HYDROGEOLOGY OF THE SURFICIAL OUTWASH AQUIFER AT CORTLAND, CORTLAND COUNTY, NEW YORK

By Richard J. Reynolds

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CONVERSION FACTORS AND ABBREVIATIONS

The following factors may be used to convert the inch-pound units of measurement in this report to metric (International System) units.

Multiply	by	To obtain
	Length	
inch (in.)	2.54	centimeter (cm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.59	square kilometer (km^2)
	Flow	
cubic foot per second (ft^3/s)	0.028	cubic meter per second (m^3/s)
cubic foot per second (ft^3/s)	28.32	liter per second (L/s)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
million gallons per day (Mgal/d)	0.0438	cubic meters per second (m^3/s)
million gallons per day (Mgal/d)	0.003786	cubic meters per day (m^3/d)
gallon per day per foot (gal/d/ft)	0.00014	liter per second per meter (L/s/m)
	Hydraulic Units	
transmissivity, feet squared per day (ft^2/d)	0.0929	meter squared per day (m^2/d)
hydraulic conductivity, feet per day (ft/d)	0.3048	meter per day (m/d)

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Abstract

A previously developed two-dimensional finite-difference ground-water model that simulates ground-water flow in a glacial outwash valley at Cortland, N.Y., was (1) enlarged to incorporate an additional 2.4-square-mile area of outwash aquifer in the adjacent East Branch Tioughnioga and Tioughnioga River valleys, and (2) modified to better simulate induced infiltration from rivers. Calibration of ground-water discharge to streams was based on data from a seepage investigation of the West Branch, East Branch, and Tioughnioga Rivers. The data indicated a ground-water discharge rate of 11.6 cubic feet per second over the 2.84-mile reach and a vertical streambed hydraulic conductivity of 1.04 foot per day.

Hydraulic conductivity of the surficial outwash aquifer in the Tioughnioga River valley was calculated to be 330 feet per day from specific-capacity data obtained from a production well. Simulation of steady-state hydrologic conditions produced heads that were generally within 0.5 foot of observed levels. Comparison of model-generated ground-water-discharge rates with data obtained during a seepage run under similar low-flow conditions indicates that further model refinement is needed with respect to the areal distribution of vertical streambed hydraulic conductivity and areal recharge.

INTRODUCTION

The development and management of the ground-water resources of the Susquehanna River basin requires detailed information on the location and characteristics of potentially productive aquifers, especially in areas that are experiencing, or have the potential for, rapid economic growth. Particularly subject to rapid growth and increased water demands are urban areas and their suburbs. Other areas of increased water demand include outlying industrial parks, housing developments, and major highway corridors. Of particular interest to the Susquehanna River Basin Commission, the agency charged with managing the water resources of the basin, is the extent to which large ground-water withdrawals from aquifers supplying such areas will cause flow in nearby rivers to decrease.

The only aquifers within the Susquehanna River basin in New York that are capable of yielding more than a few tens of gallons per minute to individual wells are sand and gravel aquifers within valley-floor deposits of stratified drift. Most of these aquifers are in direct hydraulic contact with major rivers; consequently large ground-water withdrawals can be expected to cause significant reductions in streamflow by induced infiltration. Several broad valleys in New York are known or suspected to contain extensive aquifers that are largely "separated" from nearby rivers, either by distance or by intervening ridges of till or bedrock. Such "separated-valley" aquifers deserve special attention in water-resources planning because they can be used as storage reservoirs from which large ground-water withdrawals can be made during seasonal low-flow periods without immediately decreasing flow in nearby rivers.

In October 1979, the U.S. Geological Survey, in cooperation with the Susquehanna River Basin Commission, began a series of studies to investigate the ground-water potential of two separated valley aquifers--one near Smyrna, in Chenango County (the subject of a companion report by Reynolds and Brown, 1985), the other at Cortland. The purpose the latter study was (1) to investigate the hydrogeology of the Cortland area and adjacent Tioughnioga River valley, and (2) to enlarge and refine a previous ground-water flow model of the Otter Creek-Dry Creek valley aquifer that was developed by Cosner and Harsh (1978). The model was expanded to include parts of the three adjacent valleys of the Tioughnioga River system to allow model simulations of pumping schemes involving conjunctive use of induced infiltration from the Tioughnioga River and water pumped from storage within the aquifers in the Otter Creek-Dry Creek basin.

Purpose and Scope

This report describes (1) the hydrology and geology of the surficial aquifer at Cortland, (2) the expansion and refinement of the earlier ground-water flow model of the Otter Creek-Dry Creek valley aquifer, (3) the hydrogeology of the parts of the aquifer system that were added to the model, (4) the selection of model boundaries, (5) a revised method of simulating stream leakage to the aquifer, and (6) the results of a steady-state simulation. It also presents results of a seepage investigation of a 2.84-mile reach of the Tioughnioga River and gives hydraulic-conductivity estimates of the surficial outwash aquifer and the streambed material. Selected well and test-hole logs and the results of pumping-test analysis are given as appendices.

Location and Extent of Study Area

The valley aquifer at Cortland, in the northwestern part of Cortland County (fig. 1), is situated at the convergence of three valleys--the West Branch Tioughnioga River, East Branch Tioughnioga River, and the Otter Creek-Dry Creek valleys. This investigation encompassed all of the Otter Creek-Dry Creek basin (investigated by Cosner and Harsh, 1978) and parts of the adjacent West Branch, East Branch, and Tioughnioga River valleys (fig. 1). The location and major geographic features of the study area are shown in figure 1.

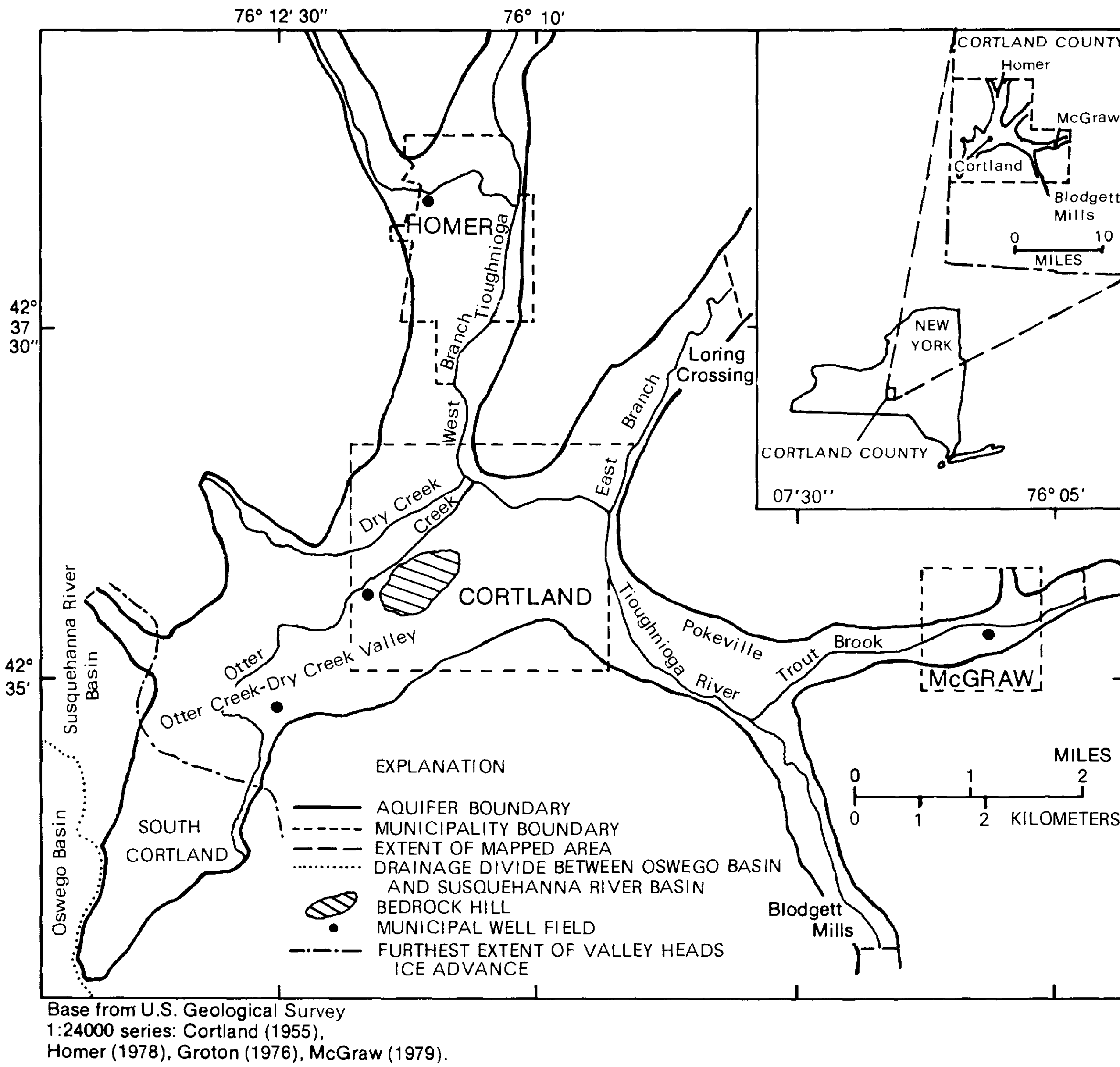
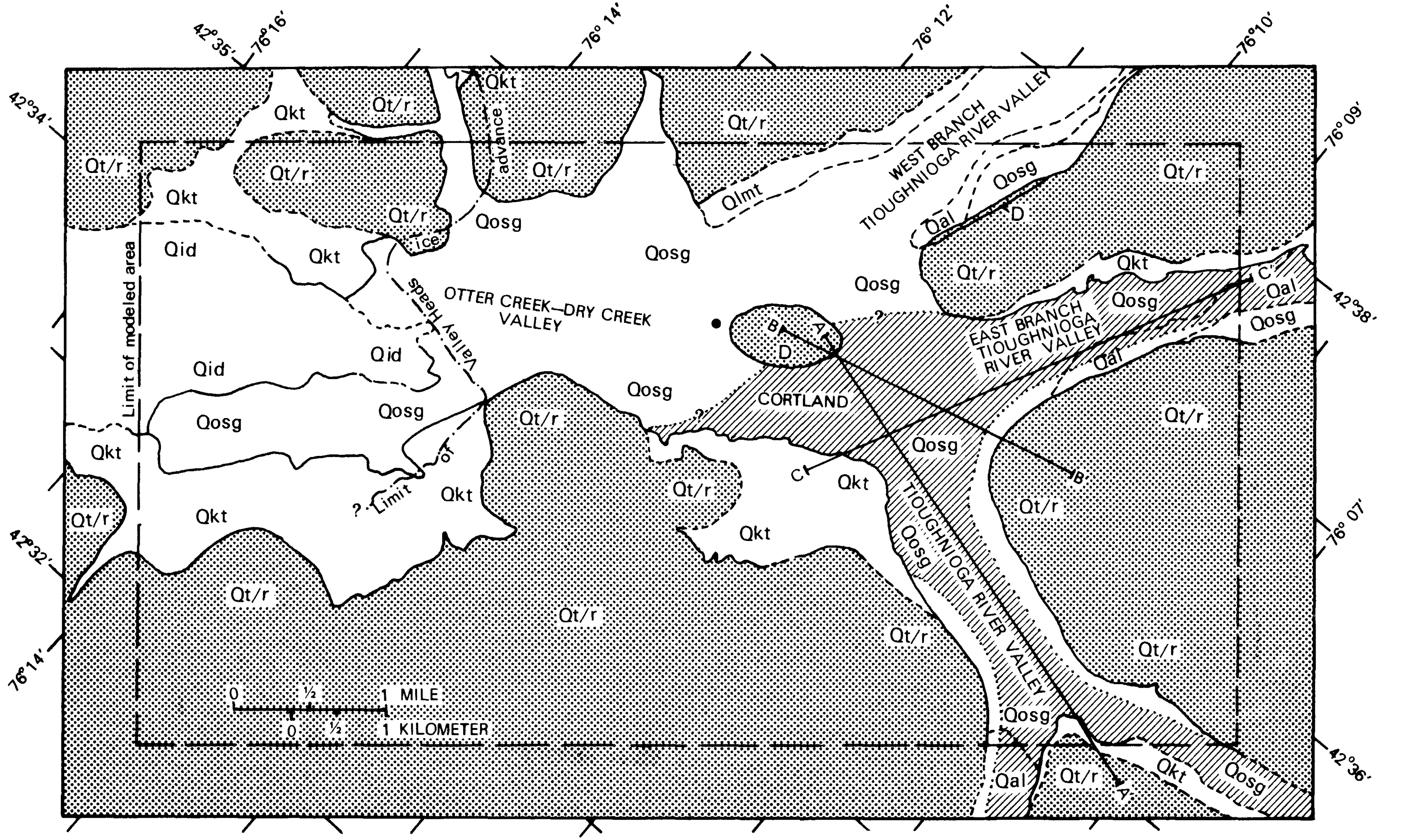


Figure 1.--Location and major geographic features of the study area.

GEOLOGY OF THE CORTLAND AQUIFER SYSTEM

Geologic investigations were conducted from 1979 to 1981 to define the hydrogeologic framework of the valley sections to be added to the Otter Creek-Dry Creek model developed by Cosner and Harsh (1978). This work consisted of a well inventory, a review of drillers' logs, drilling several test holes, and installing several observation wells. (Selected well and test-hole logs are presented in appendix I.) These and other logs were used to construct geologic sections of the area and to aid in developing several of the input data matrices for the model. In addition, a seepage investigation over a 2.84-mile reach of the Tioughnioga River was conducted to evaluate ground-water discharge from the surficial outwash aquifer.



Base from New York State Department of Transportation 1:24,000 quadrangles: Cortland (1979), Homer (1974), McGraw (1974)

Hydrology by R.J. Reynolds, 1982. Surficial geology modified from R.G. LaFleur, 1965 (unpublished field maps) and Cosner and Harsh, 1978

(hatched)	Approximate extent of confined aquifer
Qt/r	Till over bedrock
Qid	Ice-disintegration complex (Valley Heads end moraine)
Qlmt	Lateral moraine till
Qkt	Kame terrace
Qosg	Outwash sand and gravel
Qal	Recent alluvium
D	Devonian bedrock with thin till cover
A├──┤A′	Line of section
── ──	Limit of modeled area
•	Long-term observation well C-19
─ - - -	Contact between surficial geologic units; dashed where approximate
—·—·—	Approximate limit of Valley Heads ice advance

Figure 2.--Generalized surficial geology and extent of confined aquifer in the Cortland area.

The aquifer of primary interest in the Cortland area is a thick outwash deposit that extends from the Valley Heads end moraine, southwest of Cortland (fig. 2), northeastward throughout the city of Cortland and into the adjacent Tioughnioga River valley (Buller and others, 1978; Cosner and Harsh, 1978). The surficial outwash aquifer is of interest because of its hydraulic connection to small streams that cross it and to major rivers that border it. This outwash deposit is flanked in places by kame terraces composed of ice-contact stratified sand, gravel, silt, and clay; locally it contains discontinuous interbedded lenses of fine-grained silt and clay. The outwash becomes more silty and clayey southwestward and gradually grades into the Valley Heads moraine southwest of Cortland (fig. 2; pl. 1).

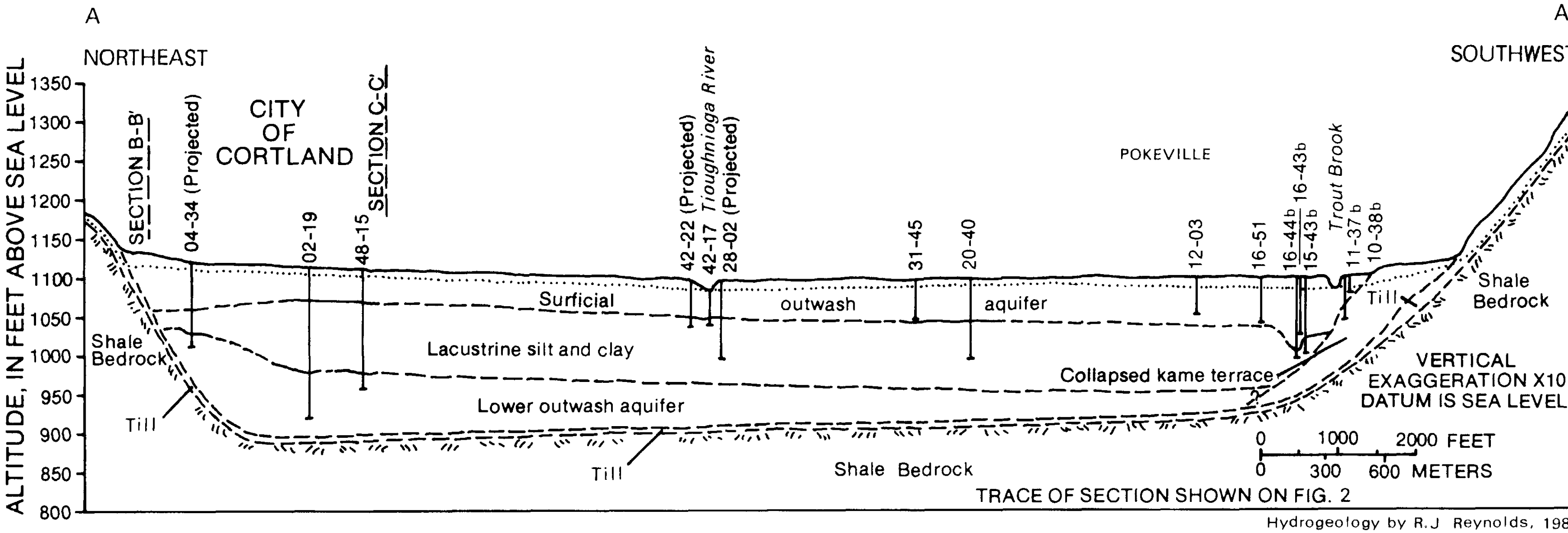
The Valley Heads moraine is an ice-disintegration complex of till and interbedded discontinuous lenses of gravel and sand (Cosner and Harsh, 1978). Surrounding this valley system are till-mantled bedrock hills that rise to a maximum of 700 ft above the valley floor. Within the city of Cortland, a bedrock hill protrudes above the valley floor and partly isolates the Otter Creek-Dry Creek outwash aquifer from the adjacent Tioughnioga River.

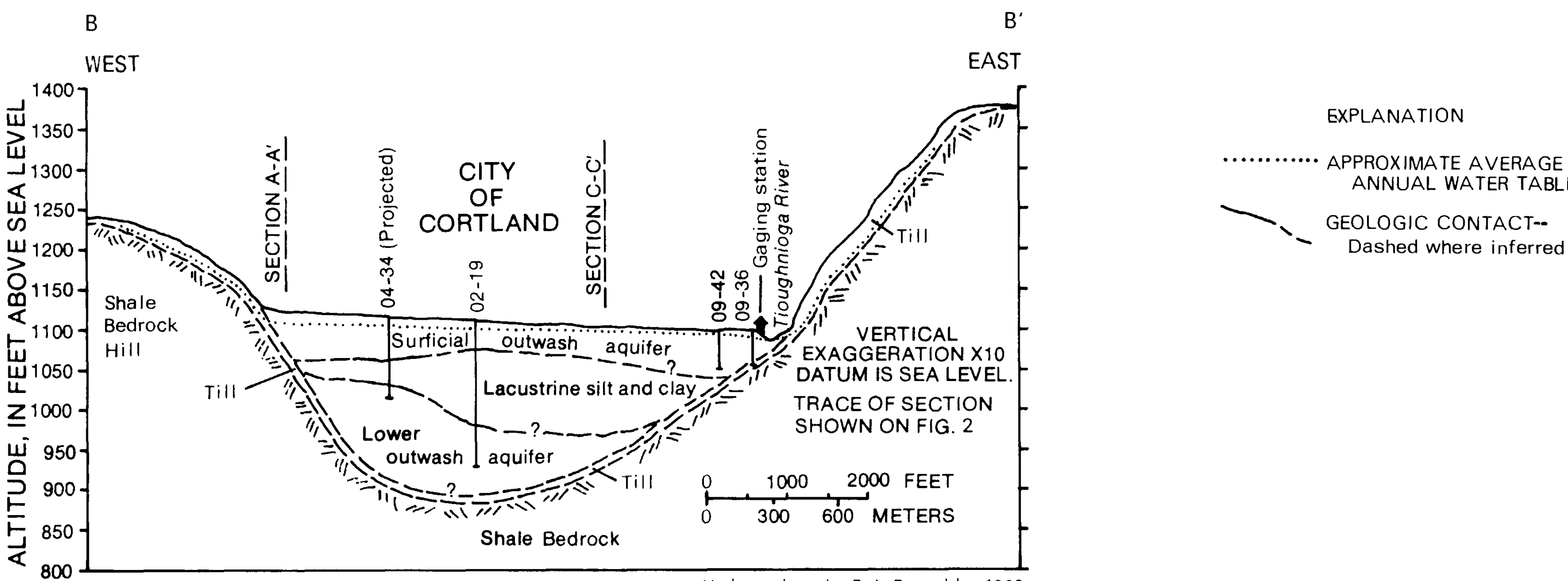
Well-completion data (Randall, 1972) and new test-hole data (appendix I, pl. 1) confirm the presence of a thick sequence of lacustrine silts and clays in the Tioughnioga River valley extending from Cortland southeast down the river valley and northeastward up the East Branch (fig. 2). This lacustrine unit, which pinches out a short distance southwest (upvalley) of Cortland in the Otter Creek-Dry Creek basin, separates the surficial outwash aquifer from a confined, deeper sand and gravel aquifer (fig. 2) in these valley limbs. The relative position of these units is shown in cross section in figure 3.

Surficial Outwash Aquifer

The surficial outwash aquifer is hydraulically connected to the overlying East Branch, West Branch, and Tioughnioga Rivers. Because of this connection, pumping from this aquifer at some locations could induce infiltration from these rivers in addition to capturing ground water that would normally enter them; thus possibly reducing flow in the river to unacceptable levels during annual low-flow periods.

In the East Branch and Tioughnioga River valleys, the outwash forms a water-table aquifer consisting of variably silty sand and gravel. In the Tioughnioga River valley between Cortland and Pokeville, the saturated thickness of the aquifer generally ranges from 30 to 50 ft (fig. 3A). Near Trout Brook, southeast of Pokeville, the saturated thickness increases to approximately 80 ft where the lacustrine unit beneath it pinches out and the surficial aquifer merges with the sand and gravel of a collapsed kame terrace (fig. 3A). In the East Branch Tioughnioga River valley (fig. 3C), the surficial aquifer thins from a maximum saturated thickness of 50 ft at Cortland to about 25 ft at Loring Crossing.


A
A′
NORTHEAST
SOUTHWEST
ALTITUDE, IN FEET ABOVE SEA LEVEL
1350
1300
1250
1200
1150
1100
1050
1000
950
900
850
800
SECTION B-B′
04-34 (Projected)
CITY OF CORTLAND
02-19
48-15
SECTION C-C′
42-22 (Projected)
42-17
Tioughnioga River
28-02 (Projected)
31-45
20-40
POKEVILLE
12-03
16-51
16-44b
16-43b
15-43b
Trout Brook
11-37b
10-38b
Shale Bedrock
Surficial
outwash
aquifer
Lacustrine silt and clay
Collapsed kame terrace
Lower outwash aquifer
Till
Shale Bedrock
VERTICAL EXAGGERATION X10
DATUM IS SEA LEVEL.
0
1000
2000 FEET
0
300
600 METERS
TRACE OF SECTION SHOWN ON FIG. 2
Hydrogeology by R.J Reynolds, 1982


B
B′
WEST
EAST
ALTITUDE, IN FEET ABOVE SEA LEVEL
1400
1350
1300
1250
1200
1150
1100
1050
1000
950
900
850
800
Shale Bedrock Hill
SECTION A-A′
04-34 (Projected)
CITY OF CORTLAND
02-19
SECTION C-C′
09-42
09-36
Gaging station
Tioughnioga River
Till
Surficial
outwash
aquifer
Lacustrine silt and clay
Lower outwash aquifer
Shale Bedrock
VERTICAL EXAGGERATION X10
DATUM IS SEA LEVEL.
TRACE OF SECTION SHOWN ON FIG. 2
0
1000
2000 FEET
0
300
600 METERS
Hydrogeology by R.J. Reynolds, 1982
EXPLANATION
APPROXIMATE AVERAGE ANNUAL WATER TABLE
GEOLOGIC CONTACT-- Dashed where inferred

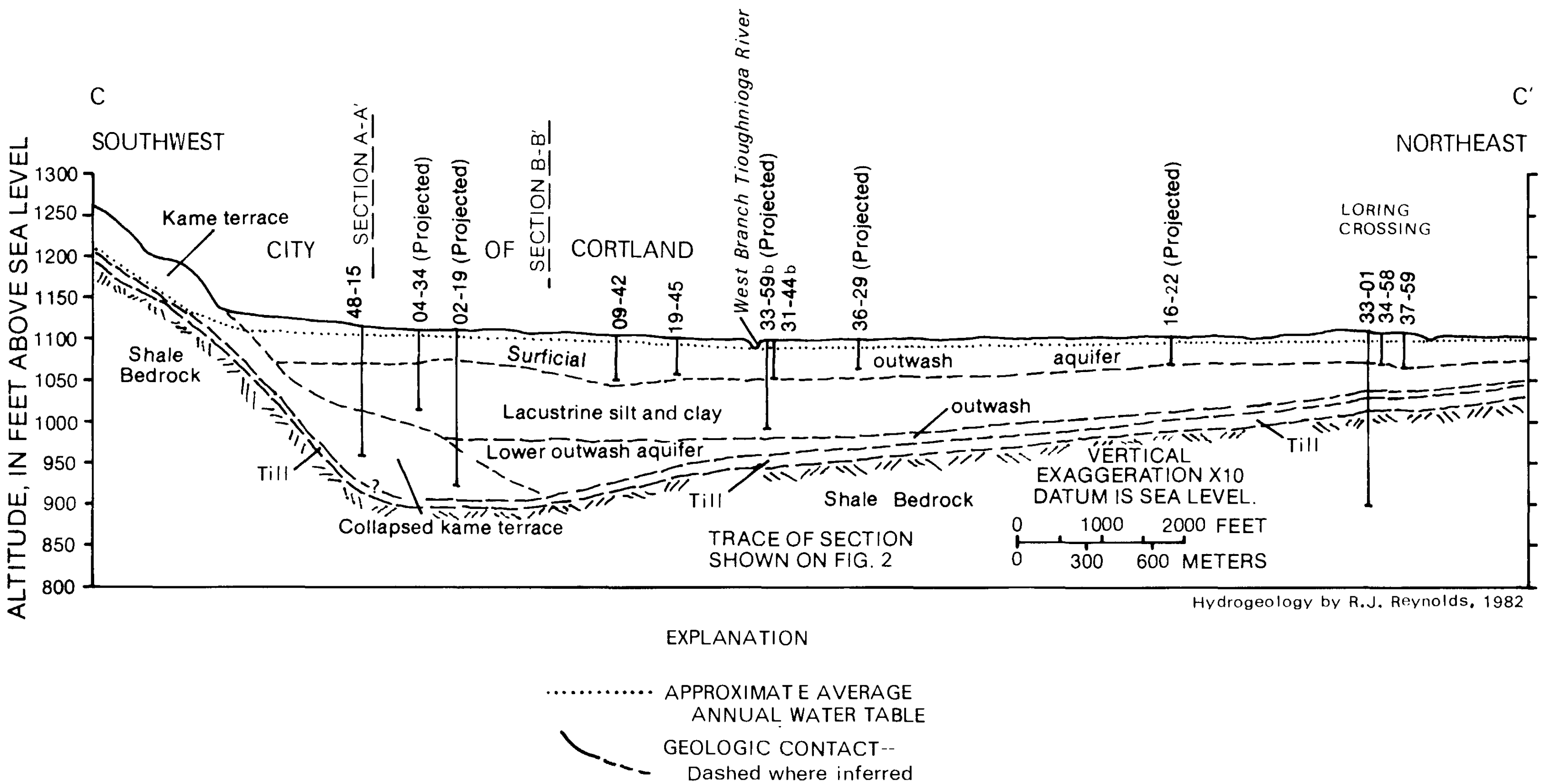


Figure 3.--Geologic sections: A-A', from Cortland to Pokeville along the Tioughnioga River valley near Cortland; B-B', across the Tioughnioga River valley at Cortland; C-C', along the East Branch Tioughnioga River valley from Cortland to Loring Crossing. (Trace of sections is shown on pl. 1 and fig. 2.)

Kame Terraces

Kame terraces are ice-contact, glaciofluvial deposits of stratified sand and gravel with interbedded layers of silt and clay that were deposited by meltwater streams flowing in channels between the glacier and the valley walls. Material was often deposited atop or against the ice bordering these channels, so that when the ice melted, the deposits formed flat-topped, irregularly shaped terraces with components spreading outward into the valley. The two kame terraces that flank the Otter Creek-Dry Creek valley on its southern and northern edges (fig. 2, pl. 1) are hydrogeologically significant because they provide avenues for additional recharge to the outwash aquifer. The buried, collapsed edges of these kame terraces commonly extend downward under the surficial outwash and toward the valley center and may merge with or directly overlie previously deposited outwash. In areas where this collapsed kamic material and outwash is overlain by lacustrine deposits, together they form a confined aquifer that extends into the Tioughnioga River valley (fig. 2). This confined aquifer lies within the eastern part of the Otter Creek-Dry Creek valley and extends from the bedrock hill within the city of Cortland eastward into the Tioughnioga and East Branch Tioughnioga River valleys (fig. 2).

Lacustrine Unit

Lacustrine sediments, which consist of interbedded fine sand, silt, and clay, were deposited in proglacial lakes that developed between older sediment deposits downvalley and the retreating ice front in the Otter Creek-Dry Creek and Tioughnioga River valleys. Meltwater streams carrying a mixture of gravel, silt, and clay under high velocity deposited the coarser fraction of their sediment loads at the edge of these proglacial lakes to form kame deltas, while the lighter fraction, mostly silt and clay, was carried further into the lake, where it gradually settled out to form thick lacustrine units.

Geologic data show that the silt and clay lacustrine unit, which overlies a confined sand and gravel aquifer throughout the East Branch Tioughnioga and Tioughnioga River valleys (fig. 2), is thickest in the Cortland vicinity and thins slightly southeastward toward Pokeville. Section A-A' (fig. 3A) shows the lacustrine unit to be approximately 95 ft thick at the city of Cortland and approximately 80 ft thick near Pokeville. The thickness of the lacustrine unit varies considerably and corresponds to changes in altitude of the underlying kame terrace or outwash surface. The silty clay unit is saturated throughout the study area, but its thickness and low hydraulic conductivity retard upward movement of ground water from the lower aquifer to the upper aquifer. The lacustrine unit thins out southwest of Cortland, as evidenced by the log for well 04-34 (appendix I and fig. 3B), which indicates only 35 ft of this unit.

The underlying aquifer within the city of Cortland probably consists of collapsed and buried material from the large kame terrace south of Cortland (fig. 3B) that extends toward the center of the valley near the southwest city boundary. This collapsed, buried kame terrace provided an elevated surface upon which the lacustrine unit was deposited and thus accounts for the smaller thickness of the lacustrine unit in this area.

The lacustrine unit thins rapidly northeastward from a maximum thickness of 95 ft at Cortland to about 35 ft at Loring Crossing as a result of the upward slope of the underlying units (fig. 3C). Figure 2 shows the approximate areal extent (at depth) of the lacustrine unit and underlying confined aquifer within the valleys.

Confined Outwash Aquifer

Beneath the lacustrine unit in the East Branch Tioughnioga and Tioughnioga River valleys (fig. 2) is a confined sand and gravel aquifer of highly variable thickness that is probably older outwash. Its variable thickness is due, in part, to erosion of its upper surface and its merging with collapsed kame-terrace deposits in some areas; its thickness also reflects the altitude of the underlying till and bedrock surfaces. South of the city of Cortland (fig. 3C), this unit merges with collapsed kame-terrace material to form the thickest confined section, estimated to be approximately 75 ft thick. From Cortland northeast to Loring Crossing (fig. 3C), this buried aquifer thins to less than 10 ft thick as a result of the upward slope of the underlying bedrock surface and probably pinches out farther north in the valley. Geologic data on the thickness of this unit in the Tioughnioga River valley (fig. 3A) southeast of Cortland, although scant, suggest that the unit may thin from 75 ft at Cortland to about 30 ft at Pokeville.

Bedrock and Till

Bedrock in the study area consists of shale units of the Upper Devonian Genesee Group (Rickard and Fisher, 1970). The bedrock configuration (Cosner and Harsh, 1978) is representative of a preglacial drainage network, but its surface configuration is not well known. Depth to bedrock in the area ranges from land surface to more than 200 ft. Where shale is exposed at land surface it is weathered and jointed, but the number and size of openings decrease with depth (Cosner and Harsh, 1978). The joints and bedding planes, which form only a small fraction of the total bedrock volume, are the only significant void space in which water can be stored and transmitted. Thus, shale bedrock in this area is a relatively low-yielding source of water and is used only for farm and domestic supplies.

Till in this area is a poorly sorted mixture of silty clay with varying amounts of sand, gravel, and boulders. A veneer of till directly overlies shale bedrock in the uplands surrounding the study area, where it may range in thickness from 2 to 20 ft. Till is also assumed to overlie bedrock in parts of the valley (figs. 3A, 3B, 3C), but test-hole data are insufficient to confirm this. Till in this region generally has a low hydraulic conductivity and therefore does not transmit or yield water readily. However, sufficient domestic supplies can usually be obtained from shallow, large-diameter dug wells excavated in till. Because of this low permeability, most of the precipitation falling on till in upland areas does not infiltrate but is routed to streams as runoff. However, some recharge does occur in areas of upland till, principally during the winter months, and provides base flow to the small streams that drain these areas (Cosner and Harsh, 1978).

HYDROLOGY OF THE CORTLAND AQUIFER SYSTEM

The valley configuration at Cortland represents a hydrologic situation that is uncommon in New York, in that the aquifers in the Otter Creek-Dry Creek valley are, to a large extent, hydraulically separated from the adjacent Tioughnioga River.

Most of the productive valley aquifers within the Susquehanna River basin and elsewhere in upstate New York are parallel to and have a direct hydraulic connection with streams or rivers that cross or border them, so that heavy pumping from the aquifers, even for relatively short periods, normally reduces adjacent streamflow by induced infiltration. The Otter Creek-Dry Creek outwash aquifer, however, is partly separated from the West Branch Tioughnioga River and Tioughnioga River by a bedrock hill in the center of the valley that serves to disrupt the ground-water flow field between the municipal pumping centers and the river. In addition, the distances from the bordering rivers to the municipal well fields are large enough to ensure that induced infiltration from the rivers would not readily take place. Moreover, the difference between the average altitude of the water table at the Cortland municipal well fields and the average river stage would seem to reduce the likelihood of causing induced infiltration under normal circumstances.

Another type of separated aquifer, a confined outwash aquifer, lies beneath the thick deposits of lacustrine clay and silt within the city of Cortland and in the East Branch Tioughnioga and Tioughnioga River valleys. Confined valley aquifers such as this are hydraulically separated from the overlying rivers by thick lacustrine units, which generally have a very low hydraulic conductivity. Pumping from the confined aquifer in the Cortland area, therefore, would not readily induce infiltration from the Tioughnioga River because the lacustrine unit would inhibit any downward leakage.

System Boundaries

The aquifer system at Cortland consists of variable lithologic units that form a surficial outwash aquifer and a confined outwash aquifer. The study area comprises the Otter Creek-Dry Creek basin and parts of the East Branch, West Branch, and Tioughnioga River valleys.

The aquifer system in the Cortland area is bounded on the southwest by the Valley Heads terminal moraine, which forms the surface-water and ground-water divide between the Oswego drainage basin and the Susquehanna River basin (fig. 1). The Tioughnioga River, on the east and southeast side of Cortland, is the discharge point for much of the ground water moving through the aquifer system. The bedrock valley walls flanking the Otter Creek-Dry Creek basin and the Tioughnioga River are treated as impermeable boundaries because shale bedrock generally transmits little water; in other words, the amount of ground water seeping from the bedrock to the aquifer is negligible compared to the amount contributed by direct precipitation and adjacent stratified drift.

The lacustrine unit in the eastern part of the Otter Creek-Dry Creek basin and in the Tioughnioga River valley (figs. 3A, 3B, 3C) overlying the confined outwash aquifer (fig. 2) acts as a confining unit and produces artesian conditions in the underlying aquifer.

Recharge

The aquifer system at Cortland is recharged primarily through infiltration of precipitation, although some recharge occurs as leakage from streams, as ground-water flow from flanking kame-terrace deposits, and from runoff from adjacent bedrock hills. Rates of recharge to stratified-drift aquifers such as these vary seasonally and areally and must be estimated because they cannot be measured directly. Randall and others (1966) show that average annual recharge to stratified drift may reach 1 (Mgal/d)/mi^2. Cosner and Harsh (1978) used a recharge rate of 28 in/yr, which equals approximately 1.25 (Mgal/d)/mi^2. Over a modeled area of 7.9 mi^2, this amounts to 9.89 Mgal/d of recharge. Although about 30 percent of the average annual precipitation (41 in.) falls during the growing season (Sealy and others, 1961), nearly all of it is lost through evapotranspiration, and only a small amount recharges the aquifer (MacNish and others, 1969).

Surficial aquifer

The surficial aquifer is recharged primarily by precipitation and by streambed leakage over losing reaches of small streams in the Otter Creek-Dry Creek basin and probably elsewhere in the valleys. Recharge from Otter Creek and Dry Creek occurs along the stream channels where they traverse the valley floor, from the point where the streams leave the till-covered bedrock uplands to their confluence with the West Branch Tioughnioga River (Cosner and Harsh, 1978; Buller and others, 1978).

Additional recharge to the surficial aquifer also occurs through the kame terraces that flank the north and south edges of the Otter Creek-Dry Creek basin (fig. 2). These kame terraces are hydraulically connected to both the surficial and confined aquifers, and precipitation on these kame terraces therefore recharges both the upper and lower outwash aquifers.

Confined aquifer

The confined outwash aquifer, which underlies the eastern end of the Otter Creek-Dry Creek basin and extends throughout the Tioughnioga River valley, is recharged wherever it is hydraulically connected with the upper aquifer--that is, wherever the intervening lacustrine unit is absent. The recharge area for the confined aquifer is assumed to be west of the bedrock hill in the western part of the Otter Creek-Dry Creek basin (fig. 2) because the lacustrine unit in this area pinches out, thus allowing good hydraulic connection between the two aquifers through which recharge can occur. Additional recharge to the confined aquifer occurs where the aquifer is hydraulically connected to kame terraces at the valley sides, such as near the south side of Cortland and near Pokeville (figs. 2, 3A, 3C).

Ground-Water Movement

Information about directions of ground-water flow and seasonal water-table configurations was obtained from periodic water-level measurements in a network of observation wells within the Otter Creek-Dry Creek basin during

1976 (Cosner and Harsh, 1978). Additional information about general ground-water flow directions and water-table fluctuations in the extended-model area was obtained during 1980-81 from periodic water-level measurements in two observation wells and from historical water-level data. The 1976 water-level data and the 1980-81 data together provided the basis for a steady-state calibration of the enlarged model.

Ground-water movement in the surficial outwash aquifer within the Otter Creek-Dry Creek basin is generally northeastward (downvalley) toward the West Branch Tioughnioga and Tioughnioga Rivers and moves eastward, more perpendicular to the river, within the city of Cortland. Ground-water flow in the East Branch Tioughnioga River valley is predominantly eastward across the valley toward the East Branch Tioughnioga River, and flow in the Tioughnioga River valley between Cortland and Pokeville is westward across the valley toward the Tioughnoiga River. Along some valley walls that border the aquifer, such as southwest of Cortland, ground water from kame terraces flows into both the surficial aquifer and confined aquifer. General directions of ground-water flow in the surficial aquifer during spring conditions are shown on plate 2.

Historic water-level data on the lower outwash aquifer are scant; however, hydraulic heads in this aquifer are probably above the water table in most areas within the Tioughnioga valley and seasonally above land surface in the vicinity of Cortland. Records for well 48-15, drilled in 1944 for Brewer-Tichner Corp. (pl. 1, figs. 3A, 3C) and screened in the lower unit, indicate that the well flowed for approximately 1 month after completion (Randall, 1972). Because the principal recharge area for the confined aquifer is in the Otter Creek-Dry Creek basin, ground-water flow in this aquifer is probably northeastward from the recharge area toward the Tioughnioga River, then southeastward in the Tioughnioga River valley. The discharge area for ground water moving through the confined aquifer is assumed to be further south in the Tioughnioga River valley, beyond the study area.

STREAM-AQUIFER RELATIONSHIP

Several factors determine the rate of ground-water flow between the surficial aquifer and the overlying rivers or streams. The most important of these are (1) the head difference between the stream (or river) stage and the underlying aquifer, (2) the vertical hydraulic conductivity of the streambed material, (3) the hydraulic conductivity of the aquifer, (4) the depth of incision of the stream into the aquifer, and (5) the proximity of nearby impermeable boundaries, such as a bedrock wall, which may alter ground-water flow paths near the river.

In this study, an effort was made to evaluate the amount of ground-water seepage to the Tioughnioga River and to estimate the hydraulic conductivity of the streambed and the aquifer. This was done by measuring discharge at successive points along the Tioughnioga River near Cortland to determine areas of gains or losses of flow (generally known as a seepage investigation), combined with water-level measurements at nearby wells and augmented by analysis of specific-capacity data from a nearby industrial well screened in the surficial aquifer. Data thus obtained were used in data arrays to represent certain areas of the expanded and revised ground-water model.

Ground-Water Discharge

A seepage run consisting of five discharge measurements was made along a 2.84-mi reach of the West Branch, East Branch, and Tioughnioga Rivers in the vicinity of Cortland on August 31, 1981, to determine the ground-water discharge from the surficial outwash aquifer to the adjacent rivers and to provide data that could aid in calibration of the expanded model. On this date, instantaneous flow at the gaging station at Cortland (station 01509000) was at its lowest point of the summer, 58 ft^3/s. As calculated from 43 years of record, this low flow is equaled or exeeded about 90 percent of the time. The measurement sites (fig. 4) were selected to cover the area of suspected high ground-water discharge. One measurement was made on the West Branch, one each at the confluence of the West and East Branches with the main branch, one downstream from the gaging station at Cortland, and one approximately 1 mi downstream near the point where the river crosses over and reaches the west side of the valley. Results of the seepage run are summarized in table 1.

The reach of greatest ground-water seepage was reach B, from the confluence of the East and West Branches south to a point near the Port Watson Street bridge (fig. 4), with a gain of 5.0 ft^3/s over 2,900 ft of stream channel or an average gain of 1.72 ft^3/s per 1,000 ft of channel. This high rate of ground-water seepage is attributed to two factors:

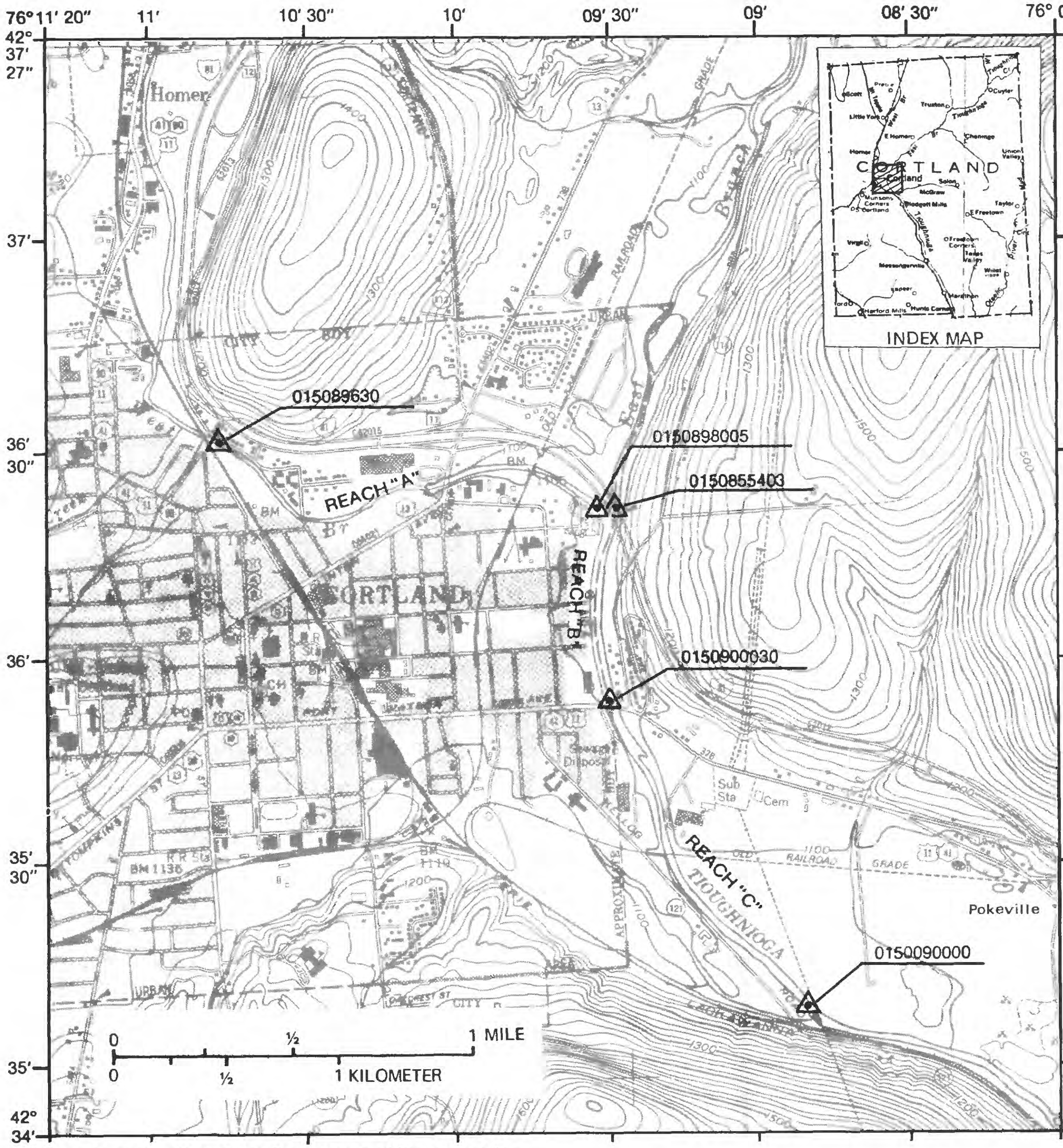
1) This reach of river is the discharge point for most of the ground water moving northeastward through the surficial aquifer in the Otter Creek-Dry Creek basin; and

2) More important, the surficial aquifer "pinches out" along reach B, such that the intersection of the upper surface of the underlying lacustrine unit with the sloping bedrock wall (fig. 3B) forces more upward leakage of ground water through the streambed than would occur without the bedrock wall.

The total ground-water discharge measured on August 31, 1981 over the 2.84-mi reach (the 1.06-mi reach C showed no net gain) of the West Branch and Tioughnioga Rivers was 11.6 ft^3/s, which is in close agreement with Cosner and Harsh's model-simulated value of 11.2 ft^3/s for a 2-mi section of the same reach under similar low-flow conditions.

Streambed Permeability

The aquifer geometry just described indicates that the predominant direction of ground-water flow in the aquifer beneath the river in this reach is vertically upward, with little or no horizontal component of flow. Streambed permeability was estimated through a variation of Darcy's law (Walton, 1962), in which the head difference across the streambed was approximated from the head difference between observation well 09-36 (pl. 1), on the streambank, and the stream-surface altitude recorded at the adjacent gaging



Base from New York State Department of Transportation
1:24,000 quadrangles: Cortland, NY, 1974

Figure 4.--Location of measurement sites in Tioughnioga River seepage investigation, August 31, 1981, near Cortland.

Table 1.--Summary of Tioughnioga River seepage investigation near Cortland, August 31, 1981.

[Site and reach locations shown in fig. 4.]

Reach	Site number	Distance along stream channel from furthest downstream site (mi)	Site location	Topographic drainage area (mi^2)	Measured discharge (ft^3/s)	Length of reach (mi)	Net gain within reach (ft^3/s)	Ground water gain per 1,000 ft of channel (ft^3/s)
	015089630	2.84	West Branch Tioughnioga River at Erie-Lackawanna RR Bridge at Cortland	*	24.6			
A						1.23	6.6	1.0
	0150898005	1.61	West Branch Tioughnioga River at mouth at Cortland	99.6	31.2			
	0150855403	1.61	East Branch Tioughnioga River at mouth at Cortland	192	22.0			
B	Combined flow for stations 0150898005 and 0150855403				53.2	0.55	5.0	1.7
	015090000	1.06	Tioughnioga River at Cortland	292	58.2			
C	0150900030	0	Tioughnioga River above Tioughnioga tributary at Cortland, N.Y.	*	66.5	1.06	0**	0**
	Subtract discharge from municipal sewage-treatment plant				- 8.3			
					58.2			

* Drainage areas not determined.
** A net gain in flow of 8.3 ft^3/s was observed over this reach and was attributed to the discharge of a municipal sewage-treatment plant upstream.

station during the seepage run. For expediency, it is assumed that this total head loss is measured across a streambed thickness of 1.5 ft, as used by Cosner and Harsh (1978). The vertical hydraulic conductivity of the streambed, based on measured seepage over stream reach B, was calculated to be 1.04 ft/d, which is comparable to the value of 1.9 ft/d reported by Haeni (1978) for the Pootatuck River in Connecticut but is two orders of magnitude greater than the value of 0.038 ft/d used by Cosner and Harsh (1978). The value selected by Cosner and Harsh was based solely on model calibration to achieve a desired total stream-leakage value for Otter and Dry Creeks. As such it is dependent on the accuracy of the estimates of other hydrologic variables used as model input.

Aquifer Characteristics

The surficial aquifer is highly permeable, with a hydraulic conductivity of 300 to 350 ft/d as estimated from specific-capacity data from an industrial well near Pokeville in the Tioughnioga River valley (appendix II). These values and the range in saturated thickness indicate that the transmissivity of the surficial aquifer in the Tioughnioga River valley ranges from approximately 7,500 to 28,000 ft/d. Most 6-inch-diameter domestic wells tapping this aquifer are finished open ended (with no well screen); therefore, reported yields are commonly less than 50 gal/min. However, properly screened and developed production wells tapping this unit could be expected to yield at least 500 gal/min (provided the saturated thickness is adequate), as evidenced by the high specific capacity of 44 (gal/min)/ft at a pumping rate of 350 gal/min at the 12-inch-diameter industrial well near Pokeville mentioned above.

The hydraulic-conductivity value obtained for the surficial aquifer in this area is substantially lower than those calculated from three wells tapping the same aquifer in the Otter Creek-Dry Creek valley. On the basis of three aquifer tests, Buller and others (1978) reported transmissivity in the Otter Creek-Dry Creek basin to range from 37,000 to 80,000 ft^2/d, with respective hydraulic-conductivity values ranging from 950 to 1,140 ft/d. These data indicate the surficial outwash aquifer to be somewhat less permeable in the Tioughnioga River valley than in the Otter Creek-Dry Creek basin.

SIMULATION OF GROUND-WATER FLOW

Model Description

A finite-difference ground-water flow model developed by Trescott and others (1976) was used to simulate the response of the surficial outwash aquifer to imposed stresses. The model simulates two-dimensional ground-water flow in response to artifically imposed stresses such as pumping or to naturally occurring stresses such as drought. Given specific values for aquifer characteristics such as hydraulic conductivity and specific yield, the model can be used to simulate water levels that would result under both steady-state (no change in heads with time) and transient-state hydrologic conditions and to calculate changes in water levels that would result from pumping at specific sites. Sources of water may include aquifer storage, recharge from precipitation, inflow across aquifer boundaries, and induced infiltration through leaky streambeds. Water may be discharged through pumping wells, evapotranspiration, and leakage to streams.

The model uses finite-difference techniques to represent the following partial differential equation (Bredehoeft and Pinder, 1970), which describes ground-water flow in two dimensions:

$$\frac{\partial}{\partial x}\left(K_{xx}\, b\, \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_{yy}\, b\, \frac{\partial h}{\partial y}\right) = S_y \frac{\partial h}{\partial t} \pm W(x,y,t) \qquad (1)$$

where: K_{xx} and K_{yy} = principal components of the hydraulic conductivity tensor (Lt^{-1});
h = hydraulic head (L);
S_y = specific yield of the aquifer (dimensionless);
b = saturated thickness of the aquifer (L);
t = time;
x,y = rectangular coordinates along the principal major and minor flow axes; and
$W(x,y,t)$ = volumetric flux of recharge or withdrawal per unit surface of the aquifer (Lt^{-1}).

Trescott and others (1976) provide a detailed description of the theoretical development, finite-difference approximations, and the various solution algorithms used in the model program.

Applying the Trescott program to model the surficial outwash aquifer at Cortland necessitates certain simplifying assumptions to enable simulation of the ground-water system. Those assumptions are:

1. Flow in the surficial outwash aquifer is horizontal, and the aquifer is isotropic in the x,y plane.

2. Recharge to the aquifer from precipitation is constant during any given simulation period.

3. The average stream stage remains constant during any given simulation period.

4. The aquifer can be areally divided into a finite number of rectangular blocks within which the aquifer properties are assumed to be uniform. The center of each block is termed a "node," and water levels calculated for each node are representative of water levels over that block. Aquifer properties may vary from block to block.

5. All pumping wells are considered to be screened through the full saturated thickness of the aquifer and are 100 percent efficient.

Extension of The Cortland Model

The primary goal of this study was to extend the previous two-dimensional ground-water flow model of the Otter Creek-Dry Creek valley of Cosner and Harsh (1978) to include the surficial outwash aquifer that occupies the East

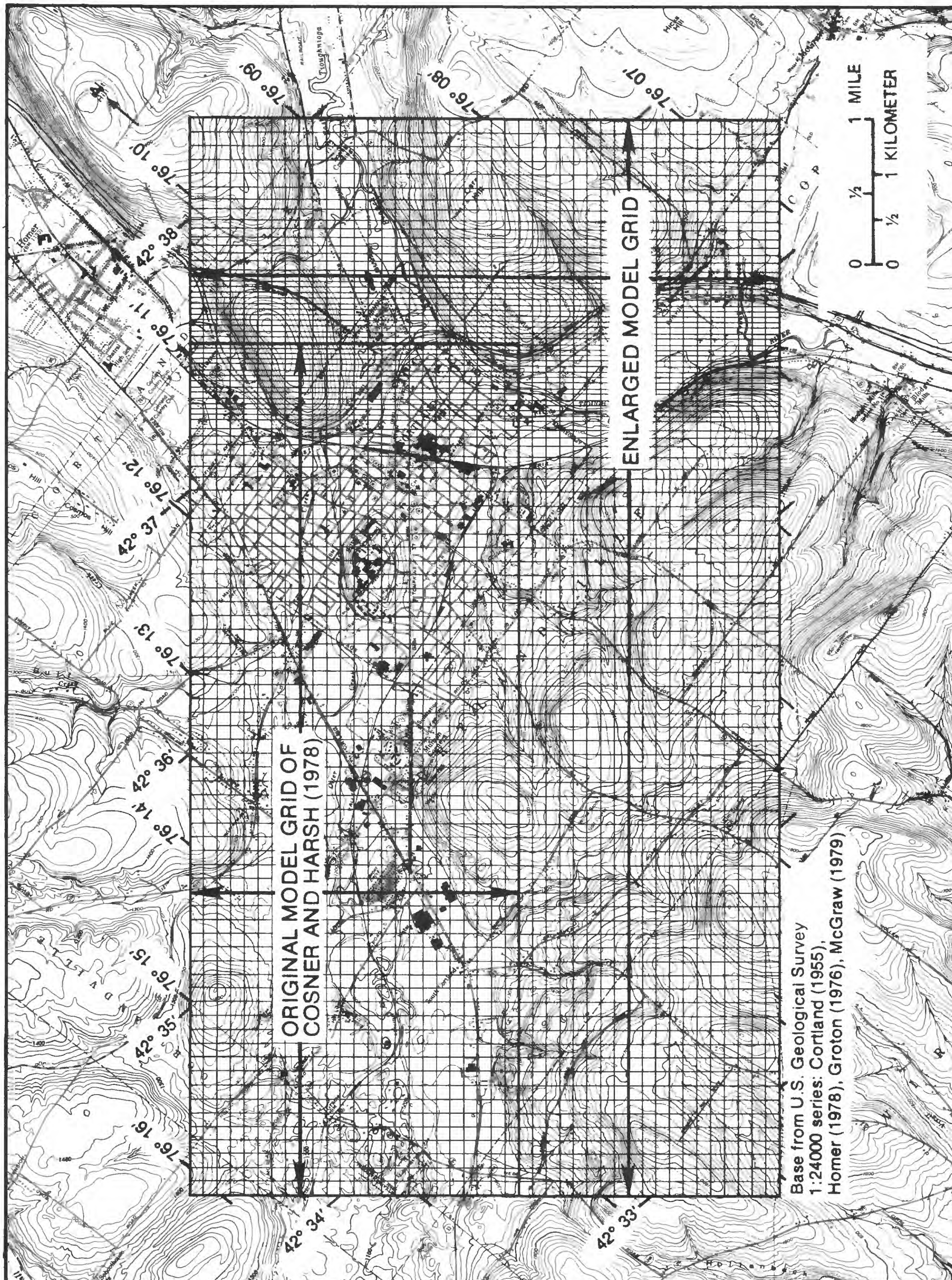


Figure 5.--Comparison of original and extended model grids.

Branch and main Tioughnioga River valleys. The grid orientation used by Cosner and Harsh for the Otter Creek-Dry Creek valley was retained, but the model area was extended east and south into the Tioughnioga River valley. The grid was expanded from the original 24 rows and 62 columns (1,488 nodes) to 43 rows and 99 columns (4,257 nodes), and the actual modeled area was enlarged from 7.9 to 10.3 mi^2. The original grid spacing of 500-ft square blocks was retained over all but the easternmost sixth of the original model area; that part and the extended area was represented by nodes ranging from 500 ft to 200 ft in width (x direction). The relative sizes and areal extent of the old and new grids are shown in figure 5. A grid of this type is known as a block-centered, finite-difference grid with variable spacing.

Initial estimates of the areal distribution of hydraulic conductivity, aquifer-base altitude, and average March water-table altitude were made from test-hole, water-level, well-inventory, and pumping-test data (appendix II) in the model-extension area. In addition, estimates of the distribution of average March river stage in the West Branch, East Branch, and Tioughnioga Rivers and distribution of the altitude of the base of the streambed were made from gage-height records, gage datums, and seepage-run data. These data were coded into data-set arrays for the appropriate nodal blocks representing those sections of aquifer and(or) stream. All other data arrays used in the original Otter Creek-Dry Creek model (Cosner and Harsh, 1978) were retained and incorporated into the enlarged model.

Model Boundaries

Boundary conditions for the enlarged model were selected to simulate actual hydrologic boundaries or were arbitrarily placed so that boundary effects on simulated pumpage would be minimal. Several types of boundaries were used in the model, including constant-flux, constant-head, and no-flow boundaries. Streams were simulated with a leaky confining unit to represent the streambed. The placement of the various types of boundaries in the enlarged model is shown in figure 6.

Constant-flux boundaries

Wherever the model grid intersects one of the tributary river valleys, the grid/valley intersections are treated as constant-flux boundaries. Constant-flux boundaries at these locations simulate the flow of ground water into the modeled area from the East Branch and West Branch Tioughnioga River valleys and the Pokeville-McGraw valley. A constant-flux boundary is also used to simulate the flow of ground water out of the modeled area through the Tioughnioga River valley south of Pokeville. The boundary fluxes were obtained by estimating saturated thickness for each node at a valley cross section at each of these boundaries, then multiplying the cross-sectional areas by the estimated hydraulic conductivity and the average water-table gradient to yield a flux for each node. The resulting fluxes were used as initial values in the model and were later adjusted through calibration.

The constant-flux boundaries used in the Cosner-Harsh model (1978), which simulate ground-water recharge to the outwash aquifer from adjacent kame terraces, were retained, but the southern boundary was extended slightly eastward to simulate recharge from the kame terrace to the model extension (fig. 6).

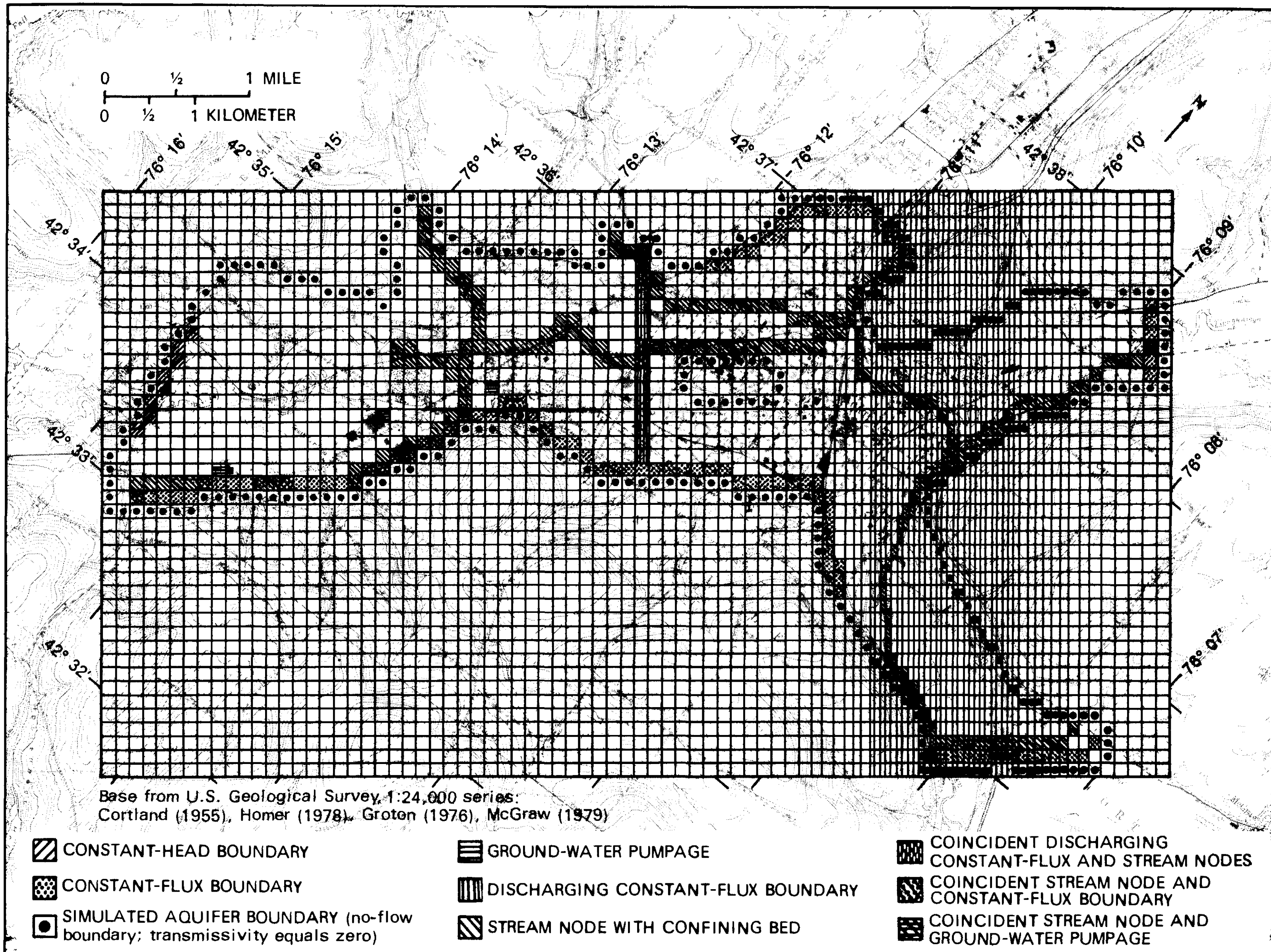


Figure 6.--Type and location of simulated aquifer boundaries and streams in extended model.

The kame terraces are not modeled as aquifer material, but the boundary between them and the outwash aquifer is treated as a constant-flux boundary to account for the additional recharge they provide.

The use of constant-flux boundaries to model ground-water flow into and out of the model is appropriate as long as the effects of pumping do not reach them. Because these fluxes are specified and are not head dependent, any model-calculated drawdown due to simulated pumping near the boundary would be greater than could actually occur. In addition, simulated pumping would not induce a greater flux across the boundary, both because the flux is specified and because a no-flow boundary is positioned behind the constant-flux boundary (fig. 6).

To account for ground water that recharges the lower confined outwash aquifer in an area just west of the city of Cortland, a discharging constant-flux boundary was used. Water leaving the surficial aquifer in the Otter Creek-Dry Creek valley to recharge the lower confined aquifer was represented by a line of fully penetrating pumping wells, perpendicular to the direction of ground-water flow, across the valley in the area assumed to be the recharge zone for the lower aquifer (figs. 2, 6 and pl. 1, 2). Ground water leaving the model and recharging the confined aquifer was estimated during steady-state model calibration to be approximately 3.5 ft^3/s.

Constant-head boundary

The constant-head boundary used by Cosner and Harsh (1978) to simulate the southwest boundary of the model was also retained (fig. 6). A constant-head boundary at this location causes a ground-water divide to form east of the boundary, which causes ground water to flow in two directions--into the modeled area and away from it, toward the constant-head boundary. The use of this constant-head boundary simulates the shift of the actual ground-water divide in response to seasonal changes in recharge. The use of a constant-head boundary here would be inappropriate, however, if attempts were made to simulate large ground-water withdrawals in its immediate vicinity. If such a simulation were attempted, the constant-head boundary would provide an unrealistically large amount of ground water under the stress of heavy withdrawal.

Impermeable boundaries

Impermeable (no-flow) boundaries are used to model the intersection of the surficial outwash aquifer with the till or bedrock surface at the valley sides. The no-flow boundary was selected because the amount of recharge contributed to the aquifer by the till and bedrock is negligible compared to the overall ground-water budget for the aquifer. A horizontal no-flow boundary was used to simulate the bottom of the surficial outwash aquifer, which is taken as the top of the lacustrine clay unit throughout the East Branch, West Branch, and Tioughnioga River valleys in the modeled area. The use of a no-flow boundary here is considered appropriate because the lacustrine unit, owing to its thickness and low hydraulic conductivity, allows little upward leakage to the surficial aquifer. Because little is known about the position or occurrence of the lacustrine surface in the area just west of the city of Cortland, the modeled boundary departs from the known stratigraphic position of this unit in this area and slopes downward to merge with the arbitrary model bottom used by Cosner and Harsh (1978).

Simulation of Stream-Aquifer Relationship

The Trescott two-dimensional model simulates the exchange of water between a stream and an aquifer by comparing the head difference between the aquifer and the stream stage at each stream block. Three possible conditions, each of which can be simulated with the model, are depicted in figure 7.

In example A in figure 7, the simulated stream is gaining; that is, the head in the aquifer beneath the river is above the stream stage, producing upward leakage through the streambed. In this situation, the model uses the following equation to calculate the upward ground-water seepage through the streambed:

$$Q_L = \frac{K_V}{M} \cdot A_S \text{ (RIVER - PHI)} \qquad (2)$$

where: Q_L = leakage over the stream-block area, in ft^3/s;
K_V = vertical hydraulic conductivity of the streambed, in ft/s;
M = thickness of streambed, in ft;
A_S = area of streambed within model block, in ft^2;
RIVER = head in river (river stage), in ft, and
PHI = head in aquifer, in ft.

In example B, the head in the aquifer beneath the river is lower than the stream stage but higher than the altitude of the streambed (or base of the streambed confining unit). This would be expected in naturally losing reaches of a stream or in areas where nearby pumping produces a slight downward hydraulic gradient through the streambed. In this situation, the model uses the same equation to calculate leakage; the only difference is that leakage is downward through the streambed instead of upward.

In example C, the head in the aquifer is below the base of the streambed, which causes a steep downward hydraulic gradient between the river surface and the potentiometric surface of the aquifer. This is the situation that would be expected where production wells adjacent to the river induce infiltration through the streambed by heavy pumping.

For convenience, the program was modified such that if the head in the aquifer falls below the base of the streambed confining unit, the model calculates leakage through the streambed to the aquifer from the head difference between the stream stage and streambed base by the following equation:

$$Q_L = \frac{K_V}{M} \cdot A_S \text{ (RIVER - BASE)} \qquad (3)$$

where: BASE = altitude of streambed base.

This is a reasonable approximation of the real-world relationship between stream and aquifer if it is assumed that:

1. Practically all head loss between the stream and the aquifer occurs across the streambed thickness, M.

2. The vertical hydraulic conductivity of the streambed is several orders of magnitude lower than that of the underlying outwash aquifer.

These assumptions imply that the rate of downward leakage through the streambed would approach some constant value if aquifer heads fall far below stream stage in response to heavy nearby pumping.

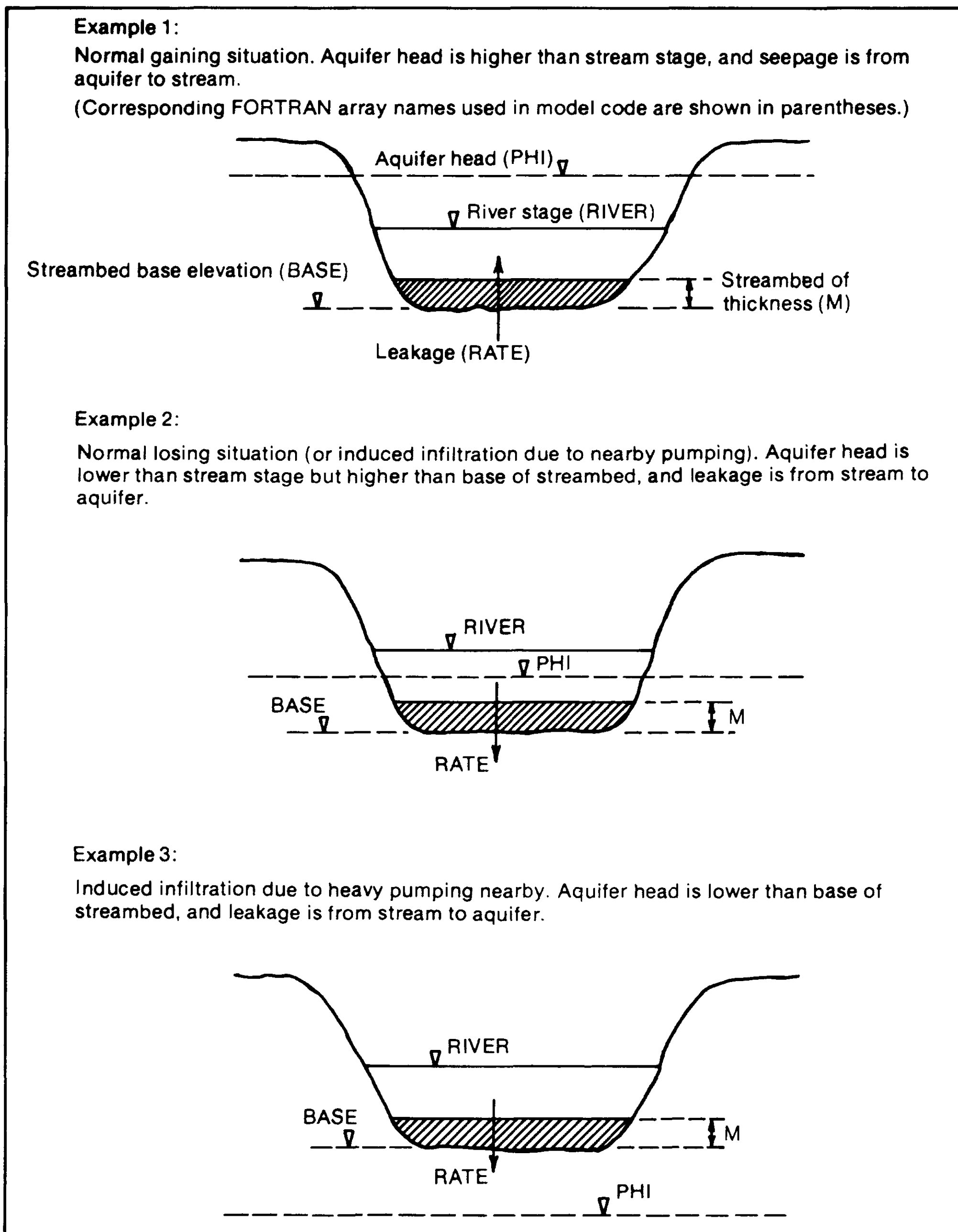


Figure 7.--Idealized relationship between aquifer head and river stage in three typical hydrologic situations.

Although the original Trescott program can be specified to use a similar equation to calculate stream leakage, use of that method for certain field conditions would require some other model approximation that would locally be too coarse. Modifying the program code as described to specify the altitude of the streambed base separately removed that limitation.

This program modification necessitated the input of a new array, termed BASE, which is the altitude of the base of the streambed. This array was compiled by subtracting the estimated river depth plus the streambed thickness, M, from the RIVER array for each stream node.

The program code was also modified to generate values of leakage into or out of constant-head nodes and streambed nodes for each individual node, rather than as a cumulative figure. This modification allows calibration of the model to the seepage-run data.

Modeled Streambed Permeability

The Trescott model approximates leakage between a stream and the underlying aquifer with equation 2 or equation 3, depending upon whether the head in the aquifer is above or below the streambed-base altitude. Because the smallest areal unit that can be simulated is a grid block, and if we use, for convenience, a constant streambed thickness, the leakage equation used in the computer program becomes:

$$Q_L = \left(\frac{\tilde{K}_V}{M_C}\right) \cdot (\text{RIVER} - \text{PHI}) \cdot A_B \qquad (4)$$

where: $\tilde{K}_V$ = equivalent vertical hydraulic conductivity, in ft/s;
M_C = standard streambed thickness, in ft, and
A_B = area of the grid block used to simulate a stream node, in ft^2.

For the leakage calculated from equation 4 to be valid, it requires that

$$\left(\frac{\tilde{K}_V}{M_C}\right) \cdot A_B = \left(\frac{K_V}{M}\right) \cdot A_S$$

That is, that product of the actual "leakance" term (K_V/M) and the actual area of the streambed (A_S) must be equal to the product of the "equivalent leakance" term (K_V/M_C) and the area of the stream grid block. Because the area of the streambed is generally some percentage of the area of the block used to represent it, the equivalent vertical hydraulic conductivity of the streambed (K_V) can be calculated by multiplying the actual streambed hydraulic conductivity (K_V) by the ratio of streambed area to block area for each stream node in the model:

$$\tilde{K}_V = K_V \left(\frac{A_S}{A_B}\right)$$

In doing this, we are assuming that the standard streambed thickness (M_C), in this case 1.5 ft, is equivalent to the actual streambed thickness, which is reasonable in this instance.

To calculate the equivalent vertical hydraulic conductivity of the streambed (K_V) for this model, areas of stream reach within each model block were measured from a 1:12,000-scale enlargement of a U.S. Geological Survey topographic map and expressed as a ratio of stream area to block area for each stream node. These stream-to-block-area ratios for Trout Brook and the Tioughnioga River and its East and West Branches were multiplied by the previously calculated streambed hydraulic conductivity (K_V) of 1.04 ft/d to obtain the equivalent hydraulic conductivity of the streambed (K_V) adjusted for block area.

Modeled streambed hydraulic conductivities for Otter Creek and Dry Creek were based on a model-derived value of 0.038 ft/d (Cosner and Harsh, 1978) and were also recalculated to adjust for differences between stream area and block area. This permitted use of a constant streambed thickness of 1.5 ft for all streams and rivers modeled.

Steady-State Simulation

Before a model can be used to simulate imposed stresses, it must be capable of reproducing observed conditions to an acceptable degree. In this study, a series of steady-state simulations was used to calibrate the model with respect to recharge, discharge, and head distribution during a period when water levels and fluxes in the aquifer were relatively constant. Steady-state conditions are commonly defined as periods during which the ground-water system is in equilibrium; that is, the rate of recharge to the system is equal to the rate of discharge from the system, with no net change in ground-water heads. True steady-state conditions are often difficult to recognize, especially during the limited data-collection periods associated with most ground-water studies. As a result, quasi-steady-state conditions are often simulated in place of true steady-state conditions as a starting point in the model calibration.

Selection of equilibrium period

Cosner and Harsh (1978) used water levels measured on March 5, 1976 to simulate a steady-state condition that represents a period of average above-normal recharge to the Otter Creek-Dry Creek aquifer. The starting heads that they used for a steady-state conditions are thus considered to be quasi-steady state but nevertheless valid. These water levels were the highest recorded that year in the Otter Creek-Dry Creek basin and thus represent a peak in ground-water storage. Because no additional water was being taken into storage at that time, we can assume that the recharge rate was approximately equal to the discharge rate of the system, and that a temporary equilibrium for an "average wet period" had been reached. Cosner and Harsh used this peak water-level distribution on March 5, 1976 for their "near steady state" conditions. Actually, the March 5, 1976 water levels are about 1.5 ft higher than the long-term average for March and 2.8 ft higher than the 36-year mean water level, as shown in figure 8, a 36-year hydrograph of well C-19, and its

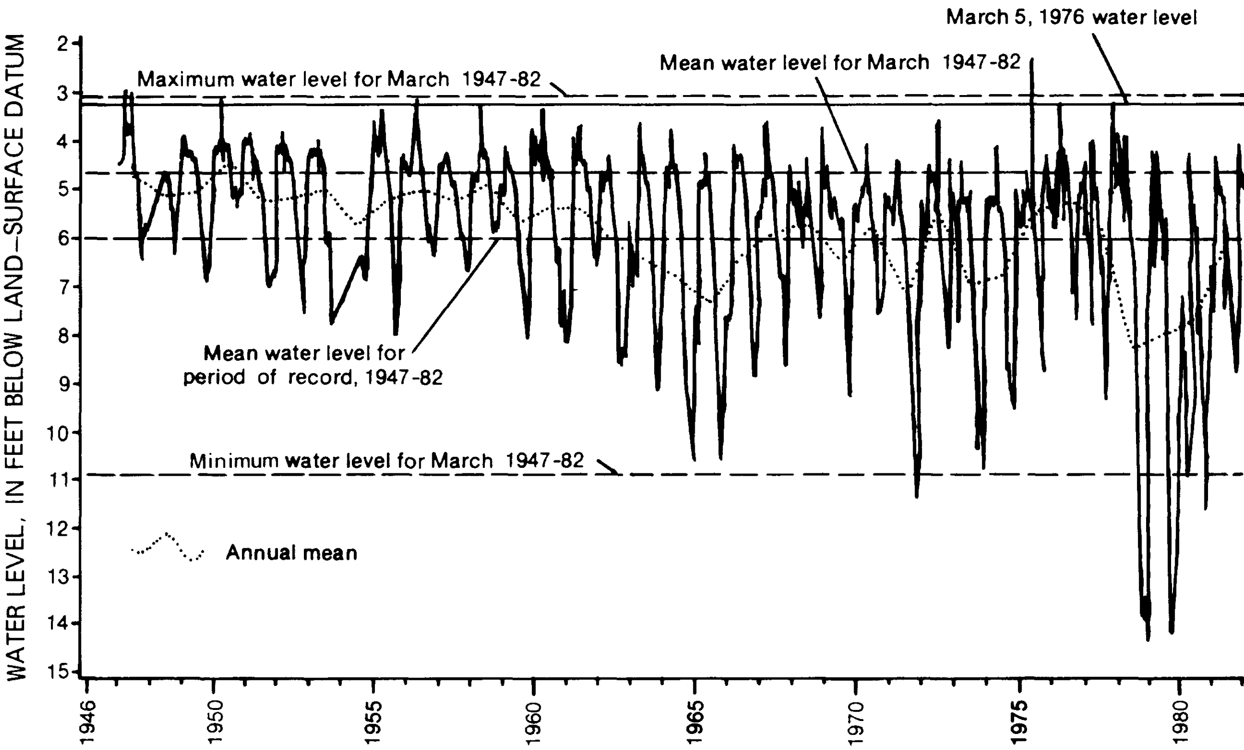


Figure 8.--36-year hydrograph of observation wells C19 at Cortland and its successor well C102. Also indicated are (1) the March 5, 1976 water level, (2) maximum, mean, and minimum March levels for 1947-82, and (3) annual mean and long-term mean levels for 1947-82. (Well locations are shown in fig. 2.)

successor, C-102, both near the municipal well field at Cortland (fig. 2) and screened in the surficial aquifer. Because no March 1976 water-level data in the model-extension area were available, a water-table map of the surficial aquifer in that area was constructed from historical water-level data to represent the average head distribution for a typical March. The objective of calibration was to acceptably duplicate the observed head distribution as defined by the 1976 Cosner-Harsh data and the additional water-level data collected in the previously unmodeled area.

Initial calibration

Steady-state simulations were made from the same areal recharge rate (28 in/yr) over the model as Cosner and Harsh (1978) had used, with adjustments to the areal distribution of hydraulic conductivity, the aquifer-base altitude in the model-extension area, and the amount of water discharged from the surficial aquifer to the confined outwash aquifer. The recharge rate of 28 in/yr represents the average recharge rate for a period of above-normal (spring) recharge. An acceptable water-level match was obtained after minor adjustments to the aquifer-base altitude in selected areas, and with hydraulic-conductivity values ranging from 50 to 350 ft/d for the surficial aquifer. Observed conditions in the expanded model were best matched with a recharge rate to the confined aquifer of 3.5 ft^3/s.

River stage for the steady-state simulation was based on data on river-surface altitudes, water depths at the five measurement sites used in the August 31, 1981 seepage run, and on the stage-discharge relationship for the gaging station on the Tioughnioga River at Cortland. The stage value used (4.0 ft) corresponds to a discharge of about 600 ft^3/s (24 percent flow duration) and approximates the average high-flow conditions that typically occur in March. By comparison, the mean daily flow at the Cortland gage on March 5, 1976 was 3,610 ft^3/s with a corresponding gage height of 7.62 ft; however, this stage is not representative of average March flow conditions.

Accuracy of results

The enlarged model acceptably duplicated the observed data in the Otter Creek-Dry Creek basin while producing an acceptable potentiometric surface in the extended model area. Both the simulated head distribution in the enlarged model area for March 5, 1976 steady-state conditions and the observed data are depicted in plate 2.

Water levels.--Comparison of simulated heads with observed water levels shows that predicted heads differ from measured water levels by values ranging from less than 0.1 to about 6.5 ft. The only exception is along the southwest border of the model, near the constant-head boundary, where the simulated heads are 36 ft too low.

Comparison of simulated heads in the extended model area with March water-level data (pl. 2) shows that predicted heads are generally within 0.5 ft of the measured heads except at well 09-42, near the Tioughnioga River at Cortland, where the predicted head is 2.3 ft lower than the measured head.

Aquifer recharge.--A ground-water budget for March 5, 1976 steady-state conditions in the Cortland surficial outwash aquifer, calculated from model-input data and model-generated data, is given in table 2. The resulting data indicate that, of the total recharge to the surficial aquifer, direct areal recharge from precipitation contributes approximately 58 percent, leakage from Otter Creek and Dry Creek provides 18.6 percent, and adjacent kame terraces account for about 16 percent.

Aquifer discharge.--Of the total discharge from the aquifer under these same steady-state conditions, ground-water leakage to the West Branch, East Branch, and Tioughnioga River, as well as to Trout Brook, accounts for approximately 55 percent, ground-water pumpage in 1976 accounts for 27 percent, and loss to the confined aquifer accounts for about 9.7 percent.

A comparison of the simulated ground-water discharge to the West Branch Tioughnioga and the Tioughnioga Rivers under steady-state conditions with seepage data collected during August 1981 (table 3) reveals some discrepancies. For example, under steady-state spring conditions (March 1976), the model calculated 4.03 ft^3/s of ground-water discharge to the West Branch, reach A, whereas the gain observed during the August 31, 1981 seepage investigation was 6.6 ft^3/s. Similarly, the model predicted a gain of 2.38 ft^3/s in reach B of the Tioughnioga River for March, whereas field observations in August showed a gain of 5.0 ft^3/s. In contrast, the model showed a gain of 2.12 ft^3/s in reach C for March, whereas the seepage investigation in August had detected no gain.

In sand and gravel valley aquifers, highest ground-water levels usually occur in March or April, which produces a corresponding increase in ground-water seepage (increased base flow) to hydraulically connected streams and rivers. The lowest water levels and therefore lowest ground-water seepage rates usually occur from August through October. The total calculated seepage to the West Branch and Tioughnioga Rivers under steady-state March conditions is probably too low, as indicated by comparison with the observed data of August 31, 1981. This discrepancy may be due to some or all of the following factors:

Table 2.--Ground-water budget for surficial outwash aquifer (expanded model) under near steady-state (March 5, 1976) hydrologic conditions.

Source		Rate of yield	
		(Mgal/d)	(ft^3/s)
Recharge to aquifer			
Direct areal recharge (28 in/yr)		13.54	20.95
Constant-flux boundaries			
recharge from kame terraces		3.82	5.91
ground-water flow from valley inlets		1.30	2.02
Constant-head boundaries (ground-water flow into model)		0.10	0.16
Leakage from streams to aquifer			
Otter Creek-Dry Creek		4.34	6.72
Trout Brook		0.25	0.39
	Total	23.35	36.15
Discharge from aquifer			
Pumping wells		6.29	9.73
Leakage from aquifer to rivers			
West Branch Tioughnioga		5.91	9.15
East Branch Tioughnioga		1.78	2.76
Tioughnioga River		4.21	6.51
Trout Brook		1.05	1.63
Leakage from aquifer to Otter Creek and Dry Creek		0.14	0.22
Recharge (ground-water flux) to lower aquifer		2.26	3.50
Constant-head boundary (ground-water flow)		1.36	2.11
Constant flux (ground-water discharge)		0.35	0.54
	Total	23.35	36.15

1. Error in discharge measurements, which may be as much as 5 percent of the measured discharge at each site, could account for wide variability in the estimate of ground-water seepage rates for the three reaches. For example, a 5-percent error in discharge measurements at all five sites could mean that ground-water seepage to reach A could range from 3.8 to 9.4 ft^3/s; seepage to reach B could range from a loss of 0.6 ft^3/s to a gain of 10.6 ft^3/s; and seepage to reach C could range from a loss of 6.2 ft^3/s to a gain of 6.2 ft^3/s.

2. The discrepancy in seepage may also be the result of inaccurate distribution of streambed hydraulic conductivity and (or) streambed thickness and insufficient recharge to certain areas of the model. At present, the river reaches are modeled with a uniform streambed hydraulic conductivity and thickness.

3. The head distribution in the aquifer in areas near the stream is not well known. Areas of high head caused by inhomogeneities in the aquifer would result in greater seepage to the stream.

4. The simulated river stage in some reaches may be too high, causing an underprediction of the amount of ground-water seepage to the stream. River-surface altitudes and mean river depths under steady-state conditions were estimated from limited data.

Table 3.--Comparison of model-calculated steady-state seepage values with observed seepage values under low-flow conditions in West Branch Tioughnioga and Tioughnioga Rivers at Cortland.

[Station locations and reaches are shown in fig.4.]

Reach	Description	Length of reach (ft)	Observed seepage August 31, 1981 (ft^3/s)	Calculated seepage March 5, 1976 (ft^3/s)
A	West Branch Tioughnioga River from station 015089630 to station 015089005	6500	6.6	4.03
B	Tioughnioga River from stations 0150898005 and 0150855403 to station 015090000	2900	5.0	2.38
C	Tioughnioga River from station 015090000 to station 0150900030	5600	0*	2.12

* A net gain in flow of 8.3 ft^3/s was observed over this reach and was attributed to the discharge of a municipal sewage treatment plant.

Future Model Refinemet

Future ground-water studies within the Cortland area should enable further refinement of the model as new data become available. A ground-water model can be only as accurate as the data from which it was assembled, and the process of calibration is often a continuous one. At present, the expanded model is calibrated only for steady-state spring ("average wet") conditions; thus, its usefulness for predicting the effects of pumping is limited to simulations during "normal wet" conditions. To maximize the model's predictive capability, it must be further calibrated with a transient-state simulation to some other hydrologic condition such as a "normal dry" period. Specific aspects in which additional detailed hydrogeologic information could improve the model's predictive accuracy are outlined below.

1. Detailed information on the areal variation in streambed hydraulic conductivity of the West Branch, East Branch, and Tioughnioga Rivers in the modeled area would allow more accurate simulation of rates of ground-water seepage rates to the river. This could be obtained by seepage investigations on these reaches during low-flow periods, measurements of aquifer heads beneath the streambed, and river-stage measurements at several places along the rivers.

2. No information is available on the areal distribution of hydraulic conductivity in the surficial aquifer in the extended model area. As pumping-test and specific-capacity data become available, the hydraulic conductivity matrix should be updated to incorporate new data.

3. Data on water levels in the extended model area are scant. Installation of several shallow observation wells in these valleys, and incorporating them into the established network, would allow periodic collection of water-level data for further calibration.

4. The poorest fit of the extended-model results to the observed data is at the southwestern edge of the original modeled area near the constant-head and southern constant-flux boundaries. Adjustment of constant-flux values, which represent flow from the southern kame terraces, may improve this fit. However, this part of the model is not critical to simulation of pumping schemes that use high-capacity wells because the low transmissivity in this area would preclude actual well construction there.

SUMMARY AND CONCLUSIONS

A previously developed two-dimensional finite-difference model of the Otter Creek-Dry Creek separated aquifer was expanded to incorporate an additional expanse of aquifer within the East Branch Tioughnioga and Tioughnioga River valleys. The modeled area was enlarged from 7.9 mi^2 (24 rows x 62 columns) to 10.3 mi^2 (43 rows x 99 columns), with a finer grid spacing in the added area. Simulation of the original March steady-state head surface produced an acceptable match with the observed data.

Well-inventory and test-drilling data show that a lacustrine unit ranging in thickness from 35 to 95 ft occupies the East Branch Tioughnioga and Tioughnioga River valleys within the study area and separates the surficial outwash aquifer from a deeper confined aquifer. Available data suggest that the average saturated thickness of the surficial outwash in the Tioughnioga River valley ranges from 25 to 80 ft. Limited data suggest that the thickness of the confined outwash aquifer is greatest (75 ft) in the city of Cortland and decreases southward to 30 ft in the Tioughnioga River valley and to 7 ft northeastward in the East Branch Tioughnioga River valley.

A seepage investigation over a 2.84-mi reach of the West Branch, East Branch, and Tioughnioga Rivers at Cortland shows that the highest rate of ground-water gain occurs in areas where the surficial outwash aquifer pinches out beneath the river and adjacent to bedrock hills. A seepage rate of 1.72 ft^3/s per 1,000 ft of river was measured in the vicinity of the gaging station on the Tioughnioga River at Cortland. The seepage-run data indicate a vertical hydraulic conductivity of 1.04 ft/d for the riverbed.

In a model simulation, a hydraulic-conductivity value of 350 ft/d for the surficial outwash aquifer in the Tioughnioga River valley provided an acceptable water-level match. This value is consistent with a hydraulic conductivity of 330 ft/d obtained by analysis of specific-capacity data from a production well.

The model program was revised with respect to the manner in which leakage from the river to the underlying aquifer is calculated. The changes allow the program to read a new array--the streambed-base altitude (BASE array)--and to use this array to calculate leakage through the streambed whenever the aquifer head falls below the streambed base. This modification should result in a more accurate simulation of induced infiltration from high-capacity pumping wells.

Further calibration of the model with respect to areal distribution of recharge and distribution and magnitude of seepage rates to the various river reaches would be needed before any serious attempt is made to simulate pumping schemes involving induced infiltration.

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APPENDIX 1. SELECTED WELL AND TEST-HOLE LOGS

(Locations are shown in pl. 1)

Wells listed in this appendix are identified on plate 1 by a hyphenated, four-digit number representing the location of the well in seconds of latitude and longitude. For example, the well at 42°35'20" latitude 76°08'40" longitude is shown as well 20-40 on plate 1.

Colors referred to in sample descriptions and shown in parentheses, for example (5Y 6/2), are taken from the Rock-Color Chart (Goddard and others, 1970) distributed by the Geological Society of America.

U.S. Geological Survey Test Hole at location 4235 20 7608 40

Drilled with U.S. Geological Survey hollow-stem auger rig July 8, 1980. Log based on field examination of drill cuttings and cores by R. J. Reynolds. Depths are in ft below land surface. Land surface is approximately 1094 ft above sea level. Finished 2-in.-diam. PVC observation well screened from 15 to 19 ft below land surface with 2-in.-diam. x 4 ft, 0.008-in. slot PVC screen. Measuring point is top of coupling, 1097.45 ft above sea level. Depth to water at time of drilling approximately 9 ft below land surface.

Depth interval	Materials penetrated
0 - 12 ft	Sand and gravel; sand medium to coarse, gravel up to 1 1/2 in., saturated at approximately 9 ft.
12 - 15 ft	Sand and gravel, dark yellowish brown, interbedded with small lenses of gray silt, gravel to 3/4 in., some larger; sorting fair, some iron staining evident.
15 - 20 ft	Sand and gravel, dark yellowish brown, sorting fair, size ranges from fine sand to 1-in. gravel. Some exotic glacial material (granite, gneiss, quartzite pebbles).
20 - 25 ft	Medium gravel with some (5 to 10%) gray clay intermixed. Occasional stones to 1 in., mostly local material (shale, limestone, siltstone, sandstone).
25 - 28 ft	Fine to medium sand with silt.
28 - 35 ft	Medium to coarse sand, poorly sorted, with medium gravel to 3/4 in., intermixed with 10% olive-gray clay. Clay binds gravel together but washes out easily. Gravel consists of mostly local material (black shale, green siltstone, red sandstone) with occasional exotics (granite and quartzite).
35 - 40 ft	Coarse to very coarse sand with gravel to 2 in., poorly sorted, mixed with 10 to 15% grayish-olive clay and silty sand. Gravel is predominantly well rounded red sandstone, black shale, green siltstone.
40 - 47 ft	Coarse to very coarse sand with gravel to 3/4 in., poorly sorted. Gravel is well rounded and contains of about 50% local material (red sandstone, shale, siltstone, limestone) and 50% exotics (gneisses, granites, and quartizite).
47 - 50 ft	Clay, brownish-red, mixed with 1-in. to 2-in. stones of shale and limestone. Stones are very weathered.
50 - 55 ft	Clay, varved, with thin interbeds of silt, pale olive gray (5Y 6/2). Some ice-rafted pebbles. (Lacustrine material.)
55 - 99 ft	Variable silty clay, varved, pale olive-gray (5Y 6/2), with occasional ice-rafted pebbles and occasional thin lenses of fine to very fine sand. (Bottom of test hole - limit of drilling capability.)

U.S. Geological Survey Test Hole at location 4235 28 7609 21

Drilled with U.S. Geological Survey hollow-stem auger rig July 9, 1980. Log based on field examination of drill cuttings and cores by R. J. Reynolds. Depths are in ft below land surface. Land surface is approximately 1096 ft above sea level. Finished 2-in.-diam. PVC observation well screened from 20 to 24 ft below land surface with 2-in.-diam. x 4 ft, 0.008-in. slot PVC screen. Measuring point is top of coupling, 1095.94 ft above sea level. Depth to water at time of drilling approximately 10 ft below land surface.

Depth interval	Materials penetrated
0 - 10 ft	Clayey soil, coarse gravel, cobbles, well-rounded stones to 3 in. Saturated at approximately 10 ft.
10 - 15 ft	Clayey-silty sand and gravel, with some medium to fine sand. Pebbles rounded to subrounded to 1/2 in. diam.
15 - 20 ft	Gravel, up to 1/2 in., mixed with medium sand and some clay. Gravel is local material, red sandstone (heavily oxidized), shale, siltstone.
20 - 25 ft	Coarse to very coarse sand with medium (1/2 in.) gravel and some clay. Sand is angular to subangular.
25 - 30 ft	Coarse to very coarse sand, fairly well sorted, mixed with 10% to 15% olive-gray clay on top, grading into heavy gravel and cobbles mixed with some clay on bottom. Gravel and cobbles mostly shale, red sandstone, siltstone - very difficult drilling.
30 - 35 ft	Interbedded coarse to very coarse sand and gravel and cobbles up to 4-in. diam., mixed with 10 to 15% olive-gray clay.
35 - 40 ft	Medium to coarse sand with some clay intermixed, interbedded with lenses of slightly sandy olive-gray clay (5Y 6/1).
40 - 43 ft	Medium to fine, well-sorted sand.
43 - 50 ft	Clay, varved, silty, compact, pale olive-gray (5Y 6/2). (Lacustrine material.)
50 - 99 ft	Clay, varved, variably silty, very compact, pale olive-gray (5Y 6/2). (Bottom of test hole - limit of drilling capability.)

U.S. Geological Survey Test Hole at Location 4235 42 7609 17

Drilled with cable tool rig May 8-9, 1980. Log based on field examination of cuttings by R. J. Reynolds. Depths are in ft below land surface. Land surface is approximately 1102.5 ft above sea level. Finished 6-in.-diam. observation well "B" screened 35 to 40 ft and 45 to 50 ft with two 4-in.-diam. x 5-ft-long PVC screens, 0.010 in. slots, connected with 5 ft of blank PVC pipe. Measuring point is top of inside casing, 1105.06 ft above sea level. Static water level at completion approximately 16 ft below land surface. Well is about 330 ft from river and is used in conjunction with well 4235 42 7609 22 to determine aquifer diffusivity.

Depth interval	Materials penetrated
0 - 5 ft	Coarse to very coarse sand with medium gravel to 1 in. diam. and approximately 10% light olive-gray clay (5Y 5/2) and silt mixed throughout. Sorting is fair, sand is mostly shale, limestone, red sandstone, tan sandstone, green siltstone, and an abundance of feldspar and quartz grains. Pebbles to 1/2 in. are rounded.
5 - 10 ft	Coarse to very coarse sand, with medium gravel, occasional stones to 2 in. with gray clay (5Y 5/1) mixed throughout. Sand and gravel consists predominantly of shale, limestone, and some red and tan sandstone, with minor amounts of feldspar and quartzite.

U.S. Geological Survey Test Hole at location 4235 42 7609 17 (cont.)

Depth interval	Materials penetrated
10 - 15 ft	Coarse to very coarse sand (70%), with fine to medium gravel (10%), occasional stones to 1 in., medium sand (5%) and mixed with gray clay (15%).
15 - 20 ft	Medium to coarse sand (60%) with some fine sand (20%), intermixed with olive-gray clay (20%). Occasional rounded stones to 3/4 in. diam.
20 - 25 ft	Medium to coarse sand (65%) with some very coarse sand (10%), mixed with olive-gray clay (20%). Occasional rounded stones to 1 in. diam. Mostly shale and limestone grains with minor amounts of quartz and feldspar.
25 - 30 ft	Coarse sand (70%), fairly well sorted, with some medium sand (10%), gravel (5%) and mixed with olive-gray clay (15%). Stones to 1 in. diam.
30 - 35 ft	Medium sand (65%), well sorted, with some coarse sand (20%), fine gravel (5%), and mixed with olive-gray (5Y 5/2) clay (10%).
35 - 40 ft	Medium sand (60%), with coarse to very coarse sand (30%) and olvie-gray clay mixed throughout (10%).
40 - 45 ft	Coarse to very coarse sand (60%), with medium sand (20%), some medium gravel and stones (10%), and olive-gray clay (10%) throughout.
45 - 52 ft	Coarse to very coarse sand, well sorted (75%), some medium sand and fine gravel (15%), mixed with brownish clay (10%) throughout.
52 - 57 ft	Silty, clayey medium sand. Approximately 50% silt and clay and 50% medium sand. Clay is dark yellowish brown (10 YR 4/2) and is compact.
57 - 60 ft	Variably silty clay, compact, medium olive-gray (5Y 4/2) with occasional ice-rafted stones. (Lacustrine material.)

U.S. Geological Survey Test Hole at location 4235 42 7609 22

Drilled with cable tool rig May 5, 1980. Log based on field examination of cuttings by R. J. Reynolds. Depths are in ft below land surface. Land surface is approximately 1099.5 ft above sea level. Finished 6-in.-diam. observation well "A" screened from 30 to 35 ft and from 40 to 45 ft below land surface with two 4-in.-diam. x 5 ft long PVC screens, 0.010 in. slots, connected with 5 ft of blank PVC pipe. Measuring point is top of inside casing, 1102.16 ft above sea level. Static water level at completion approximately 12 ft below land surface but fluctuates freely with river stage. Well is located approximately 30 ft from the Tioughnioga River and is used in conjunction with well "B" (4235 42 7609 22), located 300 ft east, to determine aquifer diffusivity.

Depth interval	Materials penetrated
0 - 2 ft	Artificial fill - predominantly clayey soil, sand, gravel, and cobbles.
2 - 5 ft	Coarse sand with gravel (85%), cobbles (5%), and clay (10%) intermixed. Clay is light olive-gray (5Y 5/2). Sand composed of mostly shale, sandstone, and limestone.
5 - 10 ft	Very coarse sand (60%), with some coarse sand (15%), and some gravel (15%), intermixed with olive-gray clay (10%). Mostly local material but some feldspar and quartzite.
10 - 15 ft	Medium to coarse sand (40%), with very fine sand and silt (40%), olive-gray clay (15%), and very coarse sand (5%).
15 - 20 ft	Medium to coarse sand (60%), with some fine to very fine sand (15%), and fine to medium gravel (20%), intermixed with olive-gray clay (5%). Occasional stones to 1 1/4-in.-diam. Slightly greater amount of "exotics" (feldspars, quartzite).

U.S. Geological Survey Test Hole at location 4235 42 7609 22 (cont.)

Depth interval	Materials penetrated
20 - 40 ft	Coarse to very coarse sand (70%), with fine to medium gravel up to 1/4 in. (25%) and little clay (<5%). Some stones to 1 in. diam.
40 - 47 ft	Coarse sand (95%) with a trace of very coarse sand (3-4%), and trace of clay (<1%). Clay color change to pale yellowish brown (10YR 6/2) noted at 42-43 ft.
47 - 60 ft	Clay, silty and slightly sandy, olive-gray (5Y 4/2), very sticky. Lacustrine material - upper few feet mixed with some very coarse sand and/or fine gravel.

U.S. Geological Test Hole at location 4236 09 7609 36

Drilled with cable tool rig November 6-9, 1979. Log based on field examination of cuttings by R. J. Reynolds. Depths are in ft below land surface. Land surface is approximately 1099.9 ft above sea level. Finished 6-in.-diam. observation well "A" screened from 24.5 ft to 29.5 ft below land surface with a single 5-in.-diam. x 5-ft-long, 0.010 in. slot, PVC screen. Measuring point is top of inside casing, 1102.38 ft above sea level. Static water level approximately 10 ft below land surface at completion but fluctuates freely with river stage. Well approximately 20 ft from the Tioughnioga River, adjacent to gage house, and is used in conjunction with well "B" (4236 09 7609 42), located 500 ft west, to determine aquifer diffusivity.

Depth interval	Materials penetrated
0 - 5 ft	Clayey soil, some gravel, construction debris, cobbles.
5 - 10 ft	Coarse to very coarse sand with gravel, occasional stones to 1-in.-diam., mixed with gravel (5%), light-brown clay (10 YR 2/2). Sand is predominantly local shale, sandstone, and siltstone with some medium gravel (15%) and intermixed with brown clay (15%). Poorly sorted.
15 - 20 ft	Medium to coarse sand (70%) with some very coarse sand to fine gravel (15%), fine sand (10%), and brownish-tan clay (5%).
20 - 30 ft	Coarse to very coarse sand (70%) with medium to fine sand (15%), brownish-tan clay (10%), and trace of gravel (5%). Material mostly local shale with minor amounts of red sandstone, feldspar, and quartzite.
30 - 37 ft	Medium to very coarse, well sorted sand, with little brownish-tan clay intermixed.
37 - 44 ft	Clay, thick, gray, intermixed with very coarse sand with gravel. Clay is medium gray (5Y 5/2). Occasional shale chips to 1 in. diam. (Till? - RJR).
44 - 45.5 ft	Bedrock, fissile gray shale.

U.S. Geological Survey Test Hole at location 4236 09 7609 42

Drilled with cable tool rig November 12-15, 1979. Log based on field examination of cuttings by R. J. Reynolds. Depths are in ft below land surface. Land surface is approximately 1101.8 ft above sea level. Finished 6-in.-diam. observation well "B" screened from 29.5 to 34.5 below land surface with a single 5-in.-diam. x 5-ft-long, 0.010 in. slot, PVC screen. Measuring point is top of inside casing, 1103.54 ft above sea level. Static water level approximately 14 ft below land surface at completion, but fluctuates freely with river stage. Well is 500 ft west of well "A" (4236 09 7609 36) and used in conjunction with well "A" to determine aquifer diffusivity.

U.S. Geological Survey Test Hole at location 4236 09 7609 42 (cont.)

Depth interval	Materials penetrated
0 - 5 ft	Very coarse sand with gravel, some clay (10%). Clay is yellowish brown (10 YR 5/4). Mostly local material, (shale, sandstone) with some red sandstone and quartzite.
5 - 10 ft	Gravel, with very coarse sand, and clay. Occasional gravel to 1-in.-diam. Shale, siltstone, red sandstone, and some feldspar and quartzite particles. Clay (15%) is yellow brown (10 YR 5/4).
10 - 15 ft	Coarse to very coarse sand with gravel and clay. Poorly sorted, gravel to 1/2-in.-diam. Clay is 10 YR 5/4, slow drilling.
15 - 20 ft	Coarse gravel with very coarse sand, little clay. Stones to 2-in.-diam., much less clay (10 YR 5/4) than in overlying sections. Poorly sorted.
20 - 25 ft	Medium to coarse sand with some gravel to 1/2-in.-diam., and some clay (10 YR 5/4). Local material (shale, red sandstone, tan sandstone) with occasional quartzite.
25 - 30 ft	Coarse to very coarse gravel with coarse to very coarse sand. Stones to 2-in.-diam., some clay (10 YR 5/4).
30 - 35 ft	Medium to coarse sand with some gravel, some clay, no stones. Sand consists of fairly well sorted shale particles with some red sandstone and minor amounts of quartz.
35 - 40 ft	Coarse to very coarse sand with gravel, stones to 1-in.-diam., trace of clay. Fairly well sorted.
40 - 45 ft	Medium to coarse sand with medium gravel, little clay.
45 - 50 ft	Medium sand (60%) with some fine sand (15%), some coarse sand (15%), and gravel and clay (10%). Bottom of test hole.

U.S. Geological Test Hole at location 4236 35 7609 26

Drilled with U.S. Survey hollow-stem auger rig July 10, 1980. Log based on field examination of cuttings and cores by R. J. Reynolds. Depths are in ft below land surface. Land surface is approximately 1098.5 ft above sea level. Finished 2-in.-diam. steel observation well "A" screened from 14 to 18 ft and 20 to 24 ft below land surface with two 2-in.-diam. x 4-ft-long PVC screens, 0.008 in. slot, separated by 2 ft PVC blank. Measuring point is top of coupling below grade, 1098.36 ft above sea level. Static water level approximately 13 ft below land surface at completion but fluctuates freely with river stage. Well approximately 30 ft from the East Branch Tioughnioga River, used in conjunction with well "B" (4236 36 7609 29) 200 ft west, to determine aquifer diffusivity.

Depth interval	Materials penetrated
0 - 5 ft	Coarse stream gravel to 1-in.-diam., cobbles, gravel, and clay (fill).
5 - 10 ft	Gravel, cobbles, clay, poorly sorted.
10 - 11 ft	Clay stringer, gray.
11 - 13 ft	Coarse gravel.
13 - 18 ft	Coarse to very coarse sand, with fine gravel, intermixed with pale brown (5YR 5/2) clay.
18 - 25 ft	Clayey sand and gravel, cobbles to 3-in.-diam. Clay is sticky, binds material together tightly.
25 - 27 ft	Medium sand, well sorted.
27 - 32.5 ft	Gray clay, hard with some small angular gravel imbedded. Clay is plastic, medium-light gray (N6). (Till-RJR).
32.5- 33 ft	Bedrock, gray shale.

Driller's Log of Well at location 4237 33 7609 01

Drilled in 1965 with cable tool rig to 159 ft; redrilled at later date by Randolph Well and Pump Co., McLean, N.Y. to 209 ft. Depths are in ft below land surface. Land surface is approximately 1105 ft above sea level. Static water level at completion approximately 8.5 ft below land surface. Appears in Bulletin 69 (Randall, 1972) as 4237 34 7608 58; Cortland Asphalt Co. Well No. 1.

Depth interval	Materials penetrated
0 - 4 ft	Gravel
4 - 10 ft	Sand, loam
10 - 20 ft	Sand
20 - 30 ft	Sand
30 - 35 ft	Gravel, saturated
35 - 38 ft	Gravel
38 - 72 ft	Gray clay (lacustrine unit-RJR)
72 - 79 ft	Gravel (lower outwash-RJR)
79 - 94 ft	Clay and hardpan (Till-RJR)
94 -123 ft	Gray shale
123 -159 ft	Gray and brown shale
159 -209 ft	Redrill-no log available

Driller's Log of Well at location 4236 04 7610 34

Drilled with cable tool rig August 14-16, 1979 by Parry Well Drilling Inc., Syracuse, N.Y. for Cortland County Office Building. Depth are in ft below land surface. Land surface is approximately 1118 ft above sea level. Finished 6-in.-diam. well screened from 92.6 to 101 ft with a 10 ft Johnson Everdure Screen, 0.010 in. slot. Static water level at completion approximately 15 ft below land surface.

Depth interval	Materials penetrated
0 - 5 ft	Fill
5 - 16 ft	Clay and gravel
16 - 55 ft	Sand and fine gravel
55 - 85 ft	Clay, blue (lacustrine unit-RJR)
85 -102 ft	Medium gravel and sand (lower outwash-RJR)

Driller's Log of Well at location 4237 26 7609 25

Drilled with cable tool rig April 21, 1969 by Randolph Well and Pump Co., McLean, N.Y., for Sun Pipe Line Co., Cortland, N.Y. Depths are in ft below land surface. Land surface is approximately 1140 ft above sea level. Finished 6-in.-diam. well is open ended. Static water level at completion approximately 20 ft below land surface.

Depth interval	Materials penetrated
0 - 10 ft	Hardpan and gravel
10 - 17 ft	Sandy gravel
17 - 19 ft	Sandy gravel
19 - 25 ft	Gravel
25 - 28 ft	Sandy gravel
28 - 39 ft	Sand and gravel

Appendix I

Driller's Log of Well at location 4237 16 7609 22

Drilled with cable tool rig April 26, 1972 by Randolph Well and Pump Co., McLean, N.Y., for Yellow Lantern Campground, Cortland, N.Y. Depths are in ft below land surface. Land surface is approximately 1100 ft above sea level. Finished 6-in.-diam. well is open ended. Static water level at completion approximately 5 ft below land surface.

Depth interval	Materials penetrated
0 - 10 ft	Topsoil
10 - 22 ft	Sandy clay and gravel
22 - 25 ft	Cemented gravel
25 - 30 ft	Coarse gravel and some coarse sand. Clay at 30 ft.

Driller's Log of Well at location 4235 12 7608 03

Drilled with cable tool rig September 2-3, 1969 by Randolph Well and Pump Co., McLean, N.Y. for Cortland Asphalt, Pokeville, N.Y. Depths are in ft below land surface. Land surface is approximately 1090 ft above sea level. Finished 6-in.-diam. well probably open ended. Static water level at completion approximately 8 ft below land surface.

Depth interval	Materials penetrated
0 - 2 ft	Fill (coarse gravel, cobbles)
2 - 18 ft	Gravel, silt
18 - 23 ft	Sand and gravel
23 - 25 ft	Fine sand with gravel
25 - 30 ft	Coarse gravel, stones
30 - 40 ft	Fine sand with coarse gravel

Driller's Log of Well at location 4237 37 7608 59

Drilled with cable tool rig April 19-21, 1969 by Randolph Well and Pump Co., McLean, N.Y. for Hall and Stearns Co., Loring Crossing, N.Y. Depths are in ft below land surface. Land surface is approximately 1105 ft above sea level. Finished 6 in. diam. well screened approximately 39 to 43 ft below land surface. Static water level at completion approximately 8 ft below land surface.

Depth interval	Materials penetrated
0 - 8 ft	Fill
8 - 20 ft	Gravel, some water
20 - 30 ft	Sandy gravel, water
30 - 38 ft	Sand, loose gravel
38 - 41 ft	Sand and gravel
41 - 43 ft	Gravel

APPENDIX II. PUMPING-TEST ANALYSIS

A 12-inch-diameter production well to supply industrial cooling water was installed and pump tested at the ETL Testing Laboratories facility in the Pokeville Industrial Park at Cortland, N.Y., on November 11, 1980. The test was conducted by Cato Well & Pump Co. The well taps the surficial glacial-outwash aquifer, which extends throughout the Cortland area. In the test locality, the surficial sand and gravel aquifer is approximately 56 ft thick, with about 44 ft of saturated material. The well is equipped with 15 ft of 50-slot (0.050 in.) screen and screened from 40 to 55 ft below land surface. Approximate static water level was 12 ft below land surface. A complete log was available for the test hole in which a 2-in. PVC observation well was installed approximately 110 ft from the pumped well. The observation well was screened (0.040 slot) from 13 to 40 ft (the saturated zone above the production-well screen).

Geologic Logs

The geologic log of the ETL test hole is similar to logs of two test holes drilled nearby by the U.S. Geological Survey. The test hole drilled by the consultant at ETL encountered variable silty sand and gravel from land surface to 56 ft, then lacustrine silt and clay from 56 to 106.5 ft (bottom of hole). The thickness of lacustrine material corresponds to two additional U.S. Geological Survey test holes drilled to about the same depth at a site on the west side of the Tioughnioga River (test hole 28-21T, pl. 1) and at a site approximately 0.6 mi south in an abandoned gravel pit (test hole 20-40T, pl. 1).

The upper aquifer is essentially a water-table aquifer, but under pumping stress behaves as a leaky artesian system. At the ETL test hole, a 9-ft-thick silty zone of lower hydraulic conductivity lies from about 15 to 23 ft below land surface. This silty zone probably is also present at the pumped well because it also appears in the logs of ETL wells A and B (wells 42-22 and 42-17 respectively, pl. 1) as a nebulous zone of slightly finer material from 15 to 25 ft. This silty zone acts as a poor confining unit (aquitard), creating a leaky artesian condition during short pumping intervals. This was seen in the plot of raw drawdown data from a step pumping test of the ETL production well. After each pumping-step increase, drawdowns leveled off rapidly to some constant level, except during an 8-hour period in which the well was pumped at 300 gal/min, when drawdowns show relatively slow adjustment to the newly imposed stress. One possible explanation for the slow decline in head for 8 hours during the 300-gal/min phase is that this was the first time the cone of depression was lowered into this silty zone, and its slower gravity drainage caused the well to take longer to reach equilibrium.

Analysis

Several methods were used to analyze the data for acceptable values of transmissivity (T) and (or) specific yield (S). Measurement accuracy (reported to 0.1 ft for the pumped well) made interpretation difficult. A semilog plot of time versus drawdown was analyzed through an equation given by Lohman (1972, p. 21, eq. 56) in an attempt to obtain acceptable values. Drawdown did not increase steadily during the test but rather increased rapidly after each change in pumping rate to some new level in a stepwise fashion. The rate of decline in head after each new pumping stress was imposed was approximately the same, 2.05 to 2.45 ft per log cycle. A semilog solution to these fractions of the plot yield hydraulic-conductivity values between 40 and 100 ft/d, which is considered unrealistically low.

Because the test used a step-drawdown technique, conventional analysis was not possible. Attempts were made to evaluate the screen-loss factor (Lennox, 1966; Sheahan, 1971) through step-drawdown test techniques. Results were inconclusive but suggest that screen loss, especially with 50-slot screen, was low (less than 1 ft) and that the effects of partial penetration would far outweigh any screen loss.

Because the geologic conditions suggest a leaky artesian situation over short periods of time, an attempt was made to match a log-log plot of drawdown data in the first 10 min of pumping at 100 gal/min to leaky-artesian type curves presented in Walton (1962). A close fit was obtained to the curve (r/B = 2.5, indicating a very leaky condition); however, the calculated T was unreasonably small.

A plot of recovery data in the pumped well on a semilog paper (Theis recovery method, in Jacob, 1963) showed two distinct slopes of recovery data. The early rapid recovery data formed a slope of 4 ft per log cycle, which corresponded to a T of 23,100 (gal/d)/ft (3,088 ft^2/d) or a hydraulic conductivity (K) of 70 ft/d. This seems rather small; however, the latter part of the curve, which comes closer to the origin, has a slope of 0.8 ft per log cycle, which corresponds to a T of 115,500 $(gal/d)/ft^2$ (15,440 ft^2/d) or a K of 350 ft/d. This value is much more realistic and is in agreement with Cosner's data (1978) and in close agreement with transmissivity estimated from specific capacity.

Transmissivity from Specific Capacity

Because log-log and semilog solutions did not give acceptable results, a solution based on specific capacity of the pumped well was sought. At the end of the 24-hour test, the drawdown had stabilized at 8 ft with the pumping rate at 350 gal/min. The unadjusted specific capacity (Q/S) therefore would be 43.75 (gal/min)/ft; however, because the well is partially penetrating, an upward adjustment of specific capacity is necessary. The first step was to make these adjustments for partial penetration.

Turcan (1963) presents equations and graphs that can be used to calculate the theoretical specific capacity of a well from the percentage of the aquifer thickness that the well screens. The formula uses what Turcan (1963) calls a productivity ratio (PR), defined as:

$$Y_a = Y[PR] \tag{5}$$

where: Y_a = observed (field) specific capacity in $(ft^3/s)/ft$ of drawdown,
Y = theoretical specific capacity of a 100-percent efficient, fully penetrating well, and
PR = productivity ratio, which is equal to:

$$PR = \left[p\left(1 + 7\sqrt{\frac{rw}{2\,pm}} \cdot \cos\frac{p\pi}{2}\right)\right] \tag{6}$$

where: p = ratio of screen length to the full saturated thickness of the aquifer, in percent,
rw = radius of pumped well, in ft, and
m = saturated thickness of aquifer, in ft.

By rearranging terms we obtain:

$$Y = \frac{Ya}{\left[p\left(1 + 7\sqrt{\frac{rw}{2pm}} \cdot \cos\frac{p\pi}{2}\right)\right]} \tag{7}$$

Equation 7 was used to calculate a theoretical specific capacity of 67.55 (gal/min)/ft for a fully penetrating well at this site.

Hurr (1966) presents a method of estimating transmissivity from specific capacity data that accounts for the duration of pumping and the aquifer specific yield. The specific capacity calculated from equation 7 was used to estimate transmissivity as follows: first, an apparent specific yield of 0.17 was estimated from figure 4 in Hurr (1966) from the lower of the two transmissivity curves presented. Next, the value of {u[W(u)]}/Sa was determined from the relationship:

$$\{u[W(u)]/Sa\} = 1.632 \times 10^{-2}\ (r^2 s/tQ) \qquad (8)$$

where: r = radius of drawdown measurement, in ft (in this case, assumed to be 1 ft),
s = drawdown in well, in ft (in this case drawdown based on theoretical specific capacity was used),
t = time since pumping started, in fraction of a day, and
Q = pumping rate, in gal/min.

A value of $\{u[W(u)]/Sa\} = 2.415 \times 10^{-4}$ was calculated. This value and the previously estimated value of apparent specific yield were used in fig. 2 of Hurr (1966) to find the corresponding value of u/Sa, which was 1.7×10^{-5}. The value of u/Sa was then substituted into the following equation to calculate transmissivity:

$$T = \frac{1.87r^2}{t(u/Sa)} \qquad (9)$$

where: T = transmissivity, in $(gal/d)/ft^2$;
r = radius of measurement, in this case assumed to be 1 ft; and
t = time of measurement since pumping started, in fraction of a day.

From equations 8 and 9, a T of 110,000 $(gal/d)ft^2$ (14,700 ft^2/day) was estimated. With a saturated thickness of 44 ft, this translates to a hydraulic conductivity of 334 ft/d, which is a reasonable value and is close to the 350 ft/d (T = 15,440 ft^2/d) obtained through the Theis recovery method.

Walton (1962) presents a graphical method to estimate transmissivity from specific-capacity data. Walton's method provides graphs of specific capacity versus transmissivity for various pumping periods ranging from 2 minutes to 180 days. As a guide, two curves are given on each graph, one for a storage coefficient of 0.0001, a value reflecting confined conditions, and another for a storage coefficient of 0.20, a value reflecting water-table conditions. Since specific capacity varies with the logarithm of 1/S (where S = storage coefficient), relatively large errors in estimated storage coefficients result in comparatively small errors in transmissivity estimated from specific capacity data. The theoretical specific capacity calculated from equation 7 was used with a plot of transmissivity against specific capacity, presented in Walton (1962), to obtain a transmissivity value of 100,000 (gal/d)/ft (13,369 ft^2/d) or a hydraulic conductivity of 304 ft/d. This value agrees closely with the values calculated by Hurr's (1966) method.