overlain by fine-grained sediments that are, in turn, overlain by an unconfined aquifer composed of alluvial and deltaic sand and gravel with a maximum thickness of 550 feet, except in areas near the valley walls where the aquifer extends to land surface. The confined aquifer is recharged at these exposures by precipitation and infiltration from streams and upland runoff, and at depth by underflow from upgradient parts of the buried bedrock valleys to the north, east and south. Ground-water flow was simulated with a one-layer numerical model representing average steady-state conditions in the confined aquifer. Two alternative models were considered: in model A a large conductance value (100 feet squared per day) was used along the head-dependent flow boundary to represent a large rate of recharge from the uplands, and in model B, a small conductance value (20 feet squared per day) was used to represent less recharge from the uplands. In model A, the areas contributing recharge to the municipal well within the aquifer and upland areas were 0.29 and 0.95 square miles, respectively, and all water pumped from the municipal well was derived from upland areas as underflow or unchanneled runoff. In model B, the contributing areas were 0.87and 2.0 square miles, respectively, and 85 percent of the

pumped water was derived from upland areas, while the remainder was derived from underflow through a buried-tributary valley now occupied by Clear Creek.

The Town of Holland's municipal well is screened in a deep, confined aquifer within a narrow, glacially scoured bedrock valley bordered to the south by the Lake Escarpment moraine. The confined aquifer is overlain by lacustrine silt and clay deposited in a proglacial lake that formed between the moraine and the receding ice front. Ground-water flow through the confined aquifer was simulated with a two-layer model representing the confined aquifer and overlying glacial sediments, but the model was not calibrated because little information on ground-water levels and flow rates is available. The area contributing recharge to the Holland municipal well from morainal sediments covers an area of 0.063 square miles and provides about 35 percent of the water pumped from the well; the remainder of the recharge to the well flows through a part of the confined aquifer that covers 0.34 square miles within the modeled area and extends southward to upgradient areas outside the modeled area. Decreasing the vertical hydraulic conductivity of the morainal sediments by a factor of 10 in an alternative model resulted in a decrease in the contribution from leakage through the morainal sediments to less than 10 percent of the flow to the well; the remainder of the water pumped from the well was derived mainly from underflow through the southern boundary.

INTRODUCTION

Accurate delineation of areas that contribute water to wells is an essential step in protecting the ground-water supplies from contamination. The U. S. Environmental Protection Agency has recommended that municipalities adopt "Wellhead Protection" strategies to prevent ground-water contamination by regulating land use in areas that contribute water to public-supply wells (U. S. Environmental Protection Agency, 1987). The objective of these strategies is to define the surface area through which water enters the ground-water system and eventually flows to wells, and to develop policies that minimize contamination in these areas.

The Erie County Department of Environment and Planning is assisting in the development of Wellhead Protection strategies for municipal wellfields screened in three confined glacial aquifers that provide water to the Village of Alden and the Towns of Collins and Holland (fig. 1). The aquifers are glacial deposits of sand and gravel that supply water to the three communities, each of which has less than 5,000 people. In 1994, the U. S. Geological Survey (USGS), in cooperation with the Erie County Department of Environment and Planning, began a 1-year study to delineate the areas contributing recharge to the municipal wells that tap these aquifers.

The Erie County Community Development Block Grant program, the Village of Alden, the Towns of Collins and Holland, and the New England Interstate Water Pollution Control Commission provided funding for the study. A companion study of the Onondaga Limestone, which provides drinking water to the Village of Akron, was conducted as part of the same project (Kappel and Miller, 1996).

Purpose and scope

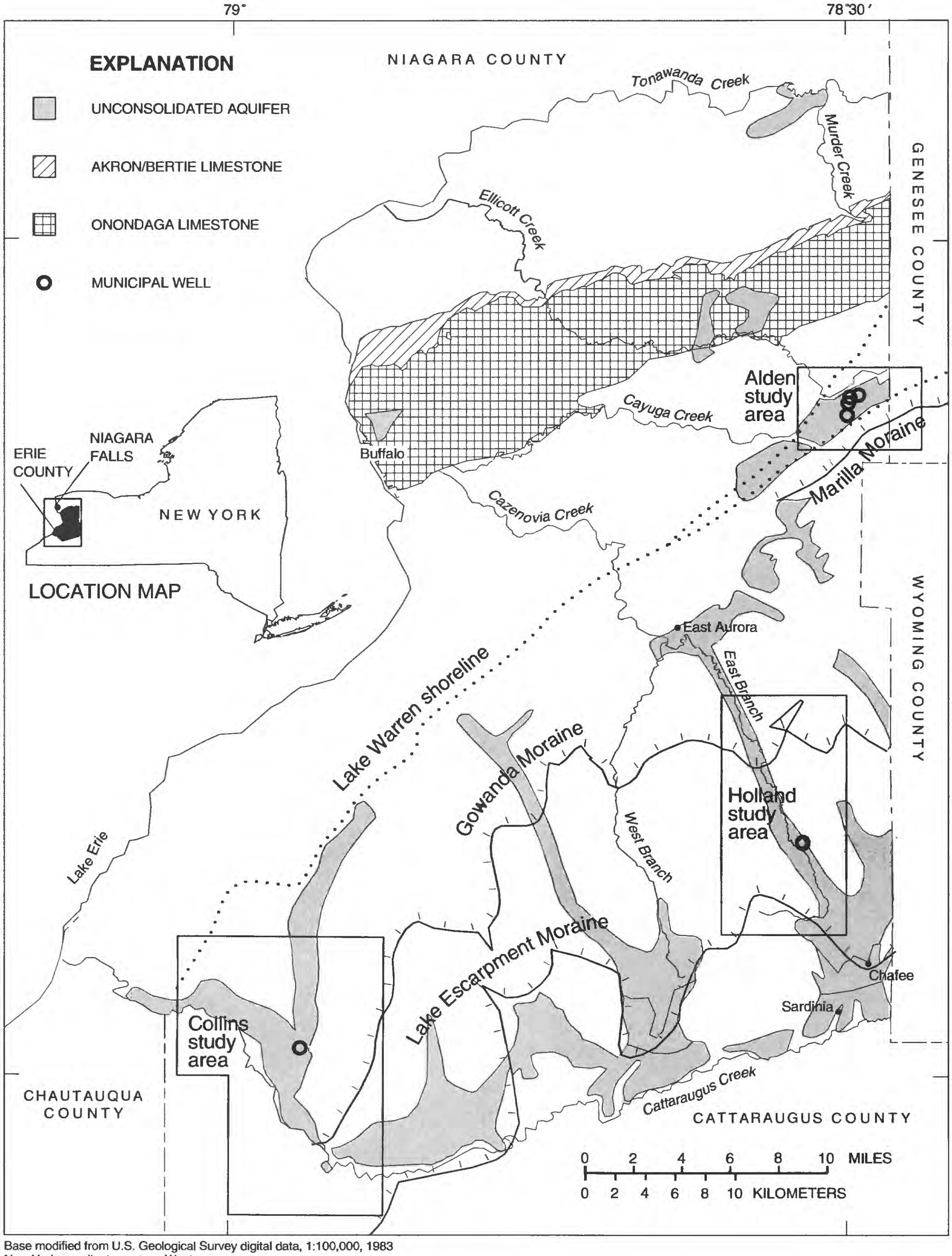
This report describes the hydrogeologic setting in each of three study areas and the ground-water-flow models used to delineate the contributing areas to confined wells in each area. It presents maps showing the distribution of hydraulic head in the principal aquifer in each area and the contributing areas to the municipal wells as indicated by the models. It also includes tables summarizing the inflows and outflows to each aquifer computed for average, steady-state conditions.

Acknowledgments

Spencer Schofield of the Erie County Department of Environment and Planning initiated this study and arranged for funding necessary for its completion. Bruce Sitzman, Superintendent of Public Works, Village of Alden; Gerald Barron, Superintendent of Water, Town of Holland; and Daniel Stroud, Collins Water Department, provided information on their respective municipal-water supplies.

METHODS OF ANALYSIS

Ground-water-flow models were developed to represent each of the three selected aquifers and to delineate the areas that contribute water to municipal wells. Hydrogeologic information from previous studies was supplemented by data from field surveys to construct numerical models of ground-water flow in each aquifer and to delineate the areas that contribute recharge to the wells. Where the boundaries and



Base modified from U.S. Geological Survey digital data, 1:100,000, 1983
New York coordinate system, West zone

Figure 1. Map showing pertinent geographic features of Erie County, N.Y. and locations of study sites

hydraulic properties of an aquifer were uncertain, alternative models of the aquifer were compared to investigate the reliability of the delineations. A relatively extensive database was available for calibration of the Alden area model, but data for models of the Collins and Holland areas were insufficient; therefore, these models were not calibrated and based on a general interpretation of the hydrogeologic framework and on limited information from wells and test borings.

The following section describes the analytical and numerical methods used to delineate the contributing areas then presents the methods used to estimate data required for model development; it also describes the delineation of contributing areas from model results.

Principles of Analytical and Numerical Methods

Areas that contribute ground water to wells can be delineated through (1) analytical methods based on equations that describe aquifers that have simple dimensions, or (2) numerical methods that use computer models to represent aquifers that have irregular and complicated dimensions. Analytical methods, such as the Wellhead Protection Area (WHPA) model (U. S. Environmental Protection Agency, 1990), are generally easy to apply and are useful for preliminary analyses of factors that control the size of contributing areas, such as the pumping rate and aquifer transmissivity. These methods apply the drawdown induced by pumping and the regional hydraulic-head distribution to delineate the area contributing recharge to the well through a particle-tracking technique. Analytical methods do not accurately represent aquifer systems with three-dimensional flow and irregular boundaries, however, and are not appropriate for delineating contributing areas to wells where the hydrologic boundaries that control the rate and direction of ground-water flow are unknown (Morrisey, 1987).

Numerical methods can be used to represent conditions that are not easily incorporated in analytical methods, such as flow to wells from streams, flow to wells that do not penetrate the full aquifer thickness, and flow in multilayer aquifer systems. Three-dimensional models of hypothetical aquifer systems can be constructed to incorporate several factors that affect the size and shape of areas contributing recharge to wells (Morrissey, 1987; Reilly and Pollock, 1993). The construction and calibration of such models are generally costly, however, and require extensive information from subsurface investigations.

Numerical methods also can be applied to conceptual models that represent hypothesized aquifer conditions to delineate ground-water contributing areas where extensive information is unavailable, but the resulting delineation is less certain than would be obtained with a calibrated numerical model. The conceptual numerical model can be used to represent conditions that can be easily measured, however, such as the topography and streambed elevations, or factors previously estimated for similar aquifer settings, such as recharge from precipitation. A preliminary calibration can be made from reported pumping rates, measured ground-water levels, and estimates of transmissivity or other aquifer properties.

Uncertainty in the location of the area contributing recharge to a well can be illustrated by comparing the areas delineated by conceptual models that represent alternative interpretations of aquifer boundaries and properties. Differences among these interpretations reflect differing estimates of aquifer properties, recharge rate, aquifer extent, or type of boundaries. Selection of the largest contributing area for an assumed set of possible aquifer conditions can allow for the uncertainty in aquifer properties and boundaries in the delineation of the area contributing recharge to a well. The analysis can also indicate what additional information is needed to determine which models best represent conditions in the aquifer; this additional information can then be used to improve the model and reduce uncertainty in the interpretation of model results.

Determination of Aquifer Characteristics

Aquifer characteristics that were measured or estimated in this study were (1) depth to bedrock, (2) aquifer transmissivity, (3) rate of recharge, (4) ground-water levels, and (5) ground-water discharge to streams.

Depth to bedrock -- Seismic-refraction surveys were conducted in areas of insufficient data (Holland and Collins) to supplement well data for use in mapping the bedrock-surface altitude. These surveys yielded the depth to bedrock that underlies the sand and gravel aquifers along a continuous transect. Seismic-refraction techniques used in this study are described by Haeni (1988). Explosives were used as an energy source because the bedrock surface is relatively deep in both study areas, typically more than 100 ft below land surface. A series of 12 geophones were spaced 50 and 100 ft apart on the ground, and arrival times of compressional waves generated by explosives were recorded and plotted as a function of source-to-geophone distances. The distribution of arrival times was analyzed (for a two- or three-layer system) to calculate depths to water and bedrock through a computer program discussed in Scott and others (1972).

Aquifer transmissivity -- This term was estimated from the specific capacities (*Q/s*) of municipal wells by the following relation (Todd, 1980, eq. 4.70):

$$Q/s = \left[\frac{2.3}{4\pi T}\log 2.25\frac{Tt}{r_w^{\ 2}S}\right]^{-1} \qquad (1)$$

where Q is well discharge (L^3T^{-1}),
s is drawdown (L),
T is transmissivity (L^2T^{-1}),
t is time of pumping (T),
r_w is well radius (L), and
S is storativity (dimensionless).

Rates of recharge -- Recharge rates were estimated from mean annual runoff values reported for western New York in Lyford and Cohen (1988, fig. 3) and rates of upland recharge reported in Morrissey and others (1988). Upland recharge rates in Morrissey and others (1988) were developed for unconfined glacial aquifers surrounded by till-covered bedrock uplands. These rates were assumed to represent the maximum recharge rate for the confined glacial aquifers in this study because hydraulic connections with till-covered bedrock uplands are limited by the lower permeability of fine-grained sediments in confining or semiconfining layers.

Ground-water levels -- Hydraulic heads were obtained from records of ground-water levels in wells and test borings.

Ground-water discharge to streams -- Streambed surveys and streamflow measurements were conducted along streams in some areas to identify possible ground-water discharge areas and to measure rates of discharge in selected areas.

Development of Ground-Water Flow Models

Ground-water flow models representing the aquifer geometry and estimates of aquifer properties of each aquifer were developed using the computer program MODFLOW (MacDonald and Harbaugh, 1988). The resulting hydraulic-head distributions were used to determine the direction and rates of ground-water flow and to calculate a water budget for each aquifer. Values representing aquifer properties and model boundaries were adjusted until the models could approximate field measurements of hydraulic head and ground-water flow.

A sufficient number of water-level measurements in the Alden area were available for use of nonlinear regression techniques to estimate values using the computer program MODFLOWP (Hill, 1992). This program computes values that minimize differences between observed and calculated values of hydraulic head and ground-water-flow rates. Values of hydraulic properties for the Holland and Collins models were estimated from specific capacities of municipal wells, or specified from the literature.

In all these models, underflow to or from areas beyond aquifer boundaries was represented by head-dependent-flow boundaries through which the rate of flow was proportional to (1) the difference in hydraulic head between the aquifer and the boundary, and (2) a conductance term representing the rate of flow from the boundary to the aquifer (eq. 2). Heads higher than the potentiometric surface of the aquifer were specified along boundaries with upgradient areas that provide a source of recharge, and heads lower than the potentiometric surface were specified along boundaries adjacent to downgradient areas that serve as regions of ground-water discharge. Head estimates were based on the altitude of perennial streams and wetland areas. The conductance term, *C*, required for head-dependent-flow boundaries is defined as:

$$C = \frac{KA}{l}, \qquad (2)$$

where K is hydraulic conductivity of the boundary material (LT^{-1}),
A is cross-sectional flow area along boundary (L^2), and
l is spacing between model cells at boundary (L).

Boundaries between aquifers and streams were represented either as drains, where leakage to the stream is proportional to (1) the difference between head in the aquifer and head in the drain, and (2) the drain conductance, or as streams with streamflow routing (Prudic, 1989), where leakage to or from streams is proportional to (1) the difference between head in the aquifer and head in the stream, and (2) the streambed conductance. The drain or streambed conductance for each model cell containing a stream was computed from equation 2 with K equal to the vertical hydraulic conductivity of the streambed, l the thickness of the streambed, and A the area of the stream reach within the model cell.

Hydraulic heads and flow rates computed by the models were then used to generate ground-water flow paths through the modeled areas with the particle-tracking routine MODPATH developed by Pollock (1989). Source areas for discharge boundaries represented in the models were delineated by a forward-tracking technique that traces particles from each model cell to a discharge boundary. The contributing areas to municipal wells were then delineated by plotting the locations of cells from which particles move to the boundary representing the well. The total rates of recharge to each contributing area were obtained as the sum of contributions from each model cell along model boundaries within the contributing area.

The term "contributing area" as used in this report refers to the areal projection at land surface of the volumetric portion of the confined aquifer from which water is diverted to a pumped well, it is also referred to as the "zone of contribution" by Morissey (1987). For an unconfined aquifer, the term "contributing area" has been defined as the area from which water entering the ground-water system at the water table flows to (and discharges from) a pumped well (Reilly and Pollock, 1993). Therefore, the vertical leakage of precipitation to the water table within the contributing area is the major source of recharge to the pumped well. For the confined aquifers in this study, however, vertical leakage within the contributing areas is limited by the low permeability of confining or semiconfining layers, so that unchanneled runoff and underflow from the adjoining bedrock uplands also is a major source of recharge to the pumped wells. Upland areas outside the modeled area that contribute recharge to municipal wells were delineated after an inspection of topographic maps to identify the watersheds that drain to the model boundary within the contributing area. The volume of recharge from upland areas was then computed as the sum of flow through the model boundary.

DELINEATION OF AREAS CONTRIBUTING RECHARGE TO MUNICIPAL WELLS

Ground-water flow models were developed for the three aquifers that provide drinking water for the Village of Alden and the Towns of Collins and Holland. The Alden area aquifer model was calibrated to measured ground-water levels and ground-water discharges to Spring Creek through MODFLOWP (Hill, 1992); the other two models were not calibrated because field data were insufficient, but were adjusted to represent assumed directions of ground-water flow. The areas contributing recharge to municipal wells in these three areas under average, steady-state conditions were delineated from model results through particle-tracking techniques.

Alden Study Area

The aquifer serving the Village of Alden, in eastern Erie County, is bordered by Ellicott and Cayuga Creeks (fig. 2). The Village's water supply consists of four wells that supply about 300,000 gal/d (40,000 ft^3/d) to 2,450 people. Three drilled wells (nos. 2, 3, and 4 in fig. 2) are finished at the top of bedrock, and well depths range from 36 to 45 ft below land surface. Well no. 1 was dug in 1900 to a depth of 14 ft. The wells are screened in a confined (or semiconfined) sand and gravel aquifer 15 to 20 ft thick that overlies the bedrock. Water from the Alden wells generally meets the State drinking-water standards but is treated to remove iron and hydrogen sulfide (Bruce Sitzman, Village of Alden, oral commun., 1995).

Hydrogeologic setting

The confined aquifer screened by municipal wells lies just north of the Marilla Moraine (fig. 1), a till deposit formed by a temporary ice stand about 12,500 years ago (Fullerton, 1980). The sand and gravel that forms the confined aquifer could have been deposited by meltwaters at the ice front as it passed through the area before the Marilla Moraine was formed, or by northward-flowing streams draining into the proglacial lake that formed between upland areas to the south and the receding ice front to the north. Some time after the deposition of the sand and gravel, a readvance of the

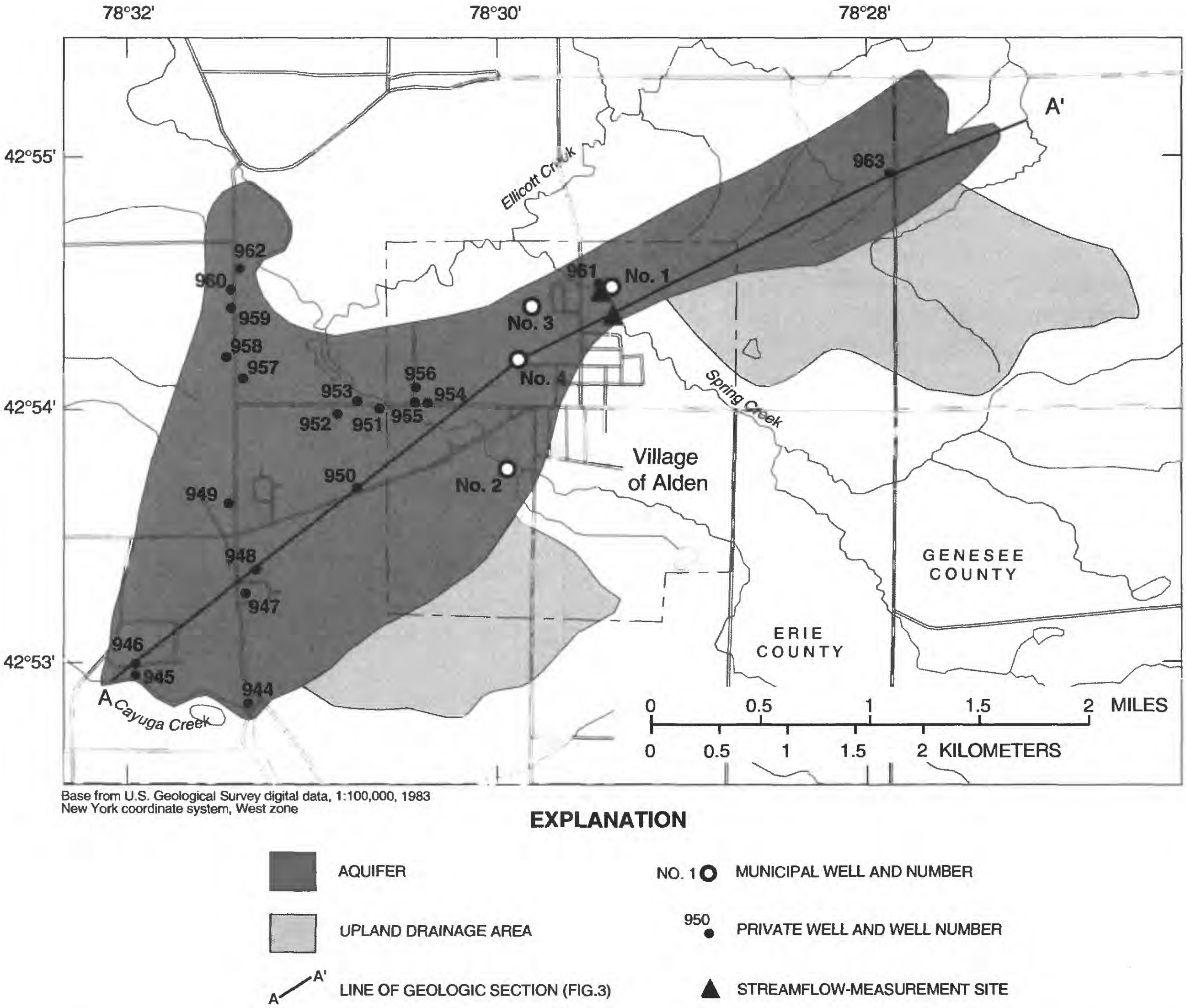


Figure 2. Map showing geographic features and aquifer boundary of the Alden area, Erie County, N.Y. (General location is shown in fig. 1)

ice front formed the Marilla Moraine and deposited a layer of fine-grained till probably derived largely from the sediments of the proglacial lake over the sand and gravel. After the ice receded northward again, the Alden area was covered by relatively long-lived proglacial Lake Warren (Calkin, 1970). Beach ridges of sand and gravel were deposited at the shore of Lake Warren, where they form an unconfined aquifer (fig. 1) overlying the till.

Relatively little information on the stratigraphy of glacial deposits in the Alden area is available; therefore, the thickness of the confined aquifer, the unconfined aquifer, and the semiconfining layer between them is uncertain in many areas. A generalized section by Miller and Staubitz (1985) (fig. 3) was based on two complete stratigraphic logs and notes from a local well driller. The confined aquifer ranges from 5 to 20 ft thick and is overlain up to 20 ft of less permeable material. The unconfined aquifer is 10 to 20 ft thick, but few wells are completed in this aquifer because the saturated thickness is generally 5 ft or less. The extent of the unconfined aquifer can be readily identified from land-surface topography and soils maps. The extent of the confined aquifer is uncertain, but it is bordered to the south by till deposits of the Marilla Moraine. If the confined aquifer was deposited by upland streams, the aquifer probably does not extend far to the north and probably has a shape of an alluvial fan. Alternatively, if the material was deposited by meltwater from the ice, the aquifer could extend further north and could be irregularly shaped. The lateral extent of the semiconfining layer separating the aquifers is uncertain also.

The confined aquifer is recharged by precipitation that falls directly on the overlying unconfined aquifer and percolates through the semiconfining layer, and by underflow and unchanneled runoff from upgradient upland areas to the south. Recharge from precipitation over the aquifer is controlled by the rate of vertical leakage through the semiconfining layer overlying the confined aquifer. The recharge rate is probably high in areas where the semiconfining layer is thin or absent, and lower where the confining layer is thick. Underflow from upgradient areas is controlled by the hydraulic conductivity of the Devonian shale bedrock that underlies the aquifer. Ground water in the bedrock that recharges the aquifer probably flows through the weathered upper 5 to 10 ft of shale.

Water discharges from the confined aquifer to wells, springs, and perennial streams that cross the aquifer, and to downgradient areas to the northwest. Streamflow measurements along a 1,000-ft reach of Spring Creek where the streambed intersects the confined aquifer (fig. 2) during low-flow conditions in November 1994 indicated a rate of ground-water-discharge to the creek of about 0.3 ft^3/s (25,000 ft^3/d). A streambed survey revealed sand and gravel in all streams bordering the unconfined aquifer, indicating that the stream channels penetrate the unconfined aquifer. Upward discharge from the confined aquifer to stream channels might occur as vertical leakage through the semiconfining layer in areas where the streambed overlies the top of the confined aquifer.

Model design

Ground-water flow was simulated as a two-layer system consisting of the confined sand and gravel aquifer (layer 1) and the underlying weathered bedrock (layer 2). Three steady-state simulations (mod-

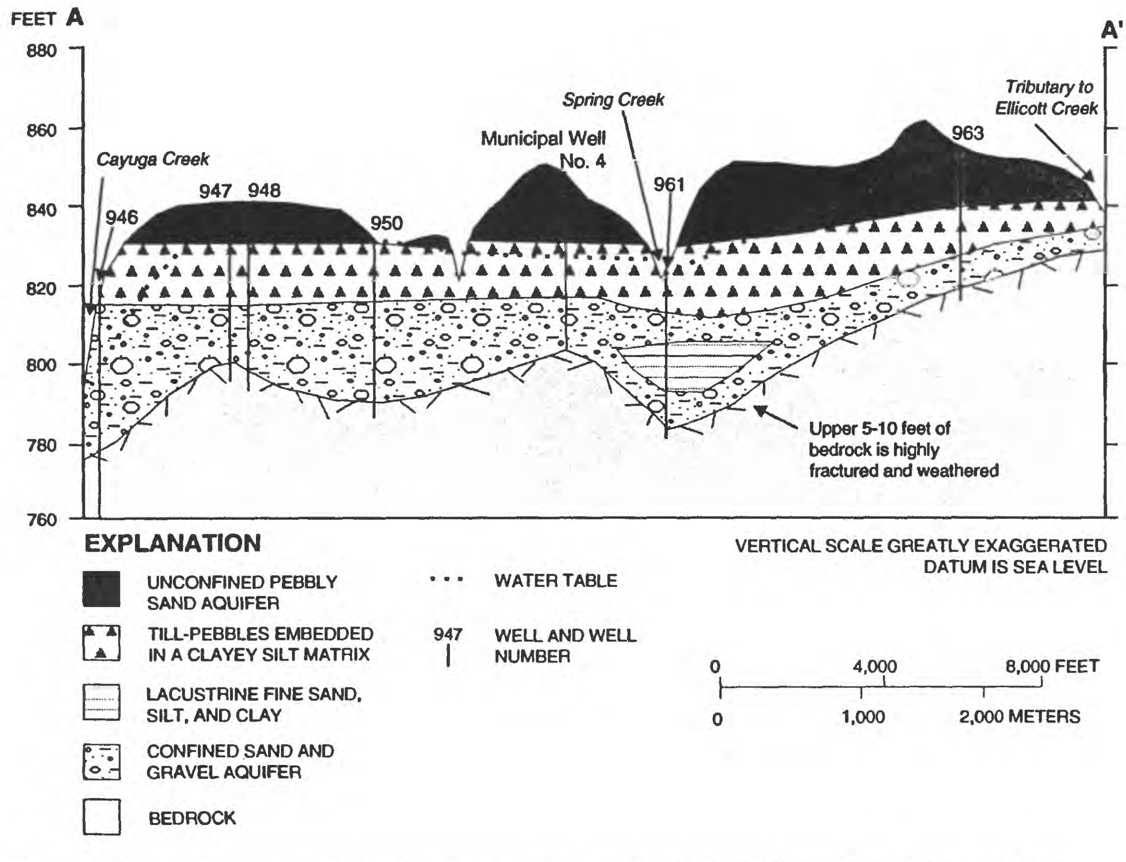


Figure 3. Generalized geologic section showing stratigraphy of glacial deposits in Alden area, Erie County, N.Y. (Trace of section is shown in Fig. 2) (modified from Miller and Staubitz, 1985)

els A, B, and C) were calibrated to ground-water levels measured in 18 wells during 1978-82 (fig 2) (Miller and Staubitz, 1985) and ground-water discharge to Spring Creek as determined by two streamflow measurements made in November 1994. Water levels in 14 of the 18 wells were measured in November 1982, a year in which the annual precipitation at Buffalo (fig. 1) was 41.1 in., which is about 2.5 in. greater than the average (National Oceanic and Atmospheric Administration, 1983). No long-term well hydrographs from the Alden area were available, but the 1978-82 measurements are assumed to represent average, steady-state conditions.

The lateral boundaries of the modeled area coincide with the boundaries of the unconfined aquifer. The modeled area encompasses 3.8 mi^2 and was represented by a uniformly spaced grid containing 64 rows and 131 columns. It contains 2,684 active cells, each of which represents 200 ft by 200 ft (about 1 acre). Discharge boundaries in layer 1 (confined aquifer) included wells, perennial streams, and springs (figs. 4, 5). The altitude of boundaries representing streams and springs was estimated from 1:24,000-scale topographic maps. Other lateral boundaries in layer 1 correspond to areas where the confined aquifer is bordered by poorly permeable materials, including till and lacustrine sediments, and are specified as no-flow boundaries. Layer 1 is recharged by downward leakage of precipitation from the unconfined aquifer through the semiconfining layer (fig. 4). Additional recharge from leakage from the village's water-distribution system (conveyance losses) is estimated to constitute 18 percent of the total amount pumped, or 7,000 ft^3/d (Bruce Sitzman, Village of Alden, oral commun., 1995). These losses are represented by increased recharge to layer 1 along the part of the southern boundary that lies within the village (fig. 5). Lateral boundaries in layer 2 (weathered bedrock) are specified as head-dependent flow to provide underflow to the modeled area from upgradient sources and from the modeled area to downgradient areas (figs. 4, 5). The distribution of hydraulic head along these boundaries in layer 2 was estimated from altitudes of ponds and streambeds of perennial streams. Streams crossing the modeled area were represented as drains.

Aquifer properties specified in the model (table 1) were assigned a fixed value from the literature or estimated by nonlinear regression. Transmissivity of the weathered shale bedrock was assigned a value of 25 ft^2/d, as calculated from a range of transmissivity values (1 to 500 ft^2/d) obtained from hydraulic tests of weathered shale in other studies in similar areas (Lapcevic and Novakowski, 1993; Dames and Moore, 1992). The vertical hydraulic conductivity of the weathered bedrock was assumed to determine the rate of vertical leakage between the weathered bedrock and the confined aquifer and was assigned a value of 1 ft/d, about one-half the horizontal hydraulic conductivity.

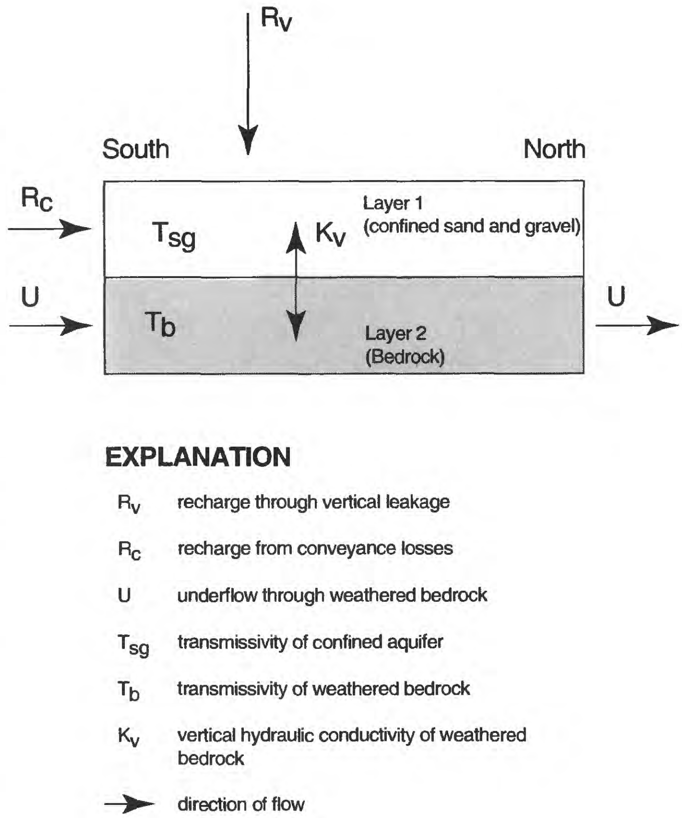


Figure 4. Diagram showing model representation of ground-water flow through glacial sediments and weathered bedrock in models of Alden area, Erie County, N.Y.

Values for five aquifer hydraulic properties were estimated through nonlinear regression for each of the three models. In model A, the transmissivity of the confined aquifer was estimated to be 2,100 ft^2/d, about one-half the transmissivity values estimated from equation 1 and the specific capacities of Alden wells no. 3 and 4 (14 and 18 [gal/min]/ft, respectively). Recharge from downward leakage through the semiconfining layer was estimated to be 5.6 x 10^{-3} ft/d or 24 in/yr, greater than the estimate

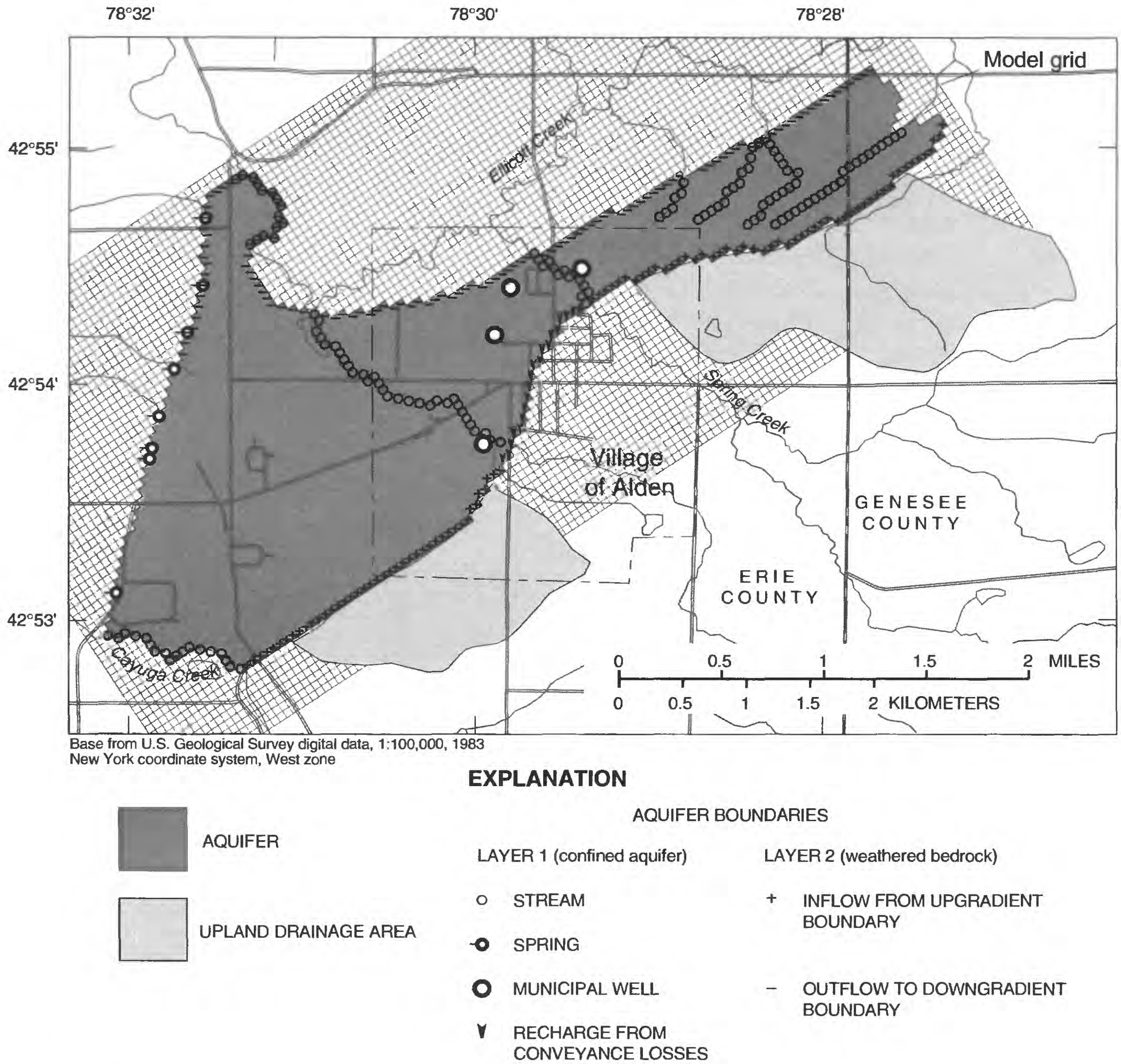


Figure 5. Map showing boundary conditions in the models of Alden area, Erie County, N.Y.

of 18 in/yr obtained from average annual base flow (Lyford and Cohen, 1988, fig. 3). Higher and lower values of transmissivity and recharge were considered in models B and C as discussed below. The conductance values for model boundaries corresponding to streams and springs represent poorly permeable materials that could impede the discharge of ground water to these boundaries. Conductance values estimated by nonlinear regression for boundaries representing small and large streams were nearly equal to or greater than the value for conductance between adjacent model cells and had little effect on model results. Conductance values estimated for boundaries representing springs were much smaller than conductance values between model cells and decreased the rate of ground-water discharge through these boundaries.

Simulated ground-water flow

The hydraulic heads computed by model A (intermediate transmissivity and recharge rate) for average steady-state conditions in the confined aquifer indicate that ground water flows northward and westward from the southern upland boundary toward streams and springs (fig. 6). A ground-water divide in the western

Table 1. Values assigned to aquifer properties in ground water flow models in Alden area

[ft/d, feet per day; ft^2/d, feet squared per day; ft^3/d, cubic feet per day]

Model Variables		Values estimated through nonlinear regression*		
Aquifer property	Value assigned from published literature	Model A	Model B	Model C
Transmissivity (ft^2/d)				
confined aquifer		2,100	4,000*	860
weathered bedrock	25			
Vertical hydraulic conductivity (ft/d)				
confined aquifer	200			
weathered bedrock	5			
Recharge (ft/d)				
vertical leakage		5.6×10^{-3}	6.6×10^{-3}	2.3×10^{-3}**
conveyance losses	10^{-2}			
Conductance (ft^2/d)				
springs		66	23	15
small streams		14,000	2900	3300
large streams		35,000	27,000	26,000
Model results				
Standard error (ft)		7.3	8.7	7.6
Computed discharge to Spring Creek (ft^3/d)		59,000	78,000	17,000

*A. intermediate transmissivity and recharge rate
B. high transmissivity and recharge rate
C. low transmissivity and recharge rate

** fixed value in nonlinear regression

part of the aquifer separates ground water flowing southwestward to Cayuga Creek from water flowing northward to Ellicott Creek and its adjoining tributary. Ground water in the eastern part of the aquifer generally flows westward toward Spring Creek. Differences between hydraulic heads computed by the model and those measured at 18 wells screened within the aquifer are generally less than 6 ft, although the computed heads at three wells were 12 to 14 ft higher-- these were wells 945 and 946 near Cayuga Creek and well 963, near the eastern aquifer boundary (fig. 6).

Rates of ground-water flow to and from the modeled area were calculated as the sum of contributions to or from each model cell within recharge areas and along discharge boundaries (table 2). Results indicate that vertical leakage through the semiconfining layer accounted for nearly all water entering the confined aquifer. About 85 percent of the ground water that exited the modeled area discharged to streams, two of which (Cayuga Creek and the middle tributary to Ellicott Creek) accounted for 70 percent of the total ground-water discharge; pumping from the four municipal wells accounted for 6 percent of the total discharge, and discharge to Spring Creek accounted for the remaining 10 percent of the total ground-water discharge (59,000 ft^3/d). This is close to the value of 62,200 ft^3/d obtained by applying the measured discharge (0.3 ft^3/s per 1,000 ft of channel) to the 2,400-ft reach of Spring Creek represented in the model, although some of the assumed discharge could be derived from the overlying unconfined aquifer.

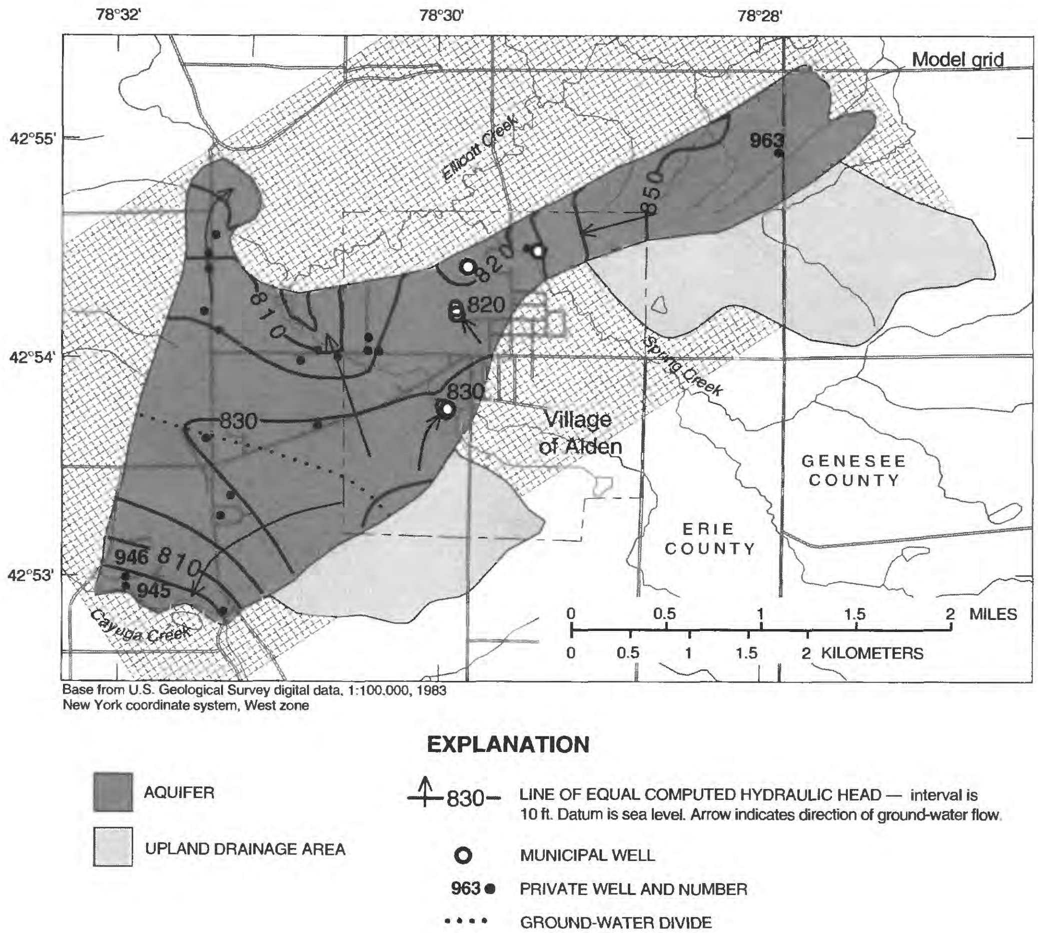


Figure 6. Map showing steady-state distribution of hydraulic head in Alden area, Erie County, N.Y. as computed by model A

Whereas water in most of the western part of the aquifer discharges primarily to Cayuga Creek and the middle tributary to Ellicott Creek, water in the eastern part discharges to Spring Creek and smaller tributaries to Ellicott Creek (fig. 7A). The areas contributing recharge to the four municipal wells range in size from 0.05 to 0.07 mi^2 (29 to 45 acres). Ground water that enters the confined aquifer (assumed effective porosity = 0.2) at the southern boundary takes 1 to 2 years to reach the municipal wells. Less than 7 percent of the water pumped from municipal wells originates as underflow and runoff from upland areas south of the aquifer, except for well no. 2, which derives 28 percent from upland areas (table 3).

Reliability of estimates

The principal factor that limits the reliability of model results is uncertainty as to the rate of ground-water flow through the confined aquifer. Nonlinear-regression estimates of aquifer transmissivity and vertical leakage through the semiconfining layer are sensitive to the rate of ground-water discharge to streams. The estimated transmissivity value was less than the minimum value calculated from specific-capacity data, and the estimated vertical-leakage rate was greater than the recharge rate calculated from average annual base flow. Therefore, two alternative models (B and C) were calibrated through nonlinear regression to determine the effect of these factors on the size of the

Table 2. Water budget for confined aquifer at Alden under average steady-state conditions in model A (intermediate transmissivity and recharge rates)

[Flow rates are in thousands of cubic feet per day]

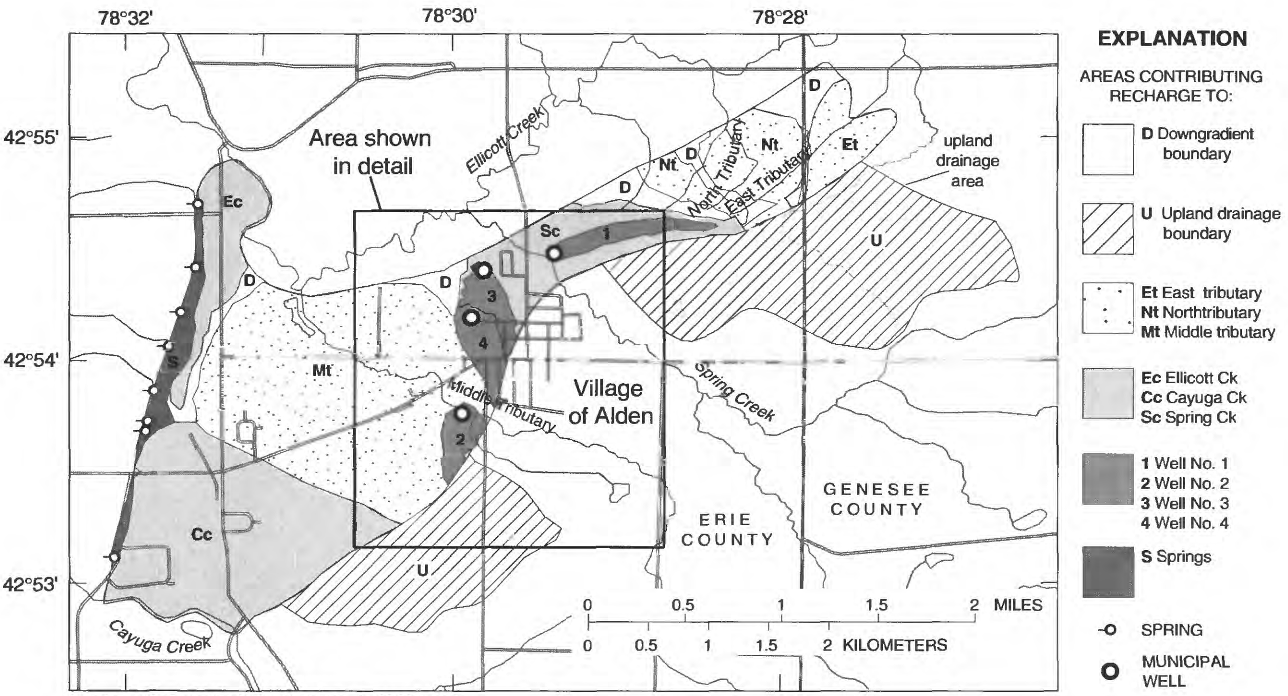
Inflow			Discharge		
Source	Rate	Percentage of total	Location	Rate	Percentage of total
Vertical leakage	593.0	97	Alden wells	38.7	6
Conveyance loss	7.0	1	Underflow	40.4	7
Underflow	13.9	2	Springs	7.6	1
			Spring Creek	58.6	10
			Ellicott Creek	38.4	6
			Cayuga Creek	152.3	25
			Other tributaries	277.9	45
Total	613.9	100		613.9	100

areas contributing recharge to municipal wells. Both gave larger differences between observed and predicted heads and flows than model A and, therefore, are not more plausible. The results illustrate the model's sensitivity to changes in aquifer transmissivity and vertical leakage, however.

In model B, the aquifer transmissivity was fixed at 4,000 ft^2/d--the average value obtained from specific-capacity data for municipal wells no. 3 and no. 4 and nearly twice the value used in model A. The regression estimate of vertical leakage from the unconfined aquifer was increased about 20 percent from 5.6×10^{-3} ft/d (24 in/yr) in model A to 6.6×10^{-3} ft/d (29 in/yr)-- a value that is slightly less than the average recorded precipitation rate of 34 in/yr and appears to be too large (table 1). The average error in predicted heads increased by 40 percent, and the predicted flow to Spring Creek increased more than 30 percent-- from 59,000 ft^3/d in model A to 78,000 ft^3/d in model B. The resulting contributing areas to the municipal wells were slightly smaller than those delineated with model A, except the area for well no. 1, which increased 24 percent from 0.045 to 0.056 mi^2 (29 to 36 acres).

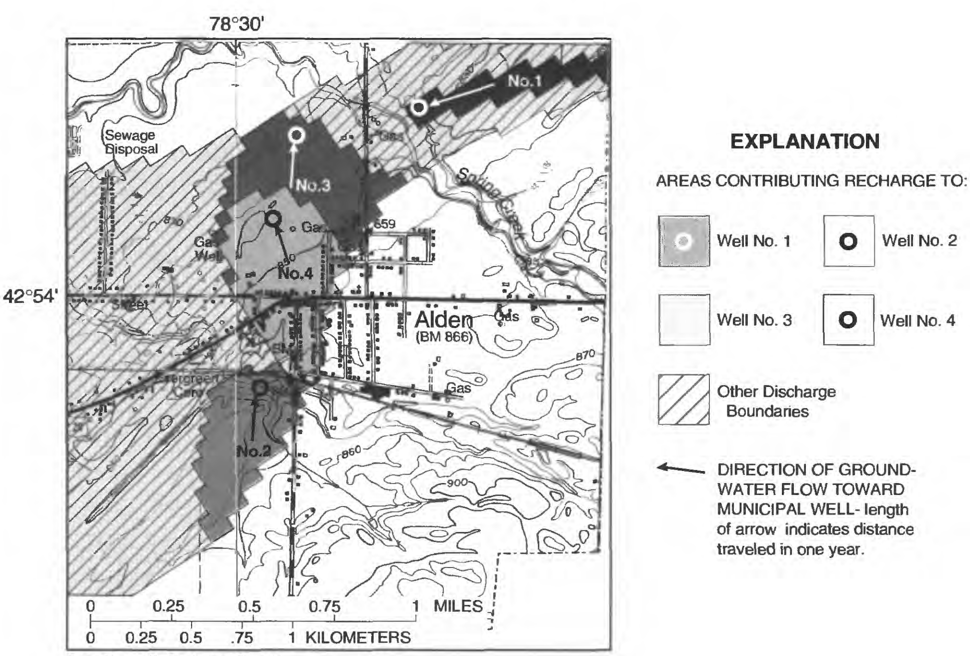
Model C represented an extensive confining layer with the vertical-leakage rate fixed at 2.3×10^{-3} ft/d (10 in/yr), which is half the value used in model A and is equal to the median hydraulic-conductivity value obtained for fine-grained glacial sediments in the Niagara Falls area, 30 mi to the northwest (fig. 1) (E. C. Jordan Co., 1985; Conestoga-Rovers and Associates and Woodward-Clyde Consultants, 1990). This is reasonable because, under a unit hydraulic gradient (1:1), the rate of vertical leakage through an extensive confining layer of saturated fine-grained sediments is equal to the vertical hydraulic conductivity. From this assumption, (1) the transmissivity value estimated by regression decreased 59 percent, from 2,100 ft^2/d in model A to 860 ft^2/d, (2) the average error in predicted heads increased by 10 percent, and (3) the predicted discharge to Spring Creek decreased 75 percent to 17,000 ft^3/d, about one-quarter the estimated inflow along the 2,400-ft reach (table 1). In model C, the amount of water pumped from the municipal wells accounted for 14 percent of the total ground-water discharge, over twice the amount computed with model A, and the sizes of the contributing areas to the municipal wells increased by 40 to 90 percent (fig. 7B). The volume of pumped water that was derived from upgradient areas beyond the boundary of the confined aquifer was 10 to 40 percent of the total volume pumped, except for well no. 1, which derived essentially all pumped water from within the boundary of the confined sand and gravel aquifer (table 3).

Results obtained from models A and C suggest that, whereas large aquifer-transmissivity values have little effect on the size of contributing areas, small transmissivity values significantly increase the size of the contributing area. Restricted vertical leakage through the semiconfining layer also would increase the size of the contributing area and require a smaller aquifer transmissivity value in model simulations to reproduce the measured distribution of hydraulic head. A vertical leakage rate smaller than the value obtained with model A would be expected if the semiconfining layer is more extensive or thicker than assumed, or if recharge to the unconfined aquifer is less than the estimated leakage rate. The large contributing areas indicated by model C (fig. 7B) would be a useful basis for



Base from U.S. Geological Survey digital data, 1:100,000, 1983
New York coordinate system, West zone

A. CONTRIBUTING AREAS AND HYDROLOGIC BOUNDARIES



Base from U.S. Geological Survey
Clarence, N.Y., 1:24,000, 1965
Corfu, N.Y., 1:24,000, 1950

B. DETAIL OF ALDEN AREA

Figure 7. Maps showing areas contributing recharge in Alden area, Erie County, N.Y. to: A. Areas contributing to all discharge boundaries in model A (intermediate transmissivity and recharge). B. Areas contributing to municipal wells in model C (low transmissivity and recharge)

Table 3. Sizes of areas contributing recharge to municipal wells, and estimated recharge from upland runoff, in models of Alden area

[Well locations are shown in fig. 7.]

Well	Pumping rate (thousands of cubic feet per day)	Model A Area (mi^2)	Model A Percentage of pumpage derived from upland runoff	Model B Area (mi^2)	Model B Percentage of pumpage derived from upland runoff	Model C Area (mi^2)	Model C Percentage derived from upland runoff
No. 1	3.3	.045	0	.056	0	.063	2
No. 2	12.1	.056	28	.044	33	.094	40
No. 3	10.6	.061	5	.044	5	.11	9
No. 4	12.6	.07	7	.061	9	.13	13

wellhead-protection measures that require a conservative estimate of contributing-area size.

The uncertainty in aquifer conditions and in size and shape of contributing areas delineated from model simulations can be decreased if certain information on the rate of ground-water flow through the aquifer were refined. For example: (1) measurements of streamflow during periods of low flow to estimate ground-water discharge to streams, especially for Cayuga Creek and the large tributary to Ellicott Creek, would allow improved estimates of aquifer transmissivity and the rate of vertical leakage through the confining layer; (2) test drilling would provide data on the extent and thickness of the semiconfining layer; and (3) installation of monitoring wells in weathered bedrock in upgradient areas south of the aquifer would provide information on the rate of underflow to the aquifer from areas beyond the boundary of the confined sand and gravel aquifer.

Collins Study Area

The Town of Collins is in the southern part of Erie County (fig. 1). Two municipal wells supply water to three water districts in the town, but only the westernmost municipal well, locally known as well no. 1 on Taylor Hollow Road (USGS well no. 932, fig. 8), was investigated during this study. Well no. 1 is 1.8 mi northwest of the Hamlet of Collins (fig. 8). The well is 129 ft deep, has an 18-in.-diameter casing, and is screened from 119 to 129 ft below land surface in a confined sand and gravel aquifer in the Clear Creek valley. It is typically pumped at a rate of 365 gal/min for 2 to 3 hours each day and supplies from 70,000 to 85,000 gal/d to 1,100 people (Daniel Stroud, Collins Water Department, oral commun., November 28, 1994). The chemical quality of water from the Collins well generally meets the State drinking-water standards (Daniel Stroud, written commun., 1995).

Hydrogeologic Setting

The study area is characterized by buried and partly buried valleys filled with glacial drift deposited during Wisconsinan time 13,000 to 14,000 years ago, mostly during the last deglaciation of western New York (Muller and Calkin, 1993). Some older, pre-Wisconsinan glacial deposits overlying bedrock are exposed in bluffs along the Clear Creek channel near Collins (Calkin and others, 1982) and may also lie in deep zones of glacial drift in buried valleys. Multiple advances and retreats of the ice in the study area have resulted in deposition of a complex and heterogeneous mix of till, and of glaciofluvial and glaciolacustrine sediments in the valleys.

The surficial deposits in the study area were mapped by Muller (1977), but little is known about their stratigraphy in deep zones of the valleys. Generalized geologic sections showing the stratigraphy in the study area (fig. 9) are based on several well records that indicate that the upper parts of the valley-fill deposits consist of 10 to 90 ft of alluvial and deltaic sand and gravel, underlain by 50-to-350-ft-thick sequence of mostly fine-grained sediments (till and glaciolacustrine fine sand, silt, and clay) that, in turn, is underlain by a basal sand and gravel 10 to 80 ft thick. This basal sand and gravel deposit probably consists of subglacial-glaciofluvial deposits such as eskers and esker fans deposited by subglacial meltwater streams and overlies Devonian shale, which forms the bottom of the aquifer. Results of seismic-reflection surveys in Finger Lake valleys of Central New York by Mullins and others (1991) suggest that meltwaters that flowed beneath the glacier were "feeders" of coarse-grained sediments to the moraines that were deposited at the ice front.

The surficial alluvial and deltaic deposits, where present, form an unconfined aquifer in the Collins study area, and the till and glaciolacustrine deposits form a confining unit that overlies the basal sand and gravel deposits, which form a confined aquifer. Municipal well no. 1 taps the confined sand and gravel aquifer in the buried-tributary valley, now occupied by Clear Creek, that is close to the junction with the ancestral Allegheny River buried valley now occupied by Cattaraugus Creek (fig. 10). During early glaciation of southwestern New York, the advancing ice sheet blocked the ancestral Allegheny River, which had flowed northwestward in a 1-mi-wide valley that extended from the southwestern part of New York to Gowanda and eventually to the Lake Erie basin (Muller and Calkin, 1993). The Allegheny River was diverted southwestward by the ice, and now flows into the Ohio River basin. Subsequent deposition filled the buried ancestral Allegheny River Valley with more than 500 ft of sediment in the study area.

Results of two seismic-refraction surveys in the vicinity of well no. 1 (fig. 8) indicate that glacial drift in the buried-tributary valley ranges in thickness from 100 to 200 ft and that bedrock is within several feet of

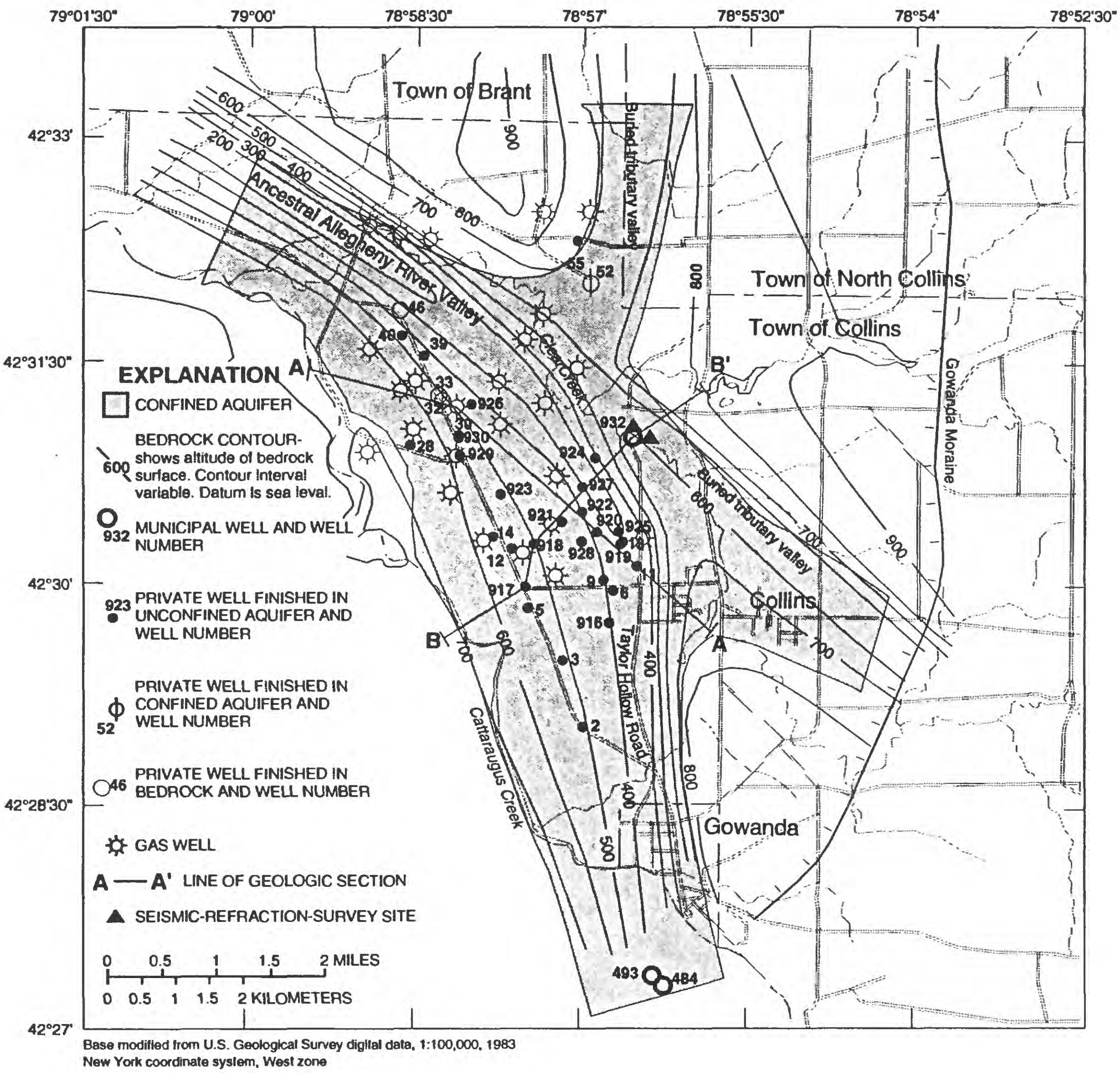


Figure 8. Map showing geographic features, aquifer boundary, and altitude of bedrock surface in the Collins area

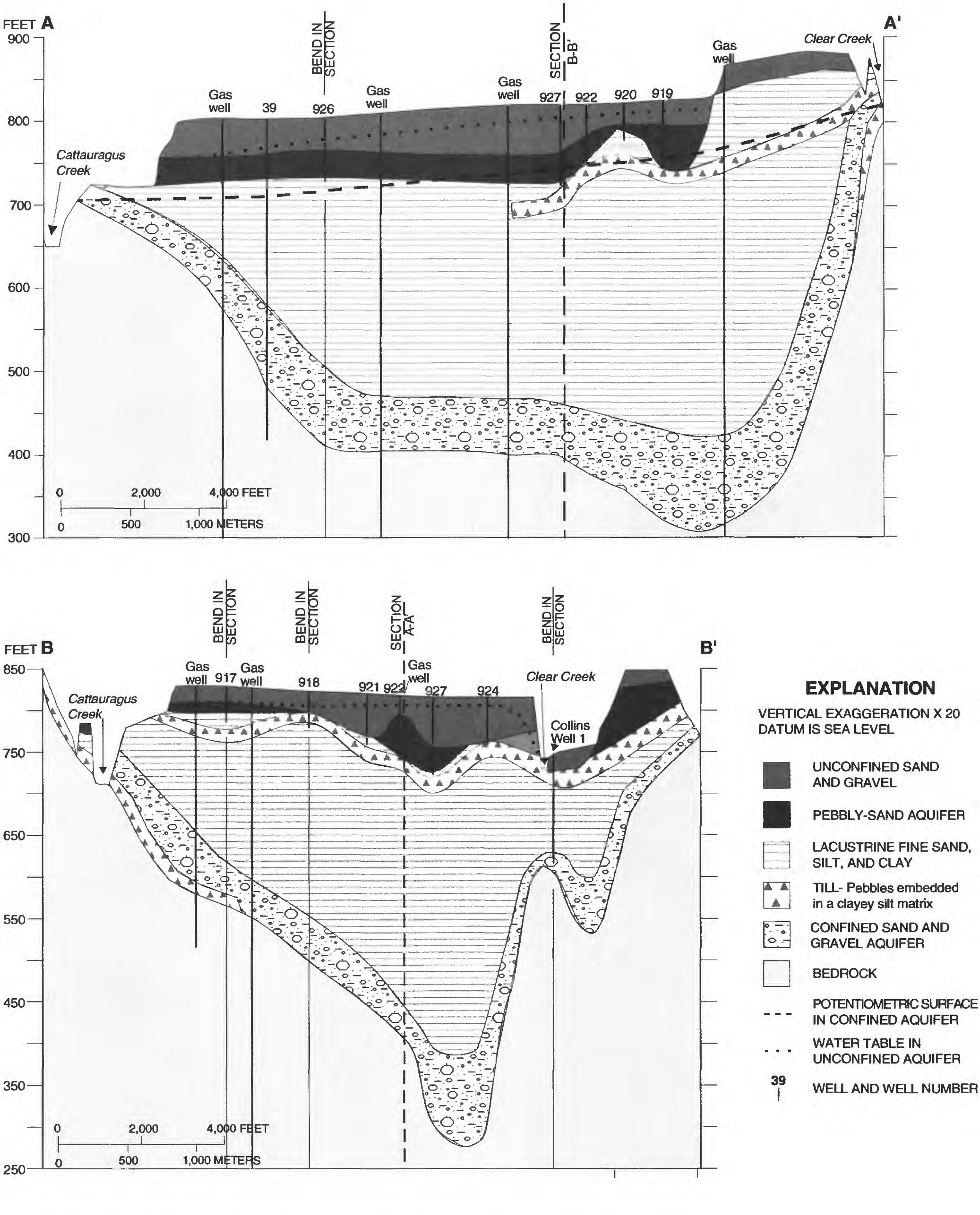


Figure 9. Generalized geologic sections A-A' and B-B' showing stratigraphy of glacial deposits in Collins area. (Traces of sections are shown in Fig. 8)

the bottom of well no. 1, suggesting that the well taps the confined, basal sand and gravel aquifer. A few well records indicate that the aquifer is typically 50 ft thick along the middle of the valley and thins to 10 to 20 ft thick along the edges. Confined aquifers that overlie bedrock are common in valleys of western and central New York (Miller, 1988, and Mullins and others, 1991). Records of gas wells and water wells in the ancestral Allegheny River valley indicate as much as 550 ft of unconsolidated sediment in the valley and a confined basal sand and gravel aquifer at the bottom of the valley. In this study, the basal aquifers in both the ancestral Allegheny River valley and Clear Creek valley were assumed to be hydraulically connected.

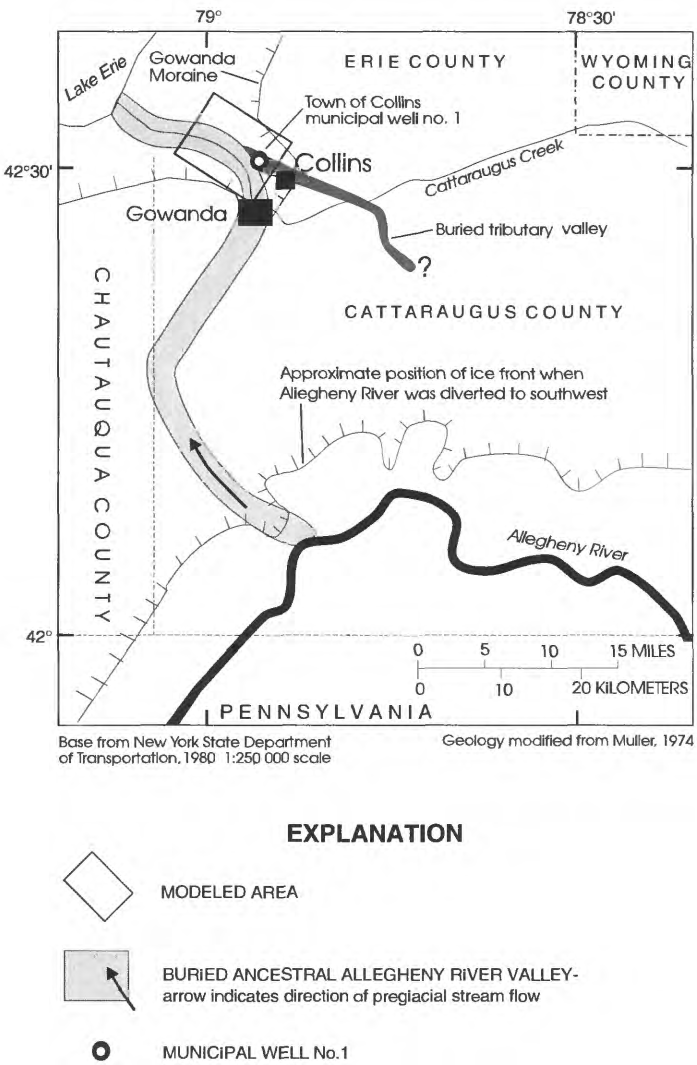


Figure 10. Map showing pertinent geologic features of Collins area, Erie County, N.Y. and location of modeled area (General location is shown in fig. 1)

Water pumped from well no. 1 could be derived from, or could affect water levels in the buried tributary valley occupied by Clear Creek and the much deeper buried ancestral Allegheny River valley because the well is near their junction (fig. 10); therefore, the modeled area includes parts of both buried valleys. These areas warrant investigation because ground water in the Clear Creek valley probably flows westward and discharges to Cattaraugus Creek in the ancestral Allegheny River valley, as does surface water.

The confined aquifer in the buried tributary valley now occupied by Clear Creek is recharged by (1) precipitation that infiltrates to the water table where the aquifer crops out at land surface at either the Gowanda moraine or the Lake Escarpment moraine along the east and south edges of the study area or both (figs. 1 and 8), (2) infiltration of runoff and lateral flow of ground water from unchanneled upland hillsides along the Gowanda moraine, where it is adjacent to the northeast side of the valley, (3) infiltration of streamflow from Clear Creek and tributary streams that flow over the unconfined aquifer where it crops out in some places along the edges of the valley, and (4) underflow from upgradient parts of the buried bedrock valleys to the north, east, and south. Water discharges from the confined aquifer (1) by flowing to pumped wells, (2) as underflow to the west in the ancestral Allegheny River valley, and (3) by seeping into the Cattaraugus Creek along the south edge of the valley.

The few exposures of glacial-drift deposits along the edges of the valley within the modeled area indicate that the sand and gravel deposits that are found at depth in the buried valley extend up the valley wall, where they are typically overlain by thin layers of till, sand and gravel, and glaciolacustrine sediments. These fine-grained deposits, where present, could restrict aquifer recharge from upland areas. Locally, the streams have removed the fine-grained deposits, leaving the basal sand and gravel aquifer exposed at land surface.

Model Design

Ground-water flow was simulated with a one-layer, numerical model representing average, steady-state conditions in the basal confined aquifer in the buried valleys. Simulation results can be interpreted only qualitatively because the ground-water-level and streamflow measurements were insufficient for model calibration. Nevertheless, the model illustrates important characteristics of the ground-water-flow system.

A rectangular finite-difference grid (fig. 11) was superimposed on a map of the study area to discretize the hydrogeologic conditions of the conceptual model. The grid represents an area of 12.1 mi^2 and contains 125 rows and 175 columns with uniform-sized cells, each representing 200 x 200 ft. The model contains 8,409 active cells.

The model represents the confined aquifer that overlies bedrock in the buried valleys. Several types of boundary conditions (fig. 11) were specified in the model to represent sources of recharge and areas of discharge. Natural hydrologic boundaries were used where possible, but arbitrary boundaries were used at certain locations to limit the size of the modeled area because the aquifer extends many miles beyond the study area. The arbitrary boundaries were placed far enough from the municipal well that they would have little effect on model results in this area.

Boundaries representing (1) the bottom of the aquifer (contact between the sand and gravel and the shale) and (2) top of the aquifer (contact between sand and gravel and overlying till and glaciolacustrine fine sand, silt, and clay) were specified as no flow because the rate of ground-water-flow through these boundaries would be negligible compared to the rate of flow through the aquifer. Most of the lateral boundaries representing the contact between the aquifer and bedrock-valley walls also were specified as no flow, except along the east boundary of the valley, where a head-dependent-flow boundary was used to represent flow from morainal deposits that are upgradient of the aquifer along the east edge of the valley (fig. 10). Head-dependent flow boundaries were also used for the arbitrary boundaries at the ends of valleys to simulate either inflow to, or outflow from, the modeled area from valleys that extend beyond the study area. Streams bordering the aquifer were represented as streams with head-dependant leakage (Prudic, 1989); streambed conductance was computed from equation 2 and an assumed vertical hydraulic conductivity (K) of 1.0 ft/d, a streambed thickness (l) of 1.5 ft, and an appropriate streambed area (A).

The heads used for the head-dependent boundaries were determined as follows:

(1) inflow from the ancestral Allegheny River valley at Gowanda-- 757 ft, the measured water level in the municipal wells for the Village at Gowanda in the southern part of the study area,

(2) inflow from the buried-tributary valley now occupied by Clear Creek-- 880 ft, the elevation of springs discharging from the backside of the Gowanda Moraine,

(3) unnamed buried tributary valley in the northern part of the study area-- 785 ft, the estimated elevation of the headwater stream in the northern part of that valley, and

(4) outflow through the Cattaraugus Creek valley-- 630 ft, the streambed elevation in Cattaraugus Creek where it exits the modeled area.

The hydraulic conductivity of the aquifer was estimated from records of pumping and drawdown at municipal well no. 1. Specific capacity, calculated from pumping records from the Town of Collins Water Department, was 9.4 (gal/min)/ft on October 26, 1992. Transmissivity (T), calculated from equation 1, was 2,060 ft^2/d, and hydraulic conductivity (K), calculated from an assumed aquifer thickness (b) of 20 ft at the municipal well site, was about 100 ft/d ($K = T/b$).

Town of Collins Municipal well no. 1 is pumped at a rate of 365 gal/min or 0.53 Mgal/d (70,300 ft^3/d). The well is pumped for only 2 to 3 hours a day, however, to meet the average daily requirement of 0.08 Mgal/d (10,700 ft^3/d). A steady-state pumping rate of 365 gal/min was specified in model simulations to represent maximum ground-water withdrawals for future needs.

Simulated Ground-Water Flow

Two simulations (models A and B) were run to represent different rates of recharge to the confined aquifer from the uplands bordering the east side of the aquifer. The rate of recharge specified along the east side of the valley affects the size and shape of the contributing area to municipal well no. 1 because of its proximity to the well. Recharge could enter the confined aquifer along the east side of the valley along two paths: (1) as precipitation infiltrating to the water table through the relatively permeable sediments that blanket the uplands, or (2) as runoff from unchanneled hillsides that seep into the ground along the edges of the valley, where the confined aquifer is close to land surface. A large conductance value (100 ft^2/d), equivalent to the hydraulic conductivity of a medium sand, was specified for the head-dependent-flow boundary in model A to represent a high rate of recharge from the uplands, and a much smaller conductance value (20 ft^2/d), equivalent to the hydraulic conductivity of a fine sand, was specified for model B to examine the effect of less recharge from the uplands.

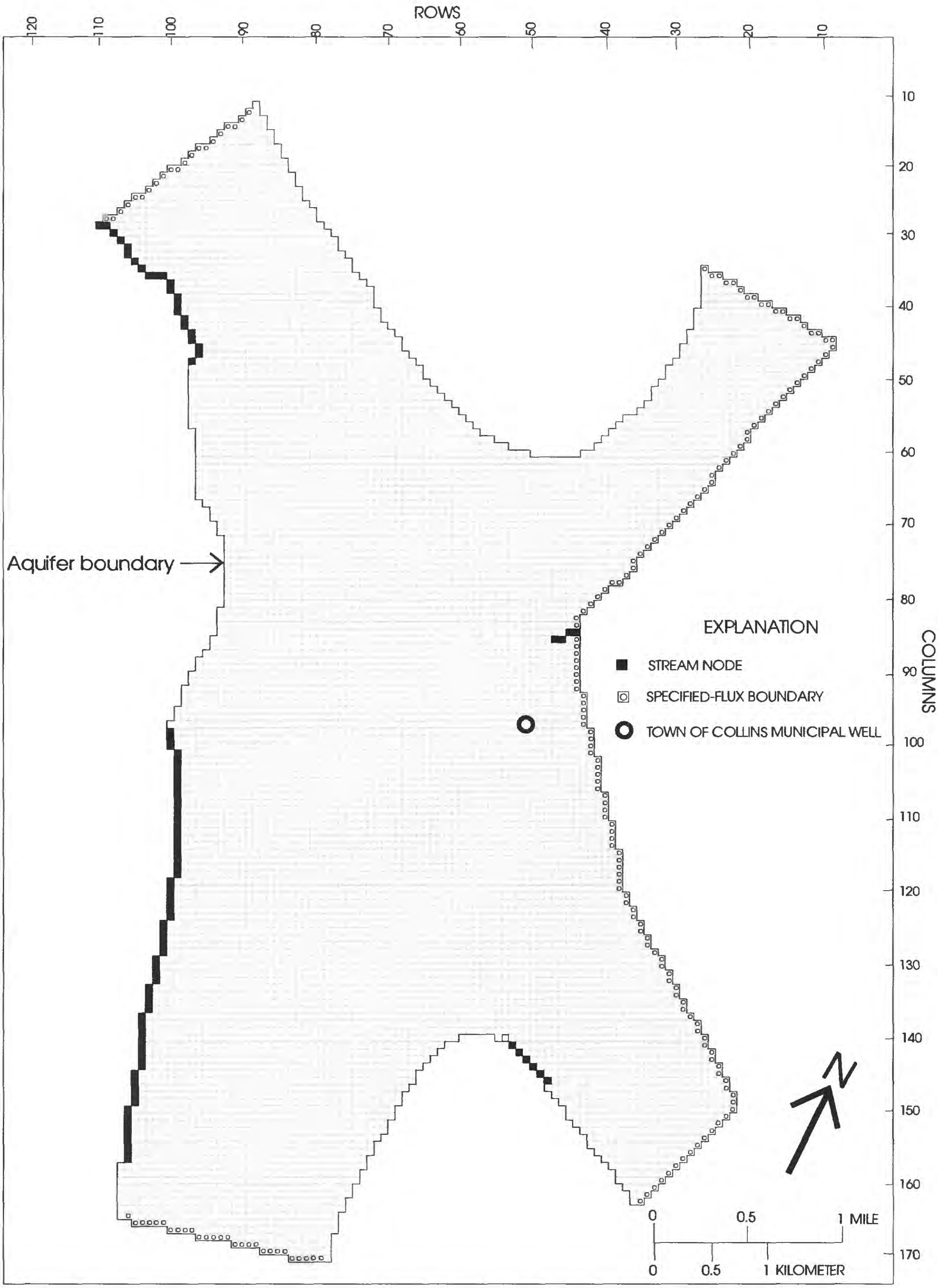


Figure 11. Model grid showing boundary conditions in the Collins study area, Erie County, N.Y.

Contoured hydraulic heads computed by the model for average steady-state conditions indicate similar general directions of ground-water flow in both models (fig. 12A and B), but the computed hydraulic heads in the buried tributary valley (now occupied by Clear Creek, fig. 10) were 15 to 20 ft higher in model A than in model B. Ground water in the buried tributary valley flows northwestward, where some moves toward the Town of Collins municipal well no. 1, and some continues northwestward into the ancestral Allegheny River buried valley.

With a pumping rate at municipal well no. 1 of 365 gal/min (70,300 ft^3/d), little or no ground water flows to the well from the ancestral Allegheny River buried valley or from the unnamed tributary valley to the north; all water in the ancestral Allegheny River buried valley flows northwestward and discharges from the modeled area either as underflow to the northwest or as discharge to Cattaraugus Creek along the southwest valley wall, and water in the unnamed tributary valley (fig. 12A) flows southward, then northwestward into the ancestral Allegheny River buried valley.

In both models, the rates of flow to and from the modeled area were calculated as the sum of contributions from each model cell within recharge areas and along discharge boundaries (table 4). Comparison of the two models indicates that only about one third as much recharge enters from the uplands along the east side of the valley in model B (88,400 ft^3/d) as in model A (245,600 ft^3/d) and, consequently, more water enters the confined aquifer from the other inflow boundaries in model B than in model A (table 4), except at Clear Creek, in which the rate of inflow is equal for both simulations.

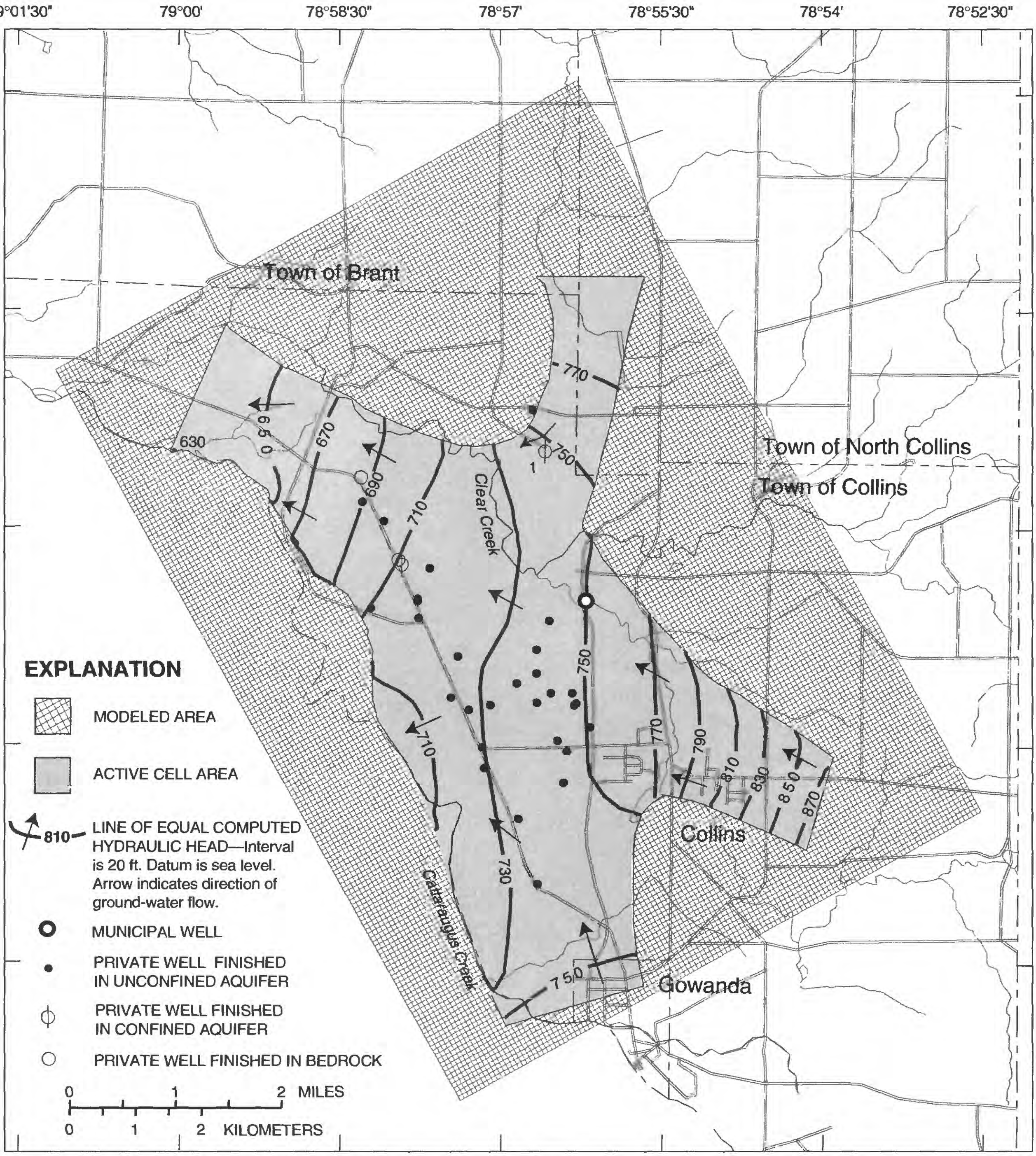
The areas contributing recharge to municipal well no. 1, including areas within the confined aquifer and in adjoining uplands, as indicated by models A and B, are shown in figure 13. In model A, the aquifer and upland contributing areas to municipal well no. 1 were 0.29 mi^2 and 0.95 mi^2, respectively, and all water pumped from the municipal well (70,300 ft^3/d) was derived from recharge from the uplands to the east. In model B, the aquifer and upland contributing areas were larger (0.87 mi^2 and 2.0 mi^2, respectively) and 84 percent of the water pumped (59,050 ft^3/d) was derived from the uplands to the east, while the remaining 16 percent (11,250 ft^3/d) was derived from underflow in the buried-tributary valley now occupied by Clear Creek.

Reliability of Model Results

The amount of recharge entering the confined aquifer from the eastern uplands is an important factor in determining the area contributing recharge to well no. 1, and uncertainty in the recharge rate is a principal limitation in the interpretation of model results. If the recharge rate is relatively large (model A), the contributing area within the aquifer and uplands will be relatively small (fig. 13A), whereas a smaller recharge rate (model B) would result in larger contributing areas in the aquifer and uplands (fig. 13B). The recharge rate could be relatively large if the confined aquifer crops out at land surface near the edge of the valley, where it would receive unchanneled flow from the adjacent uplands, but the recharge rate would be relatively small if the confined aquifer is overlain by fine-grained deposits at the edge of the valley that impede the infiltration of runoff from the uplands.

The recharge rate along the 0.82-mi reach of valley wall in the upland contributing area in model A is equivalent to 1.0 ft^3/s per mile of valley wall, and the recharge rate along the 2.3-mi reach of valley wall that is in the upland contributing area in model B is equivalent to 0.3 ft^3/s per mile of valley wall. The presence of fine-grained deposits at the edge of the valley suggests that model B represents aquifer conditions in that area more accurately than model A and that the contributing area delineated with model B is larger and, therefore, more suitable as a basis for wellhead-protection measures, which require a conservative estimate of contributing-area size. Both rates of recharge are considerably less than the recharge rate of 2.5 ft^3/s per mile reported by Morrissey and others (1988) for unconfined aquifers in valleys in areas of high relief. The thin layers of fine-grained deposits that locally overlie sand and gravel deposits along the valley wall in this study area suggests that the recharge rate is less than that determined by Morrissey and others (1988).

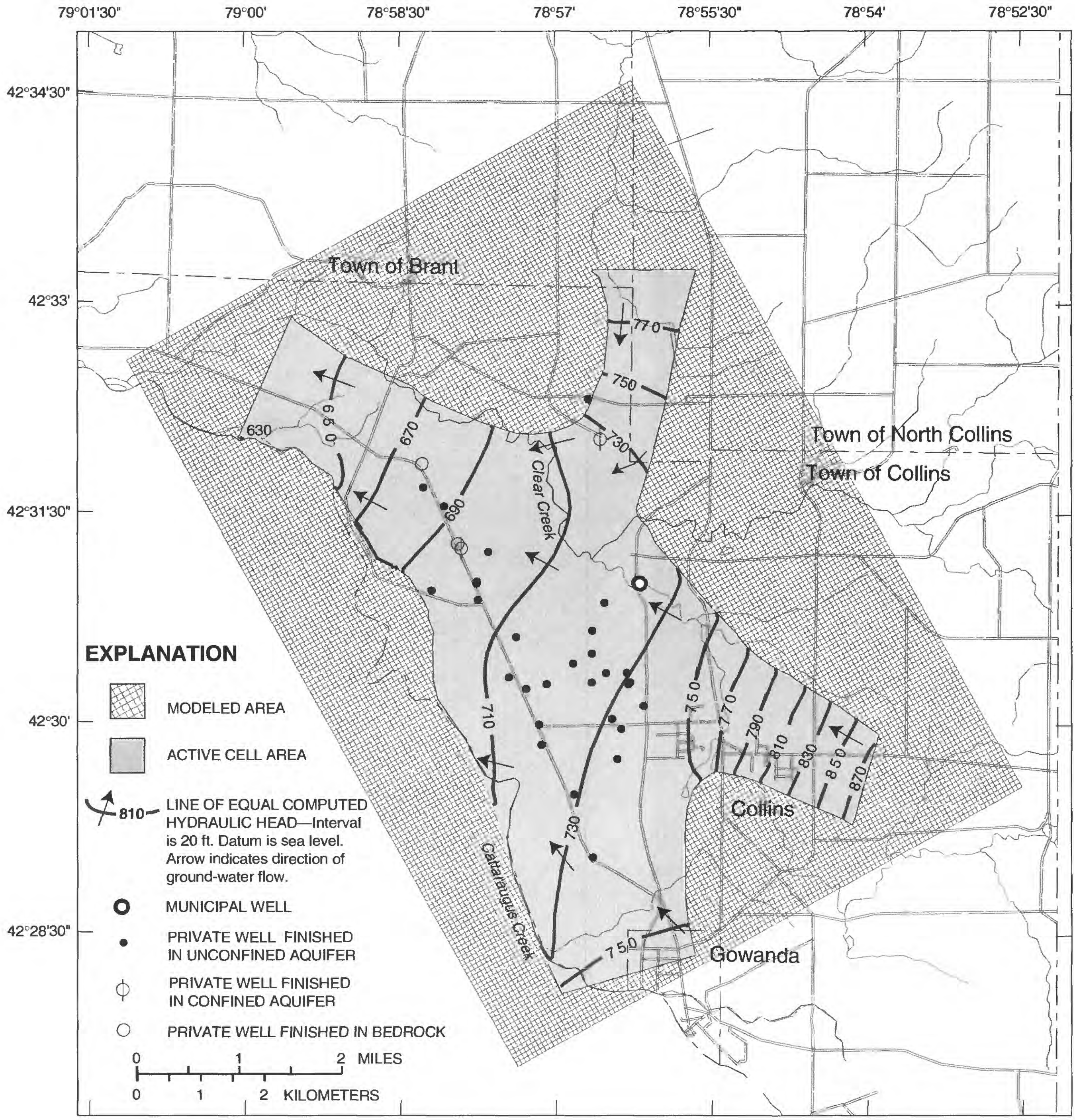
The results of the simulations indicate that most or all of the water pumped by municipal well no. 1 is derived from the confined aquifer in the buried tributary valley now occupied by Clear Creek; therefore, this valley would be a logical area for future investigations to determine the extent, thickness, and hydraulic properties of the confined aquifer through test drilling and installation of observation wells for water-level measurements and aquifer tests. The simulation results also indicate that most of the water pumped from the municipal well is derived from runoff from the uplands to the east. The most suitable location to



Base modified from U.S. Geological Survey digital data, 1:100,000, 1983
New York coordinate system, West zone

A. HIGH RECHARGE RATE

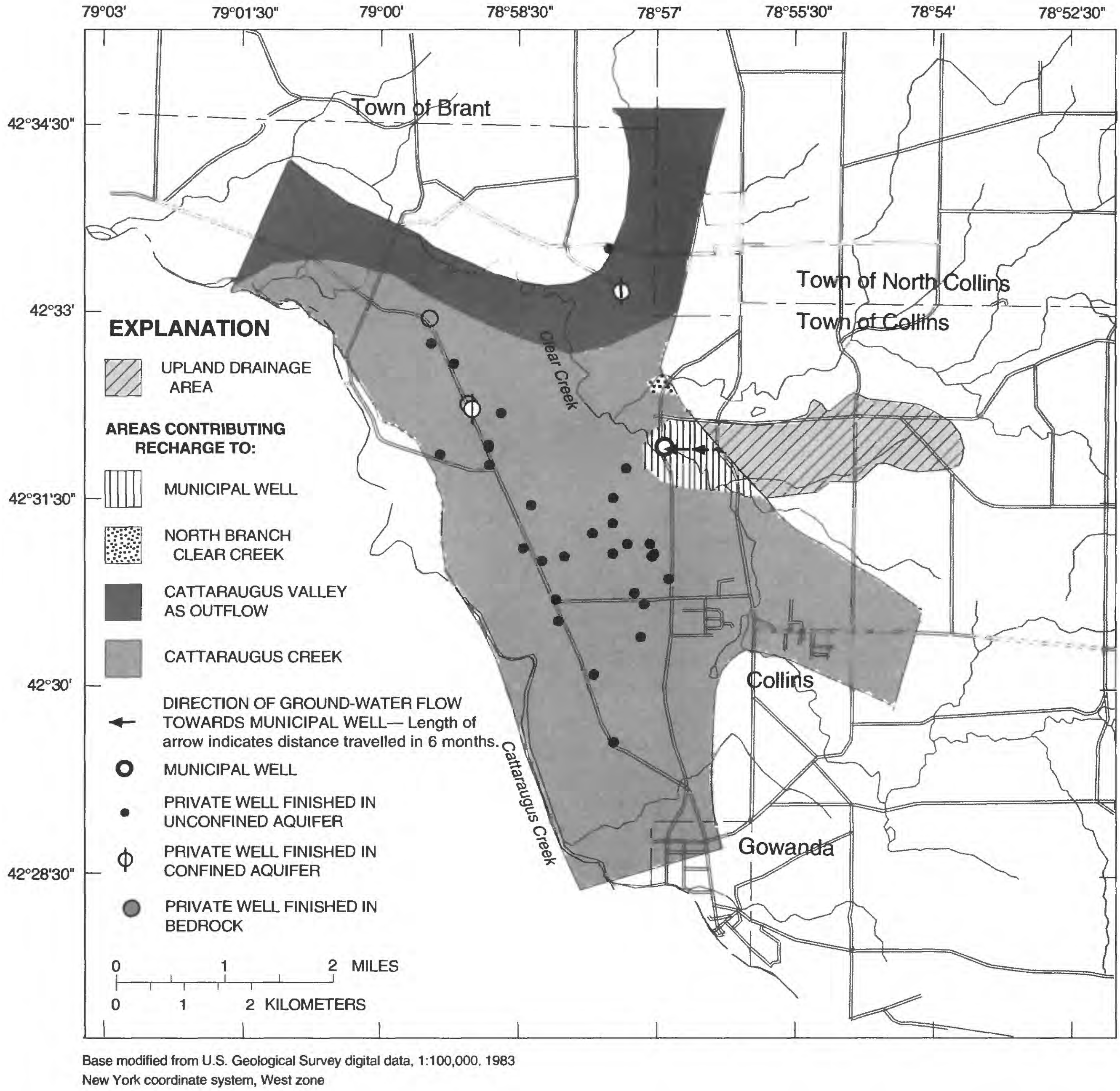
Figure 12 A. Map showing hydraulic head and directions of ground-water flow in Collins area, Erie County, N.Y. computed by steady-state simulation in Model A (high recharge rate)



Base modified from U.S. Geological Survey digital data, 1:100,000, 1983
New York coordinate system, West zone

B. LOW RECHARGE RATE

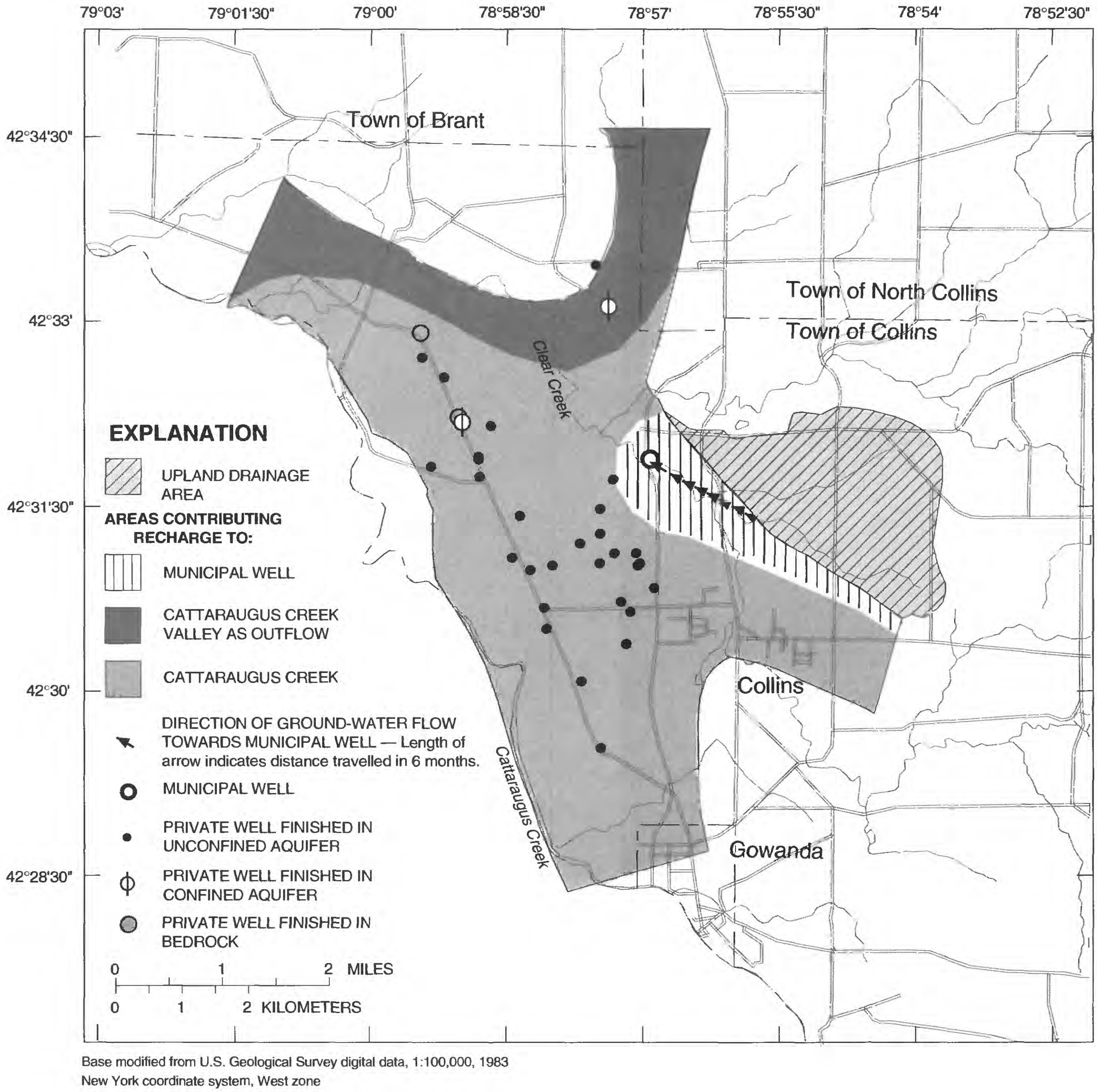
Figure 12 B. Map showing hydraulic head and directions of ground-water flow in Collins area, Erie County, N.Y. computed by steady-state simulation in Model B (low recharge rate)



A. HIGH RECHARGE RATE

Figure 13 A. Map showing areas contributing recharge to municipal well No. 1, Town of Collins, Erie County, N.Y. in Model A (high recharge rate)

install test wells to investigate the hydraulic connection between the confined aquifer and water draining the uplands would be near the east edge of the valley in the aquifer and in the uplands. A chemical comparison of water samples of this area with samples from upland runoff could verify whether the uplands are the principal source of recharge to the confined aquifer.



B. LOW RECHARGE RATE

Figure 13 B. Map showing areas contributing recharge to municipal well No. 1, Town of Collins, Erie County, N.Y. in Model B (low recharge rate)

Holland Study Area

The Town of Holland is on the East Branch of Cazenovia Creek in southeastern Erie County (fig. 14). The Town's water supply consists of two wells that supply 200,000 gal/d to about 1,700 residents. The two wells are 6 ft apart and were drilled to bedrock at a depth of 198 ft in 1932 (LaSala, 1968); both are reported to be screened in a basal, confined sand and gravel aquifer and the upper 10 ft of the bedrock. Usually only one of the wells is pumped at any time. No detailed stratigraphic logs from these or other wells that screen deep, confined aquifers in the Holland area are available. The chemical quality of water from the Holland wells generally meets New York State drinking-water standards, but the water is treated to remove iron and manganese and has slightly elevated concentrations of sodium (33 mg/L) and barium (1.3 mg/L) (G. Barron, Town of Holland, written commun., 1995).

Hydrogeologic setting

The confined aquifer in which the Holland municipal wells are screened is in a narrow, glacially scoured valley that trends northwest-southeast and is bordered on the south by the Lake Escarpment moraine, which rises 500 ft above the valley floor (fig 15). The valley is underlain by Middle and Upper Devonian shale and is drained by the East Branch of Cazenovia Creek, which flows northwestward from the Lake Escarpment moraine to Lake Erie. The East Branch of Cazenovia Creek occupies a buried bedrock valley carved by the preglacial Cazenovia River, which probably extends 30 mi southward to Ischua (fig. 14) (Calkins and others, 1974). Drilling logs and gravity measurements have delineated part of the buried valley between Holland and Sardinia (Calkins and others, 1974); within this reach the buried valley near Chaffee 5 mi south of Holland is filled with as much as 600 ft

Table 4. Simulated water budget for confined aquifer at Collins, in Erie County, N.Y.
[Flow rates are in thousands of cubic feet per day]

Inflow			Discharge		
Source	Rate	Percentage of total	Location	Rate	Percentage of total
Model A (high rate of recharge from adjacent uplands)					
Runoff and lateral flow from unchanneled uplands	245.6	49.8	Town of Collins municipal well no. 1	70.3	14.2
Underflow from buried ancestral Allegheny River valley	115.5	23.4	Cattaraugus Creek	349.2	70.7
Underflow from tributary buried valley now occupied by Clear Creek	94.8	19.2	Underflow to buried ancestral Allegheny River valley	53.8	10.9
Underflow from unnamed tributary valley	21.9	4.4	North Branch Clear Creek	20.5	4.2
Infiltration from Clear Creek	14.0	2.8			
Infiltration from North Branch Clear Creek	2.0	0.4			
TOTAL	493.8	100.0	TOTAL	493.8	100.0
Model B (low rate of recharge from adjacent uplands)					
Runoff and lateral flow from unchanneled uplands	88.4	22.3	Town of Collins municipal well no. 1	70.3	17.7
Underflow from buried ancestral Allegheny River valley	129.1	32.5	Cattaraugus Creek	282.1	71.1
Underflow from tributary buried valley now occupied by Clear Creek	119.2	30.0	Underflow to buried ancestral Allegheny River valley	44.6	11.2
Underflow from unnamed tributary valley	39.1	9.9			
Infiltration from Clear Creek	14.0	3.5			
Infiltration from North Branch Clear Creek	7.2	1.8			
TOTAL	397.0	100.0	TOTAL	397.0	100.0

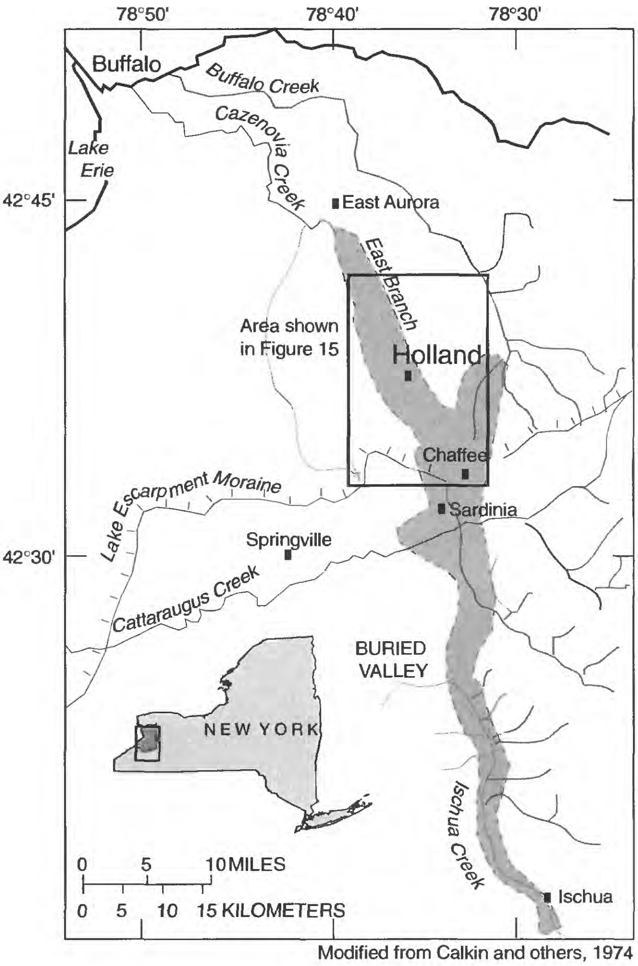


Figure 14. Map showing location of Holland study area

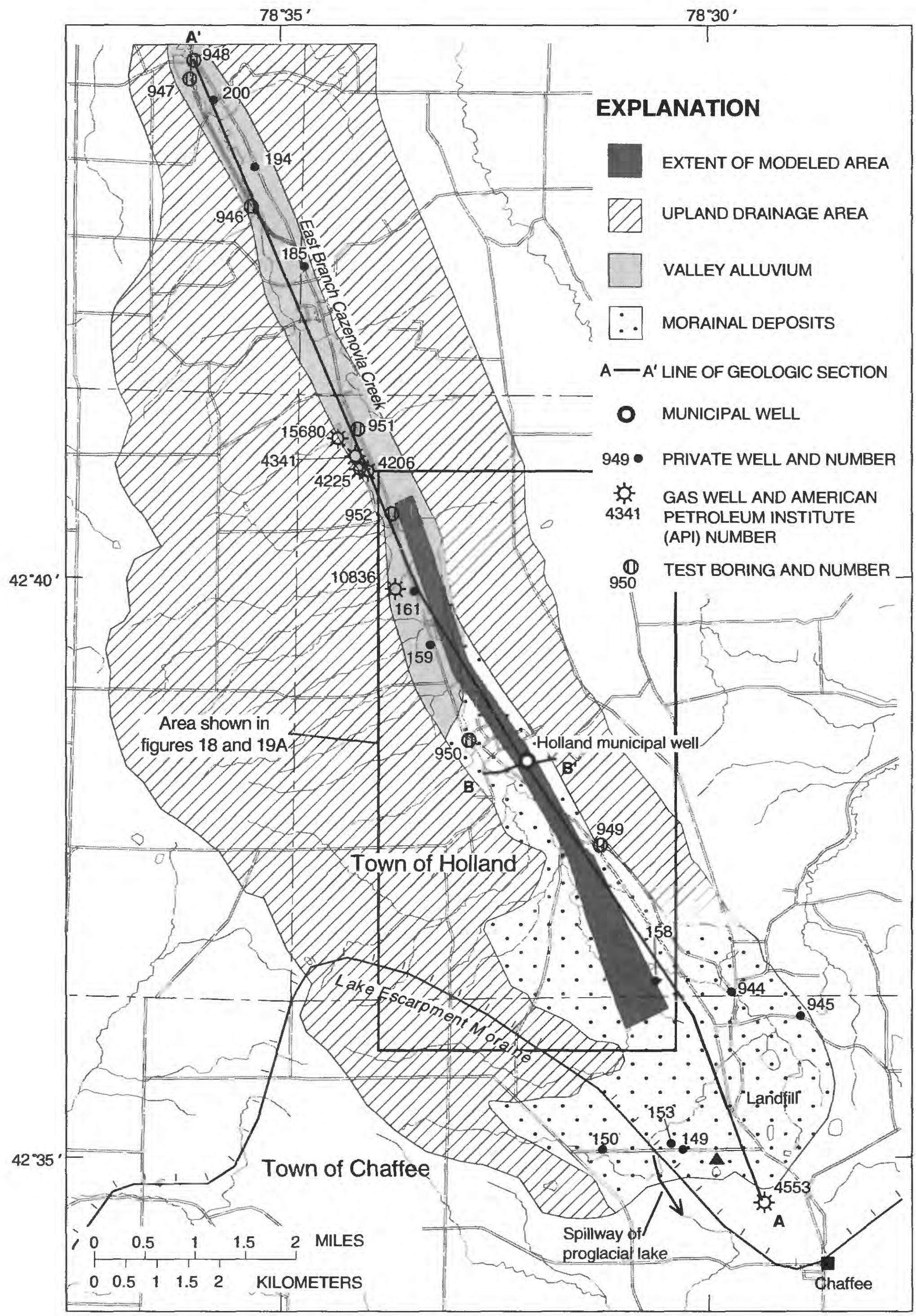
of glacial sediments consisting of surficial sand and gravel underlain by thick deposits of clay till and lacustrine sand, silt, and clay with some confined sand and gravel layers (Miller and Staubitz, 1985).

The buried valley is plugged between Holland and Chaffee by the Lake Escarpment moraine, which was deposited at the ice margin during a widespread glacial readvance 13,000 to 14,000 years ago (Calkin, 1982). The Lake Escarpment moraine is correlated with the Valley Heads moraine in central New York and was probably formed by similar processes (Calkin and others, 1974). The morainal sediments are heterogeneous -- a study of the Valley Heads moraine in central New York near Dryden, 125 mi to the east, by Miller (1993) identified several sand and gravel aquifers separated by till and lacustrine deposits, some of which extend several miles on either side of the moraine.

Little information is available on the types of sediments that form the Lake Escarpment moraine south of Holland. Till is present at land surface and is estimated to be 50 ft thick. Several sand and gravel layers similar to those described by Miller (1993) near the Valley Heads moraine near Dryden could be buried within the moraine, but their extent and depth is unknown. The southern extent of the confined aquifer also is unknown. The presence of a buried aquifer at nearby Sardinia, 3 mi southwest of Chaffe (fig. 14) (Miller and Staubitz, 1985), and possible geomorphic similarity to a continuous confined aquifer beneath the Valley Heads moraine near Dryden (Miller, 1993), suggest that the confined aquifer tapped by the Holland municipal well extends beneath the Lake Escarpment moraine in this area. Additional stratigraphic information would be needed to verify this hypothesis, however.

The East Branch Cazenovia Creek valley north of Holland, is filled with recent alluvium and glacial sediments up to 150 ft thick (fig. 16A) that form two aquifers separated by lacustrine and till deposits. The lower, confined aquifer probably is an extensive ice-contact sand and gravel deposit that overlies the bedrock in most of the valley from the moraine 10 mi northward to East Aurora (fig. 14). The upper, unconfined aquifer consists of sand and gravel 10 to 20 ft thick deposited as alluvial fans along the valley walls. The intervening lacustrine silt and clay layer (30 to 100 ft thick) was deposited in a proglacial lake that formed between the moraine and receding ice front. The till unit(s), 5 to 30 ft thick, were deposited by readvances of ice that reworked the lacustrine deposits. A spillway on the moraine drained the proglacial lake during its early phase (Miller and Staubitz, 1985). Two seismic-refraction surveys across the East Branch Cazenovia Creek valley near Holland indicate that the Holland wells are screened within the deepest part of the bedrock valley (fig. 16B).

Flowing artesian conditions are present in the confined aquifer in the valley north of Holland; that is, ground-water levels are above the top of the aquifer, and wells in the low parts of the valley flow at the land surface (fig. 16A). Artesian conditions indicate that recharge is entering the aquifer at high elevations from upland areas along the valley walls or on the moraine. Water-levels indicate that ground water discharges from the confined aquifer to the Holland wells and probably to the East Branch of Cazenovia Creek downvalley to the north, near East Aurora. Drillers



Base modified from U.S. Geological Survey digital data, 1:100,000, 1983
New York coordinate system, West zone.

Figure 15. Map showing pertinent geographic features and aquifer boundary in the Holland study area, Erie County, N.Y.

report that water from the confined aquifer north of Holland is of poor quality and contains natural gas (Howard Maurer, Maurer Drilling Co., oral commun., 1995) as a result of its proximity to the underlying shale bedrock.

A landfill near the hamlet of Chaffee, 5 mi south of Holland, is within the potential recharge area of the confined aquifer that supplies the Holland wells (fig. 15). The landfill is excavated in clay till on the Lake Escarpment moraine and has a leachate-collection system that was installed around its perimeter in 1981 (Earth Dimensions, Inc., 1981). Although leachate from the landfill could potentially migrate downward through the till and northward toward Holland, no evidence of contamination was discovered at wells in the landfill vicinity sampled by the USGS in 1982 or 1983 (Miller and Staubitz, 1985).

Model design

Ground-water flow through the confined aquifer screened by the Holland wells was represented as a two-layer system consisting of the confined aquifer (layer 2) and the overlying glacial sediments (layer 1). Three steady-state simulations (models A, B and C) were run to examine differing interpretations regarding the major sources of recharge to the confined aquifer. The results can be interpreted only qualitatively, however, because information on ground-water levels and flow rates was insufficient for model calibration. Nevertheless, the model illustrates the important features of the ground-water flow system.

The lower and lateral boundaries of the modeled area were specified as no-flow and correspond to the floor and walls of the bedrock valley that contains the confined aquifer (fig. 17A). The width of the aquifer was assumed to be 1,000 ft, as indicated by seismic-refraction profiles of the deepest part of the bedrock valley (fig. 16B). The upgradient (southern) extent of the confined aquifer is unknown; this boundary was specified as a head-dependent-flow boundary where the bedrock valley widens beneath the Lake Escarpment moraine. The elevation of this southern boundary was interpolated from the hydraulic gradient in the confined aquifer (fig. 16A). The downgradient (northern) boundary was also specified as a head-dependent-flow boundary at the location of a test boring in which water levels were measured (fig. 16A). The modeled area was divided into a variably spaced grid containing 50 rows and 200 columns. Cell widths (perpendicular to the valley axis) represented 50 ft, and cell lengths ranged from 100 ft near the Holland wells to 200 ft at the northern and southern boundaries. The model represents an area of 1.4 mi^2 (886 acres) and contains 4,878 active cells.

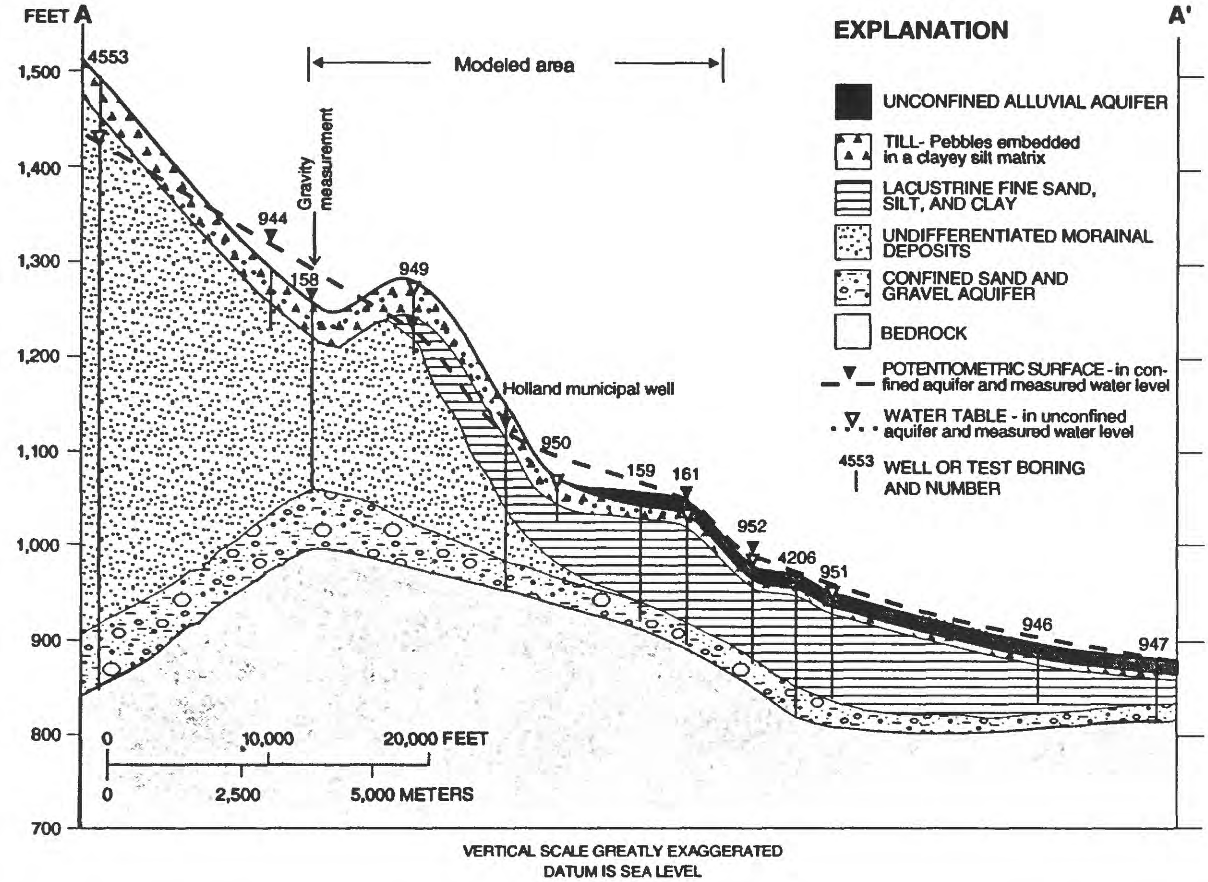
Ground water was assumed to flow horizontally northward through layer 2 (confined aquifer) from the bedrock valley south of Holland (fig. 17A), and vertically upward or downward through layer 1 (morainal and lacustrine sediments overlying the confined aquifer). The rate and direction of flow was controlled partly by the elevation of a constant-head boundary representing the water table in the upper glacial sediments. The altitude of the water table was assumed to parallel land surface and was interpolated from water levels observed in wells and test borings, and from the channel elevations of perennial streams. Flow to a single municipal well was simulated.

The rate of vertical flow through layer 1 was largely proportional to the vertical hydraulic conductivity specified for morainal sediments, which are heterogeneous, and the vertical hydraulic conductivity assigned to this unit (layer 1) represented the most permeable flow paths from land surface to the confined aquifer below. Models A and C used a value of 1 ft/d, equivalent to silty sand (table 6); model B used lower values to represent finer particle sizes. The vertical hydraulic conductivity of the lacustrine sediments in all models, based on permeameter tests of samples of lacustrine sediments deposited by a proglacial lake in the Genesee River valley south of Rochester, N. Y. (Alpha Geoscience, 1995), was specified as 3×10^{-5} ft/d.

A strip of cells along both lateral boundaries of layer 1 were assigned a vertical hydraulic-conductivity

Table 5. Values assigned to aquifer properties in ground-water flow model of Holland area

Aquifer property	Model A	Model B	Model C
Transmissivity of confined aquifer (ft^2/d)	2,500	2,500	2,500
Vertical hydraulic conductivity (ft/d):			
morainal sediments	1	0.1	1
lacustrine sediments	3×10^{-5}	3×10^{-5}	3×10^{-5}
along valley walls	1	1	1
Conductance of southern boundary (ft^2/d)	630	630	63



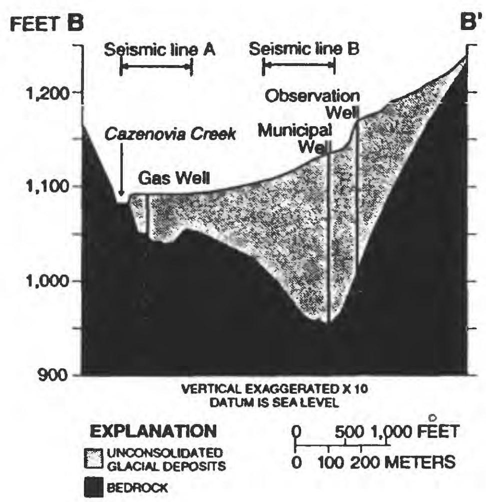


Figure 16. Generalized geologic sections A-A' and B-B' in the Holland area, Erie County, N.Y. (Traces of sections are shown in fig. 15)

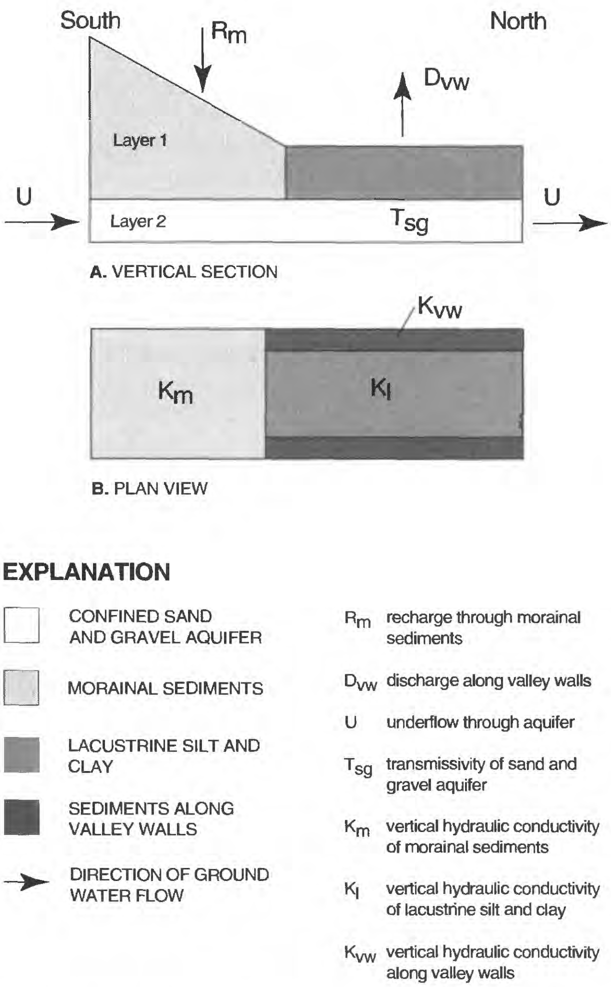


Figure 17. Schematic diagram showing layers, aquifer boundaries and hydraulic properties in model of Holland area, Erie County, N.Y., in A. Vertical section and B. Plan view of layer 1

of 1 ft/d to represent vertical leakage from upland areas through relatively permeable deposits that are assumed to mantle the valley walls (fig. 17B). Ice-contact material and inwash consisting of sand and gravel are commonly deposited along the valley walls by meltwater or by streams draining the surrounding uplands (Ridge, 1991) and could extend along the valley walls from land surface down to the confined aquifer, providing a flow path for ground-water recharge. Ground water also could enter the confined aquifer as downward flow from the weathered bedrock in the till-covered uplands. Recharge from upland areas probably is an important source of water to the confined aquifer; estimated recharge rates in areas of high relief typically exceed 2.5 $(ft^3/s)/mi$ or 41 $(ft^3/d)/ft$ in three unconfined aquifers in the glaciated northeast (Morrissey and others, 1988). This value is assumed to be the maximum rate of recharge to the confined aquifer because its hydraulic connection with the till-covered bedrock uplands is probably limited by fine-grained sediments in the confining layer.

The hydraulic conductivity of the confined aquifer (layer 2) was estimated from records of drawdown and pumping from the Holland wells (LaSala, 1968). The specific capacity of Holland well no. 1 was computed to be 9 (gal/min)/ft, which corresponds to a transmissivity value of 2,500 ft^2/d calculated from equation 1. This value was also used to compute the conductance of the head-dependent flow boundaries at the north and south ends of the confined aquifer. The hydraulic head assigned to these boundaries was interpolated from ground-water levels measured in wells or test borings screened in permeable deposits beneath the lacustrine sediments. Holland municipal wells pump water at an average daily rate of 26,700 ft^3/d (140 gal/min) (G. Barron, Town of Holland, oral commun., 1995).

Simulated ground-water flow

The hydraulic heads in the confined aquifer computed for average steady-state conditions in model A (fig. 18) indicate that ground water flows from the Lake Escarpment moraine northward toward the Holland wells, then northward to East Aurora (fig. 14). The hydraulic gradient is steepest beneath the moraine and flattens to the north. The computed hydraulic head at the Holland wells was 1,093 ft, slightly higher than the observed hydraulic head of 1,087 ft measured during a period of pumping in the 1960's (LaSala, 1968). Computed heads 2 mi north (downgradient) of the Holland wells were 25 ft lower than the observed level of 1,046 ft measured in a well screened in the confined aquifer; this discrepancy could indicate that the hydraulic conductivity of the confined aquifer decreases north of Holland, or that discharge from the confined aquifer along the valley walls is less than assumed. Neither of these possibilities was investigated with model simulations, however.

Rates of flow to and from the modeled area were calculated as the sum of flow into or out of each model cell within recharge areas and along discharge boundaries (table 6). Underflow from areas south of the modeled area accounted for 45 percent of the computed inflow to the confined aquifer; nearly half (48 percent) of the inflow was derived from recharge through the morainal deposits within the modeled area, and the rest was derived from recharge along the valley walls. Discharges from the modeled area included (1) pumpage from one Holland municipal well (23 percent), (2) upward flow through morainal sediments (58 percent) and along the valley walls (13 percent), and (3) underflow to downgradient areas north of the modeled area (6 percent).

The distribution of upward and downward vertical flow between layer 1 and layer 2 indicates that most of the predicted recharge through the morainal sediments discharges to land surface at lower topographic elevations within the moraine and could represent discharge to springs. Upward flow from layer 2 in model A also occurs along the valley walls near the northern boundary of the model and could discharge through the unconfined aquifer (not represented in the model) to the East Branch of Cazenovia Creek. Almost no flow discharges through the lacustrine sediments.

The average rate of recharge throughout the morainal sediments (19 in/yr) is close to the rate of 18 in/yr obtained from calculations of annual base flow by Lyford and Cohen (1988, fig. 3), but the recharge rate of more than 36 in/yr in some cells is unrealistically large, indicating that the assigned vertical hydraulic conductivity could be too large as well. The average rate of recharge from upland areas along the valley walls is about 30 ft^3/d per foot of valley wall, which is less than the assumed maximum rate of 41 ft^3/d per foot (2.5 ft^3/s per mile) reported for unconfined aquifers in the Northeast by Morrissey and others (1988).

In model A the area contributing recharge to the Holland municipal well from morainal sediments is smaller than 0.1 mi^2 (40 acres) (fig. 19B); recharge from these morainal sediments accounts for about 35 percent of the water pumped from the wells; the remaining 65 percent is derived from an upgradient part of the confined aquifer that represents 0.34 mi^2 (220 acres) within the modeled area (fig. 19A). Water flowing through this part of the aquifer is derived from recharge from upgradient parts of the moraine outside the modeled area. Little information is available to determine the exact location of these recharge areas, however.

Table 6. Water budget for confined aquifer at Holland, Erie County, N.Y., under average steady -state conditions in model A

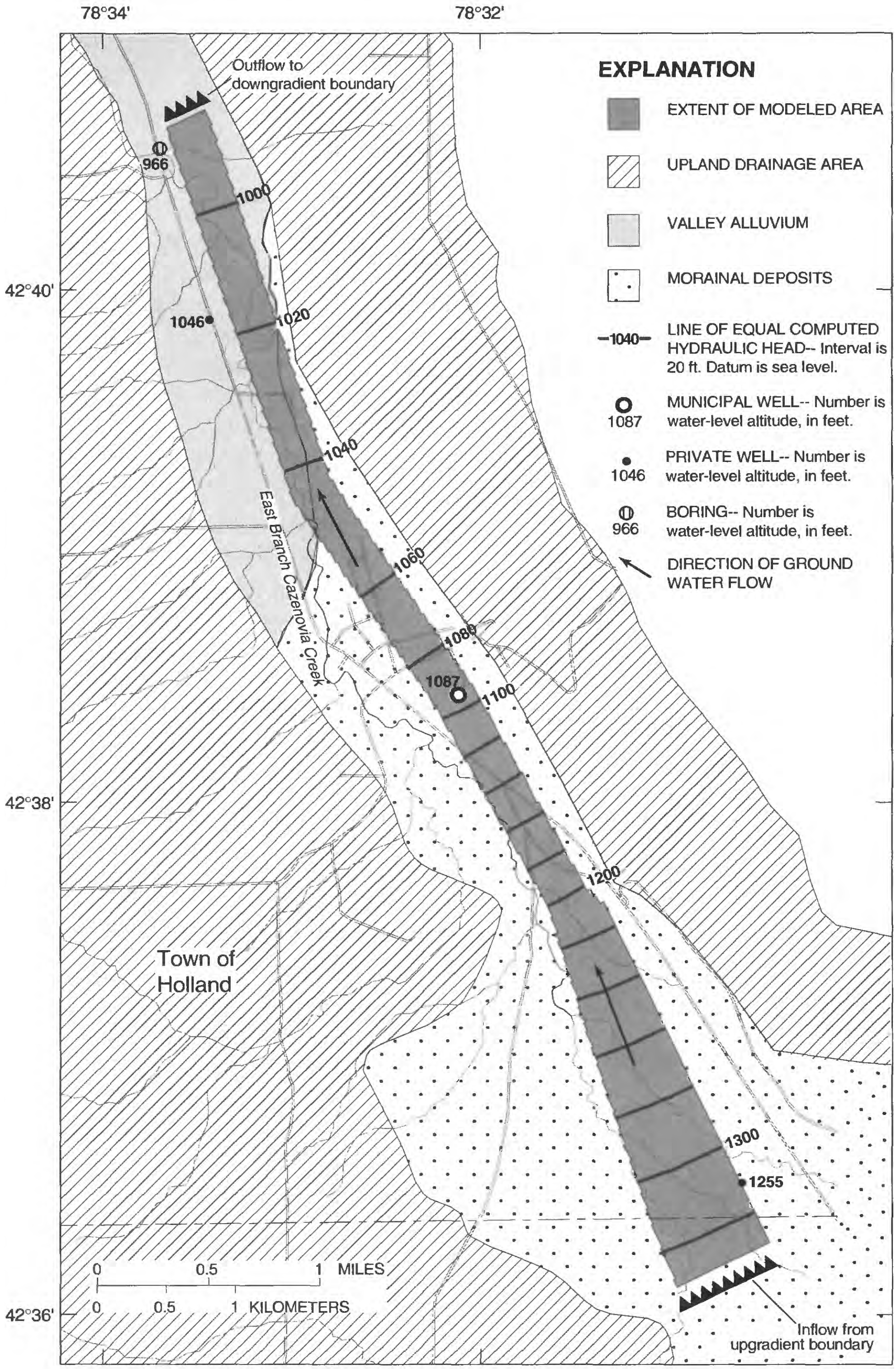
[Flow rates are in thousands of cubic feet per day]

Inflow			Discharge		
Source	Rate	Percentage of total	Location	Rate	Percentage of total
Moraine	56.1	48	Moraine	67.5	58
Valley wall	8.2	7	Valley wall	15.1	13
Underflow	52.1	45	Underflow	7.1	6
			Holland	26.7	23
TOTAL	116.4	100		116.4	100

Reliability of model results

Uncertainty as to the extent of the confined aquifer in the Holland area and the rate of ground-water flow through it are the principal factors limiting the reliability of model results. Little is known about the extent of the confined aquifer either across the valley or southward beneath the Lake Escarpment moraine; additional information on the extent of the confined aquifer would be needed to accurately delineate the areas that contribute recharge to the Holland wells. The computed rate of inflow to the confined aquifer is dependent on the aquifer transmissivity and the vertical hydraulic conductivity of morainal sediments -- changes in these values in the model would affect the size of the resulting contributing areas. The values of the hydraulic properties specified in the model could not be calibrated because too little information is available on the distribution of hydraulic head within the glacial sediments to allow comparison of computed hydraulic heads with measured ground-water levels.

Uncertainty as to the sources of recharge to the confined aquifer was considered in alternative models that represented other rates of inflow from upgradient boundaries (table 5). These results indicate that the boundary of the area within the confined aquifer contributing recharge to the Holland well is relatively insensitive to the rate of leakage through the morainal sediments (model B) or to underflow from the upgradient areas to the south (model C). The boundary of the area that contributes recharge from morainal sediments is sensitive to the rate of leakage through the sediments, however. Decreasing the vertical hydraulic conductivity of the morainal sediments by a factor of



Base modified from U.S. Geological Survey digital data, 1:100,000, 1983
New York coordinate system, West zone.

Figure 18. Map showing steady-state distribution of hydraulic head computed by model A in Holland area, Erie County, N.Y.

10 (model B), decreased the leakage rate through the moraine by 80 percent and the total inflow to the modeled area by 35 percent; it also increased the rate of underflow through the southern boundary by 20 percent and decreased the discharge of ground water upward through the moraine by 50 percent. In this model, the water pumped from the Holland well was derived mainly from underflow through the southern boundary, and leakage through the morainal sediments represented less than 10 percent of the flow to the well.

Decreasing the conductance of the southern, head-dependent-flow boundary by a factor of 10 to decrease underflow from upgradient areas south of the modeled area (model C) increased leakage through the moraine by 50 percent and caused no change in the total inflow. It also increased the size of the area contributing recharge to the Holland wells through the morainal sediments by 32 percent-- from 0.063 mi^2 (40 acres) in model A to 0.083 mi^2 (53 acres) in model C, and more than doubled the percentage of pumped ground water that was derived from leakage through the morainal sediments (from 35 percent in model A to 75 percent in model C).

All three models are equally plausible. The uncertainty in the lateral extent of the confined aquifer could be decreased if additional data could be provided from test drilling and additional seismic-refraction surveys north and south of Holland, but the southern limit of the confined aquifer beneath the Lake Escarpment moraine would be more difficult and costly to determine because the overlying glacial sediments are several hundred feet thick. Borehole geophysical-logging of existing wells could help define the stratigraphy beneath the moraine, but additional test drilling would provide information on the stratigraphy of glacial sediments that would help delineate aquifer boundaries, and also would provide water-level data. Installation of new wells at the Holland well field and in areas up- and downgradient of the well field would provide additional sites for ground-water sampling, and water-level measurements.

The rate of ground-water flow through the aquifer could be estimated through ground-water-isotope studies to indicate the age of the water and, thus, its velocity. This information could be used to estimate vertical hydraulic-conductivity values used in the model and provide estimates of the rate of the leakage through the morainal sediments. These studies would require a water-well survey between Holland and Chaffee to locate wells from which water samples could be obtained and analyzed for isotopes and at which water levels could be measured. This information could be used to determine the age of ground water and the distribution of hydraulic head needed to calibrate the model.

SUMMARY

Areas that contribute recharge to municipal wells screened in three glacial aquifers in Erie County were delineated through conceptual, steady-state ground-water flow models that were based mostly on hydrogeologic information compiled from previous studies and from additional data collected during this 1994 study. The aquifers are confined glacial deposits of sand and gravel that supply water to small water systems serving the Village of Alden and the Towns of Collins and Holland; each system serves less than 5,000 people. Field surveys, including seismic-refraction profiles, stream-discharge measurements, and streambed inspections, were conducted to obtain additional information on depth to bedrock and the degree of hydraulic connection between aquifers and streams.

The ground-water flow models represented aquifer geometry and boundaries, and used estimates of aquifer properties derived from specific capacities of municipal wells and values of hydraulic properties reported in previous studies. The Alden area models were calibrated to measured ground-water levels and discharge to Spring Creek, but the Collins and Holland area models were based on the directions of ground-water flow indicated by known water levels because the data were insufficient for calibration. The rates and directions of ground-water flow computed through steady-state simulations were used with a particle-tracking routine developed by Pollock (1989) to generate ground-water flow paths and delineate contributing areas to wells. Rates of recharge to each contributing area were obtained as the sums of contributions from each model cell along model boundaries within the contributing area. Upland areas outside the modeled area that bordered the contributing areas were delineated from maps of land surface topography and the boundaries of watersheds draining to the model boundary within the contributing area. These areas were assumed to contribute recharge as underflow and unchanneled runoff.

Uncertainty in the size and shape of the areas contributing recharge to the municipal wells was examined through a comparison of areas delineated by

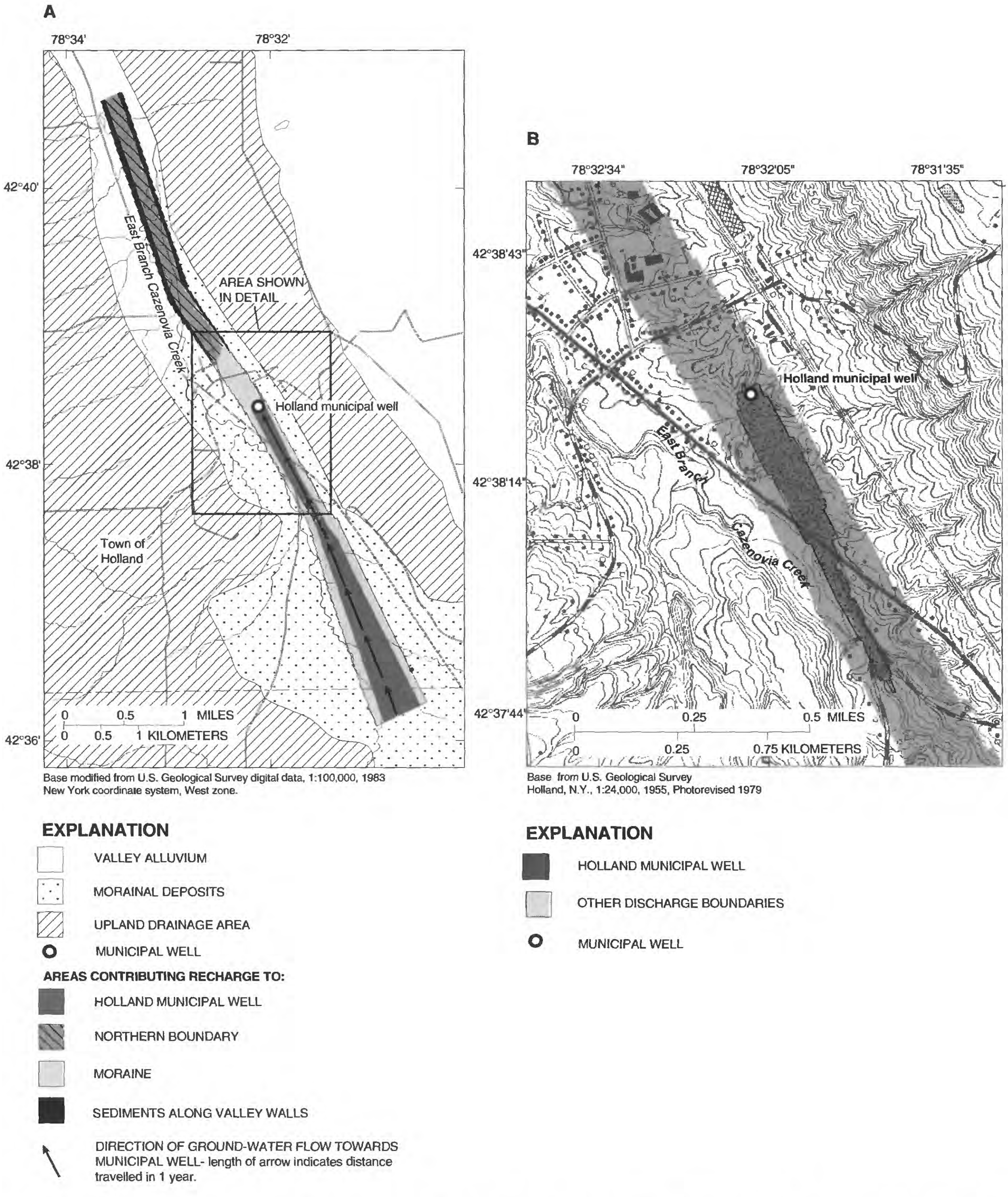


Figure 19. Map showing contributing areas in Holland area, Erie County, N.Y. in model A to: A. Discharge boundaries of confined aquifer (layer 2) and B. Municipal well from overlying glacial sediments (layer 1)

models representing alternative hypotheses concerning aquifer boundaries and properties. The largest contributing areas to municipal wells represent a conservative estimate of contributing-area size needed in the design of wellhead-protection measures. Comparison of results from the alternative models indicated what additional information would be needed to refine the models and thereby decrease the uncertainty in the rates and directions of ground-water flow and in the delineation of contributing areas to municipal wells.

Alden

The Village of Alden water has four wells that supply about 300,000 gal/d (40,000 ft^3/d) to 2,450 people. These wells tap a shallow, confined aquifer 5 to 20 ft thick just north of the Marilla Moraine. The aquifer is overlain by a semiconfining layer up to 20 ft thick and by an unconfined aquifer that consists of sand and gravel 10 to 20 ft thick. The confined aquifer overlies Devonian shale bedrock, the top 5 to 10 ft of which is probably weathered and highly fractured.

The northern extent of the confined aquifer is uncertain but is assumed to coincide with that of the unconfined aquifer, which can be readily identified from land-surface topography and soils maps. Ground-water flow was simulated with two-layer models representing the confined sand and gravel aquifer and the underlying weathered bedrock. Discharge boundaries in the models included (1) wells, (2) perennial streams and springs in the layer representing the sand and gravel aquifer, and (3) underflow to downgradient areas in the layer representing the bedrock. Recharge boundaries included underflow from upgradient area and recharge to the layer representing the confined sand and gravel aquifer to account for downward leakage of precipitation from the unconfined aquifer through the semiconfining layer of fine-grained sediments. Conveyance losses in the Village's water-supply network were represented by additional recharge along the southern boundary within the village limits.

The models were calibrated to measured ground-water levels in 18 wells during 1979-82, and measured ground-water discharge to Spring Creek, through MODFLOWP, a parameter-estimation method that uses nonlinear regression. Transmissivity of the confined aquifer was estimated to be 2,100 ft^2/d, about half the value estimated from the specific capacities of Alden wells no. 3 and 4. Recharge from downward leakage through the semiconfining layer was estimated to be 5.6×10^{-3} ft/d (24 in/yr), which is larger than the estimate of 18 in/yr obtained from average annual baseflow.

The computed hydraulic heads indicate that ground water flows northward from the upland (southern) boundary toward streams and springs that drain the aquifer. About 85 percent of the ground water that exited the modeled area discharged to streams, of which two-- Cayuga Creek and a large tributary to Ellicott Creek-- accounted for 70 percent of the total discharge. Pumping from the four municipal wells accounted for 6 percent of the total ground-water discharge. Nearly all water entering the confined aquifer was derived from vertical leakage through the semiconfining layer. The contributing areas to the municipal wells encompass from 0.05 to 0.07 mi^2 (29 to 45 acres), and ground water that enters the confined aquifer at the southern boundary takes 1 to 2 years to reach the municipal wells, as calculated from an assumed effective porosity of 0.2 for the sand and gravel. Less than 7 percent of the water pumped by municipal wells originates from upland areas south of the aquifer, except for well no. 2 which derives 28 percent of its discharge from upland areas.

Results obtained with two alternative models indicate that, whereas relatively large aquifer-transmissivity values have little effect on the extent of the contributing area, decreased transmissivity values significantly increase the contributing area. Decreased rates of vertical leakage through the semiconfining layer also increased the contributing area and required a smaller aquifer-transmissivity value to reproduce the measured distribution of hydraulic head. In one alternative model, the vertical-leakage rate was lowered 40 percent to 2.3×10^{-3} ft/d (10 in/yr), and the aquifer transmissivity value (estimated by regression) was lowered 40 percent to 860 ft^2/d; as a result, the sizes of the contributing areas to the municipal wells increased by 40 to 90 percent, and the municipal well pumpage doubled from 7 percent of the total ground-water discharge to 14 percent. The uncertainty in aquifer conditions and in contributing areas delineated from model simulations could be decreased if additional information on the rate of ground-water flow through the aquifer were obtained, such as by measuring streamflow to estimate ground-water discharge, or installing monitoring wells in weathered bedrock south of the aquifer to provide information on the rate of underflow to the aquifer from areas beyond the boundary of the confined aquifer.

Collins

The Town of Collins study area is characterized by buried valleys filled with glacial-drift deposits. The upper parts of the valley-fill deposits typically consist of a 10- to 90-ft thickness of alluvial and deltaic sand and gravel underlain by a sequence of mostly fine-grained sediments (till and glaciolacustrine fine sand, silt, and clay) that ranges in thickness from 50 to 350 ft. These sediments are underlain by a basal sand and gravel deposit that typically ranges from 10 to 80 ft thick.

The Town of Collins municipal well no. 1 taps a confined sand and gravel aquifer in a buried-tributary valley now occupied by Clear Creek. The confined aquifer is recharged by: (1) precipitation that infiltrates to the water table where the aquifer is exposed at land surface at the Gowanda Moraine to the east, (2) infiltration of runoff and lateral flow of ground water from unchanneled upland hillsides along the east side of the valley, (3) infiltration from streams that flow over the confined aquifer where it is exposed in places along the edges of the valley, and (4) underflow from upgradient parts of the buried bedrock valleys to the north, east and south. Ground water discharges from the confined aquifer: (1) to pumped wells, (2) to the west in the ancestral Allegheny River valley, and (3) into Cattaraugus Creek along the south edge of the valley.

Ground-water flow was simulated with a one-layer numerical model representing average steady-state conditions in the basal confined aquifer in the buried valleys. Two alternative models were developed: model A used a large conductance value (100 ft^2/d) along the head-dependent inflow boundary to represent a large rate of recharge from the uplands, and model B used a small conductance value (20 ft^2/d) for the same boundary. The smaller value is probably more realistic because fine-grained deposits that overlie the aquifer along the edges of the valley would retard the infiltration of runoff from uplands to the confined aquifer.

In model A (high recharge rate), the area contributing recharge to well no. 1 from the aquifer was 0.29 mi^2, and the area contributing recharge from the upland areas was 0.95 mi^2; the corresponding areas in model B (low recharge rate) were 0.87 mi^2 and 2.0 mi^2 (560 and 1300 acres), respectively. All water pumped by the municipal well (70,800 ft^3/d) in model A was derived from uplands to the east, whereas in model B, 59,600 ft^3/d (84 percent) of the pumped water was derived from the uplands to the east, and 11,200 ft^3/d (16 percent) was derived from underflow through the buried-tributary valley now occupied by Clear Creek.

Results of both models indicate that most water pumped from the municipal well is derived from recharge that enters the confined aquifer from the uplands to the east. This finding could be confirmed if test wells were drilled along the edge of the valley to examine the hydraulic connection between the confined aquifer and ground water draining from the uplands. The principal source of recharge to the confined aquifer could be identified through a chemical comparison of water samples from these wells with samples of upland runoff.

Holland

The Town of Holland's water supply is provided by two wells, pumped one at a time, to provide 200,000 gal/d to about 1,700 residents. The wells tap a deep, confined aquifer within in a narrow, glacially scoured valley now occupied by East Branch Cazenovia Creek which flows northwestward and is bordered to the south by the Lake Escarpment moraine. The confined aquifer consists of sand and gravel overlain by lacustrine silt and clay. The lacustrine sediments are overlain by till 5 to 30 ft thick and by an unconfined aquifer consisting of sand and gravel 10 to 20 ft thick.

The confined aquifer is assumed to be 10 to 20 ft thick and probably extends northwestward within the valley as far as East Aurora; its southward limit is unknown but could be near Sardinia, where a buried aquifer has been mapped. The confined aquifer in the valley north of Holland is artesian, suggesting that recharge enters the aquifer at high elevations from upland areas along the valley walls or on the moraine.

Ground-water flow through the confined aquifer was simulated with a two-layer model representing the confined aquifer and overlying glacial sediments. Data on ground-water levels and flow rates were insufficient for model calibration, but the model illustrates important features of the ground-water-flow system. The lower and lateral boundaries of the modeled area (walls and floor of the bedrock valley) were specified as no-flow, and the aquifer width was assumed to be 1,000 ft, which corresponds to the deepest part of the bedrock valley as delineated by seismic-refraction profiles. The upgradient (southern) and downgradient (northern) boundaries were specified as a head-dependent flow.

A constant-head boundary representing the water table in the upper model layer simulated recharge to the confined aquifer from downward leakage of precipitation through the glacial sediments. The rate of recharge to the confined aquifer was proportional to the vertical hydraulic conductivities specified for morainal sediments (1 ft/d) and lacustrine sediments(3 x 10^{-5} ft/d). A strip of model cells along both no-flow boundaries in the upper model layer was assigned vertical hydraulic-conductivity values of 1 ft/d to represent vertical leakage from upland areas through relatively permeable deposits that are assumed to mantle the valley walls. Transmissivity of the confined aquifer was estimated, from the specific capacity of Holland well no. 1, to be 2,500 ft^2/d.

The computed distribution of hydraulic heads in the confined aquifer indicates that ground water flows from the Lake Escarpment moraine northward toward the Holland municipal well and then northwestward towards East Aurora. Underflow from areas upgradient (south) of the modeled area were nearly equal to recharge through the morainal deposits within the modeled area and together, they accounted for 93 percent of the computed inflow to the confined aquifer. Discharges from the modeled area included: (1) pumping from Holland well no. 1 (23 percent), (2) upward flow through morainal sediments (58 percent), (3) upward flow along the valley walls (13 percent), and (4) underflow to downgradient areas north of the modeled area (6 percent). The area contributing recharge to the Holland municipal well from morainal sediments covers an area of 0.063 mi^2 (40 acres) and provided about 35 percent of the ground water pumped by the well; the remainder of the water pumped by the well flows through a 0.34-mi^2 part of the confined aquifer within the modeled area and extending southward to upgradient areas outside the modeled area.

The average rate of recharge through the morainal sediments (19 in/yr) computed by the model is close to the rate of 18 in/yr estimated from annual base flow, but the computed recharge rate in some models cells exceeds 36 in/yr, which is unrealistic and indicates that the assumed vertical hydraulic conductivity of the morainal sediments (1 ft/d) could be too large as well. When the vertical hydraulic conductivity of the morainal sediments was reduced to 0.1 ft/d (model B), the water pumped by the municipal well was derived mainly from underflow through the southern boundary, and leakage through the morainal sediments accounted for less than 10 percent of the flow to the well. When the conductance of the southern, head-dependent-flow boundary was reduced by a factor of 10 (model C) to limit underflow from upgradient areas south of the modeled area, the area contributing recharge to the municipal well through the morainal sediments increased 33 percent to 0.083 mi^2 (53 acres), and the percentage of the pumped ground water that was derived from leakage through the morainal sediments doubled.

The southern limit of the confined aquifer beneath the Lake Escarpment moraine would be difficult and costly to determine because the glacial sediments are several hundred feet thick. Seismic refraction surveys and borehole geophysical surveys in existing wells could help define the extent of the aquifer and the stratigraphy of glacial sediments that form the moraine, and a water-well survey between Holland and Chaffee could identify wells from which water samples could be obtained for geochemical analyses and at which water levels could be measured. These data could indicate the age of ground water and, thus its velocity and could provide a sufficient basis for model calibration.

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