Aquifer Model of the Susquehanna River Valley in Southwestern Broome County, New York

U.S. GEOLOGICAL SURVEY
Water-Resources Investigations
Report 85-4099

Prepared in cooperation with
SUSQUEHANNA RIVER BASIN COMMISSION

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By Allan D. Randall

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Albany, New York

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UNITED STATES DEPARTMENT OF THE INTERIOR

DONALD PAUL HODEL, Secretary

GEOLOGICAL SURVEY

Dallas L. Peck, Director

For additional information write to:

U.S. Geological Survey
P.O. Box 1669
Albany, New York 12201
(518) 472-3107

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(in pocket)

CONVERSION FACTORS AND ABBREVIATIONS

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

Multiply	by	To obtain metric unit
inch (in.)	2.540	centimeter (cm)
foot (ft)	0.3048	meter (m)
gallon (gal)	3.785	liter (L)
gallon per minute (gal/min)	0.003785	cubic meter per minute (m^3/min)
gallon per day (gal/d)	0.003785	cubic meter per day (m^3/d)
million gallons per day (Mgal/d)	0.04381	cubic meters per second (m^3/s)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.59	square kilometer (km^2)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)

AQUIFER MODEL OF THE SUSQUEHANNA RIVER VALLEY IN SOUTHWESTERN BROOME COUNTY, NEW YORK

By Allan D. Randall

Abstract

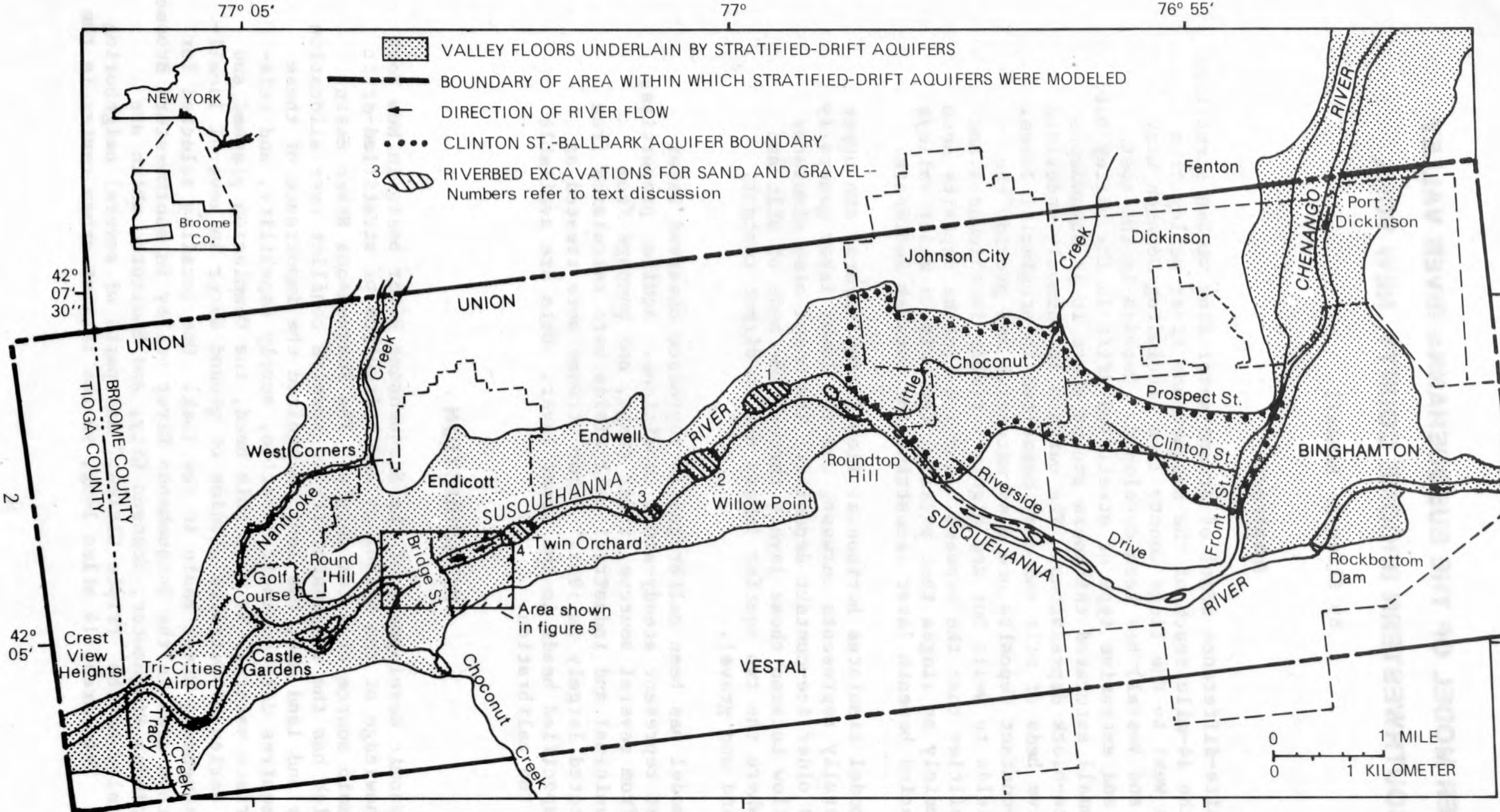
A finite-difference model of ground-water flow within stratified drift in the 14-mile reach of the Susquehanna River valley from Binghamton west to the Tioga County line (including Johnson City, Endicott, and Vestal) has been developed. Outwash is the most permeable and extensive type of stratified drift in the valley but has only small saturated thickness except where it is downwarped beneath ice-block depressions. The outwash is commonly underlain by extensive beds of silt and clay deposited in proglacial lakes. Older ice-contact deposits are also extensive and provide the largest yields to wells but are highly variable in thickness and commonly siltier than the outwash. The ice-contact deposits seem to occur mainly as ridges that parallel the axis of major valleys and are buried beneath later lacustrine and outwash sediments.

The model simulates horizontal flow in two layers; the upper layer generally represents outwash, and the lower layer generally represents older ice-contact deposits. The model also simulates vertical flow between those layers through the beds of silt and clay or, where the two aquifer layers are in direct contact, through sand and gravel.

The model has been calibrated to reproduce observed water levels that represent steady-state conditions. Aquifer properties, recharge from several sources, river stage, and pumpage from several municipal and industrial well fields were calculated from data collected largely in 1981. Major streams were treated as constant specified heads in the upper layer. Data are available to refine the calibration.

INTRODUCTION

Future economic development of the Susquehanna River basin in New York will require knowledge of the hydrology and hydraulics of stratified-drift aquifers, the main source of water supply. The Susquehanna River Basin Commission, which has the responsibility to resolve conflict over allocation of water supply and land use, has long recognized the importance of these aquifers and requires data on their location, supply capability, and relationship to surface waters. To meet this need, the Commission planned and administered a series of detailed studies of ground water in several localities of the Susquehanna River basin in New York. One locality selected for study was an urban area in the Susquehanna River valley in southwestern Broome County that includes Binghamton, Johnson City, and Endicott, which are referred to locally as the "Triple Cities," and parts of several neighboring towns (fig. 1). This area, 14 miles long, is the largest urban center in the



Base from U.S. Geological Survey
State base map, 1:500,000, 1974

Figure 1.--Location and geographic features of modeled area in Susquehanna River valley in southwestern Broome County, N.Y.

New York part of the Susquehanna River basin and was, in 1979, the only area in the basin from which large amounts of water were withdrawn from unconsolidated aquifers that had not been simulated by a digital-computer aquifer model. In 1979, the U.S. Geological Survey began a study, in cooperation with the Susquehanna River Basin Commission, that was financed on a cost-sharing basis by Federal funds through the Water Resources Council. The goal of this study was to prepare a digital computer model of aquifers in the Triple Cities area in the Susquehanna River valley to enable water-resources management and planning decisions by the Commission.

Purpose and Scope

This report describes the distribution and hydraulic properties of aquifers of the valley and the patterns of ground-water recharge and withdrawal, which were needed to design the aquifer model. It also summarizes the model design, calibration, and application. A set of maps depicts surficial geology, water-table configuration, model grid, and simulated vertical and horizontal water-transmitting capacity of aquifers.

Acknowledgments

Much of the data required for calibration were furnished by or collected with the cooperation and assistance of the Water Departments of Endicott, Johnson City, and Vestal, and several private companies, including International Business Machines Corp., Anitec Image Corp., and R. J. Martin, Consulting Engineer. Many residents of the area and other groups provided assistance and information.

STRATIFIED DRIFT IN SOUTHWESTERN BROOME COUNTY

The only earth materials in southwestern Broome County capable of supplying large yields to wells are deposits of sand and gravel within stratified drift, which is confined to the major valleys. Stratified drift along the Susquehanna and Chenango valleys in the Triple Cities is about a mile wide and generally 80 to 180 feet thick; lesser amounts occur along Nanticoke Creek (fig. 1).

Several maps of surficial geology and aquifer distribution in the Triple Cities have been published (Hollyday, 1969; Randall, 1977; MacNish and Randall, 1982; Holecek and others, 1982). The descriptions of geologic units and their distribution differ appreciably as a result of differences in purpose as well as scale and detail of the maps, and also as a result of growing knowledge of subsurface conditions, but all identify two contrasting types of valley-fill material: (1) In some areas a surface layer of sand and gravel of small saturated thickness overlies an extensive layer of clay, silt, and very fine sand that forms a barrier to downward ground-water flow. Deep gravel deposits may occur locally beneath the fine-grained layer, but little exchange of water occurs between the two layers. (2) Elsewhere, sand and gravel is generally continuous from land surface to bedrock with only small interspersed fine-grained units, so water can move vertically with relative ease. The latter areas have the best potential for large well yields but are also the most susceptible to contamination from the surface activities of man.

This contrast in aquifer geometry is evident in the surficial geologic map accompanying this report (pl. 1) and in the following pages, which describe the origin, composition, and distribution of aquifers and confining layers in the Triple Cities area and the factors controlling recharge to the aquifers.

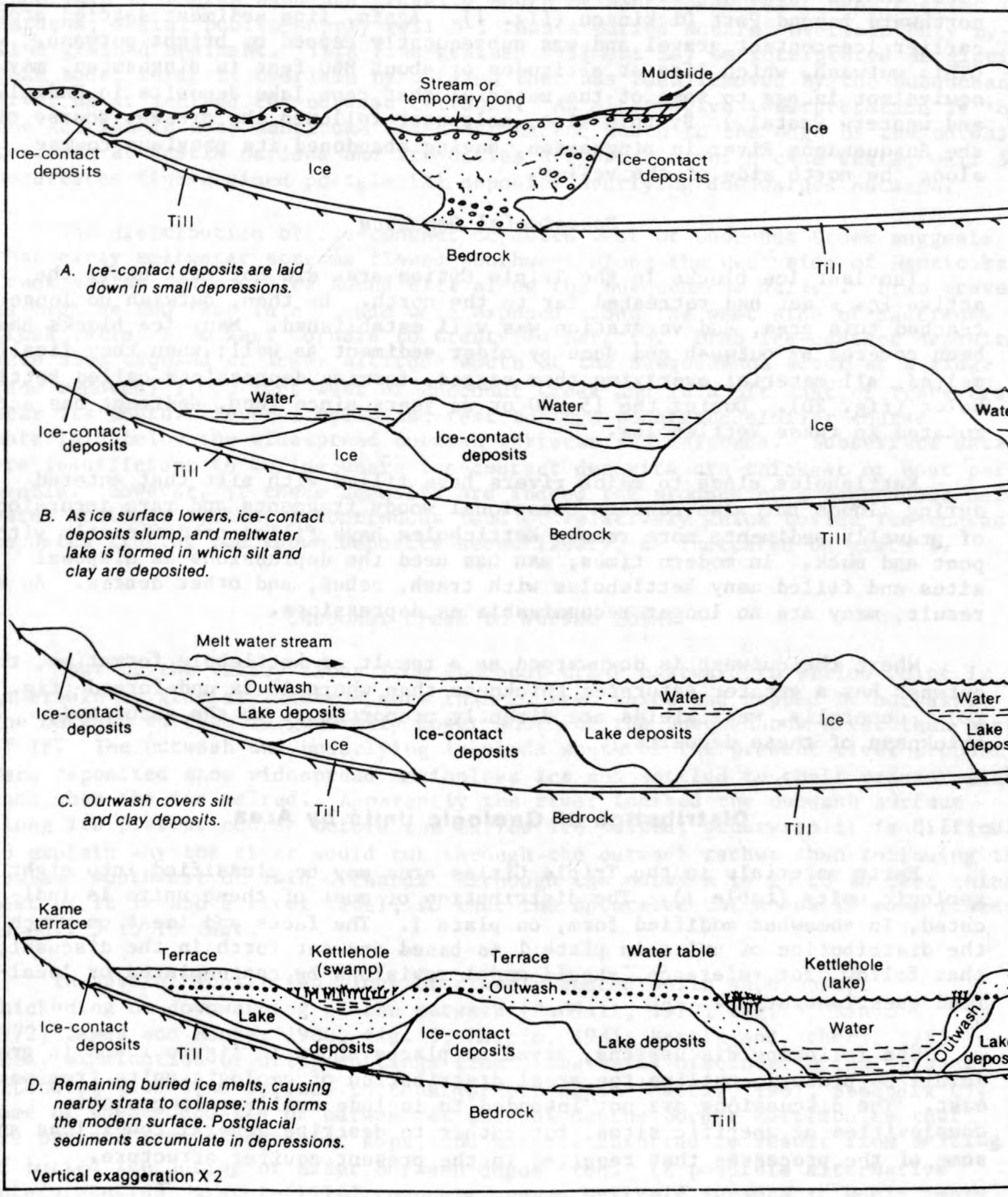
Origin of Glacial and Postglacial Deposits

About 18,000 years ago, a sheet of ice covered all of Broome County (Fleisher, 1977). This ice sheet flowed southward, scouring most of the land surface and locally depositing debris, termed till, beneath it. When the climate warmed and melted the ice faster than it was being replaced from the north, the uplands in each locality became uncovered first, while nearby valleys were still clogged with ice.

Stratified Glacial Drift

The earliest stratified deposits were formed as ice that was stranded in the valley bottoms began to melt. Sediment-laden meltwater flowed through fissures, narrow channels, or tunnels beneath and within the decaying ice and also between the ice and the valley walls (fig. 2A). Wherever the surface of the ice melted down to altitudes lower than older deposits downvalley, meltwater became ponded behind those deposits. The ponds quickly became filled with sediment--mostly coarse-grained material but interspersed with silt and with masses of muddy debris that slumped down from adjacent slopes. The early deposits consist almost totally of olive-gray shale and siltstone fragments derived from the local bedrock; they are termed "drab." Later, as headwaters of the meltwater drainage system extended further north (Randall, 1978a), the percentage of exotic pebbles from distant localities began to increase. All of the stratified deposits described so far are termed "ice-contact" because they were laid down against or atop ice. Their thickness is only a few feet or less in many places but locally reaches 50 to 100 feet. The thicker deposits seem to occur as belts parallel to the valley axes; for example, on the west side of the Chenango valley north of Prospect Street, along the Clinton Street-Ballpark aquifer (Randall 1977), generally near the center of the Susquehanna valley west of Johnson City, and along the west side of Nanticoke Creek valley (pl. 1).

At times, nearly all coarse sediment was trapped as ice-contact deposits in the upvalley (northerly) end of whatever valley reach then contained decaying ice. As a result, the lakes that formed where ice downvalley melted were not immediately filled with sediment and became quite extensive. For a while, a single lake probably extended from Johnson City westward into Tioga County. Fine-grained sediments ranging from silty clay to very fine sand gradually settled in this lake over the earlier ice-contact deposits of sand and gravel (fig. 2B). Later, meltwater deposited 10 to 40 feet of coarse sand and gravel, termed outwash, atop the fine-grained sediment in deltas and broad stream channels that covered nearly the entire width of the Susquehanna valley (fig. 2C). This coarse sediment, which originated largely from the Chenango valley, is termed "bright" because 20 to 40 percent of the pebbles are colorful rocks from areas north of Broome County (Randall, 1978). (This sediment differs sharply in appearance and composition from the older "drab" ice-contact deposits.)



A. Ice-contact deposits are laid down in small depressions.

B. As ice surface lowers, ice-contact deposits slump, and meltwater lake is formed in which silt and clay are deposited.

C. Outwash covers silt and clay deposits.

D. Remaining buried ice melts, causing nearby strata to collapse; this forms the modern surface. Postglacial sediments accumulate in depressions.

Vertical exaggeration X 2

Figure 2.--Sequence of stratified-drift deposition during deglaciation. This idealized diagram of a part of a valley cross section depicts the inferred geometry of stratified drift near kettleholes; it does not necessarily represent average conditions nor any particular locality.

A similar sequence of fine-grained lake sediment and outwash was deposited later in the Chenango valley when a large lake formed from eastern Binghamton northward beyond Port Dickinson (fig. 1). Again, fine sediment settled atop earlier ice-contact gravel and was subsequently capped by bright outwash. (This outwash, which lies at altitudes of about 860 feet in Binghamton, may be equivalent in age to some of the material that caps lake deposits in Endwell and western Vestal.) By this time, meltwater followed the present course of the Susquehanna River in Binghamton, having abandoned its previous course along the north side of the valley.

Postglacial Deposits

The last ice blocks in the Triple Cities area did not melt until the active ice sheet had retreated far to the north. By then, outwash no longer reached this area, and vegetation was well established. Many ice blocks had been covered by outwash and some by older sediment as well; when they finally melted, all material overlying them sagged, forming depressions called kettleholes (fig. 2D). During the 15,000 or so years since then, sediment has accumulated in these kettleholes.

Kettleholes close to major rivers have filled with silt that entered during floods but also contain occasional woody fragments and rare incursions of gravelly sediment; more remote kettleholes have filled predominantly with peat and muck. In modern times, man has used the depressions as disposal sites and filled many kettleholes with trash, ashes, and other debris. As a result, many are no longer recognizable as depressions.

Where the outwash is downwarped as a result of kettlehole formation, the outwash has a greater saturated thickness than where it is undeformed (fig. 2D). Generally, well yields are directly proportional to the saturated thickness of these deposits.

Distribution of Geologic Units by Area

Earth materials in the Triple Cities area may be classified into eight geologic units (table 1). The distribution of most of these units is indicated, in somewhat modified form, on plate 1. The facts and ideas on which the distribution of units in plate 1 is based are set forth in the discussions that follow, for reference, should model revisions be contemplated or localized questions arise.

The following discussions, keyed to places shown in figure 1 and in greater detail on plate 1, outline the areal distribution of geologic units from west to east. The discussions are not intended to include small-scale geologic complexities at specific sites, but rather to describe general conditions and some of the processes that resulted in the present aquifer structure.

Area West of Choconut Creek

Widespread terraces at or below 840 feet altitude consist of bright outwash over lake sediment. Numerous test borings for the Endicott sewage-treatment plant (Randall, 1972) indicate a large kettlehole near the mouth of Nanticoke

Creek that contains postglacial lake sediment, bordered on the north and underlain by coarse sand and gravel. South of Nanticoke Creek, near Castle Gardens, Vestal public-supply well 5-1 taps a buried aquifer overlain only by fine-grained sediment. This fine-grained sediment may be interpreted as glacial lake beds, formerly overlain by outwash that has been removed by the Susquehanna River as it incised the outwash terraces. An alternative interpretation is that the kettlehole near Nanticoke Creek may extend south to the edge of the outwash terraces at Castle Gardens and Tri-Cities Airport, in which case Vestal well 5-1 penetrates fine-grained postglacial deposits overlying downwarped outwash.

The distribution of ice-contact deposits west of Choconut Creek suggests that early meltwater streams flowed southwest along the west side of Nanticoke Creek valley and south of Round Hill along the Susquehanna valley. Drab gravels as high as 880 feet in altitude were exposed along the west side of Nanticoke Creek valley from West Corners to Crestview Heights. Drab ice-contact deposits were also exposed at lower altitude south of the Susquehanna River at a ridge (now removed) 1,000 feet west of Choconut Creek and in a pit east of Tracy Creek near its mouth. Several wells and test borings penetrate similar coarse materials below the widespread outwash terraces and lakebeds. Subsurface data are insufficient to define where ice-contact deposits are thickest or most permeable. However, if these deposits are indeed the product of a continuous meltwater drainage system, a continuous band of relatively thick buried ice-contact deposits linking the known deposits seems likely, as indicated on plate 1.

Choconut Creek to Willow Point

Most of the valley floor from Choconut Creek eastward to Willow Point is underlain by glacial lakebeds more than 20 feet thick and capped by outwash. The outwash surface is generally lower south of the Susquehanna River than north of it. The outwash and underlying lakebeds south of the present river probably were deposited atop widespread motionless ice and settled to their present altitude when the ice melted. Apparently the river incised the outwash surface along its present course before the buried ice melted; otherwise it is difficult to explain why the river would cut through the outwash rather than following the lowland southwest of Twin Orchard. Although the outwash is 25 to 40 feet thick, most of it is above river level, so that the saturated thickness in most places is only 5 to 10 feet.

In several localities, logs of closely spaced wells show abrupt thickening or downwarping of the outwash (Randall, 1970, fig. 5; Randall, 1972; Dames and Moore, 1980, fig. 5; Martin, 1981; Martin and others, 1983). Also, examination of drill cuttings from a few wells disclosed bright gravel extending from land surface to considerable depth (Martin, 1981, appendix C). Some of these deposits of outwash at unusual depths border kettleholes that are obvious from topographic maps, and all are inferred to result from melting of buried ice during or after outwash deposition. (A possible alternative interpretation, gravel-filled channels, seems unlikely because it would imply incision of sediment previously deposited over many miles downstream to depths greater than postglacial incision, followed by deposition of a long, continuous channel fill that has not been recognized.) As yet, detailed descriptions of pebble types have been made from only a few boreholes and exposures; further study might modify the preceding interpretation of the history and geometry of the valley fill.

Table 1.--Geologic units in the Susquehanna River valley, southwestern Broome County, New York.

Geologic unit (youngest to oldest)	Distribution, thickness, position	Materials composing unit	Hydrologic significance
Fill	In natural and excavated depressions, typically 5 to 20 feet thick; some sites can no longer be recognized as depressions.	Chiefly trash and ashes; some natural materials.	Increases concentration of dissolved solids and many individual constituents of infiltrating water, but effect decreases as age of fill increases. Not tapped by wells.
	Embankments for highways, dikes, railroads, etc.	Chiefly sand and gravel; some till, rock rubble, and other natural materials.	Minimal; reduced extent of flood inundation may slightly restrict aquifer recharge.
Alluvium	Flood plains of Susquehanna and Chenango Rivers. Some low-lying areas, never crossed by the river channel, contain only the upper (silt) unit.	About 15 feet of silt to fine sand that may include organic-rich layers, commonly overlying and interbedded with 10 to 15 feet of sandy pebble-cobble gravel that ordinarily contains only a few percent of silt; noncalcareous throughout.	Upper (silt) unit may limit infiltration from floods and possibly heavy rain. Lower (gravel) unit highly permeable, rarely tapped by wells, but may facilitate induced infiltration to deeper units.
	Flood plains and alluvial fans of tributary streams. Alluvial fans occur where tributaries enter the Susquehanna or Chenango valleys.	Gravel, 10 to 20 feet thick, moderately silty, noncalcareous. Most stones are flat pieces of local shale or siltstone.	Permeable, but too thin to sustain large-capacity wells. Seepage loss from tributaries on alluvial fans recharges underlying units.
Postglacial lake beds	Fills scattered irregular-shaped depressions, most less than 2,000 feet in diameter, left where ice blocks melted; as much as 80 feet thick.	Silt and very fine sand with some clay and scattered plant fragments; commonly grades into peat or highly organic silt at the top; thin lenses of sandy gravel may occur in areas bordering streams.	A significant barrier to infiltration and ground-water flow in places.

Table 1.--Geologic units in the Susquehanna River valley, southwestern Broome County, New York--continued.

Geologic unit (youngest to oldest)	Distribution, thickness, position	Materials composing unit	Hydrologic significance
Outwash	Present over most of the valley floors; forms broad terraces and is down-warped beneath postglacial lake beds; thickness ranges from 10 to 100 feet but is typically about 40 feet.	Sandy pebble gravel and pebbly coarse to fine sand, slight to moderate amounts of silt, highly calcareous. Contact with underlying lake beds commonly abrupt but locally gradational (finer grain size with depth) or interfingered. At least 20 percent of pebbles are limestone, quartzite, and other exotic rock types originating north of this region; at least 35 percent in upper part and in youngest terraces.	Highly permeable, tapped by several large-capacity wells, but over large areas the water table is only a few feet above the base of this unit.
Glacial lake beds	Probably forms two extensive deposits, one from Johnson City west into Tioga County, another from Binghamton north into Fenton, each interrupted in a few places by buried ridges or knolls of ice-contact deposits and by downwarping of younger units around several natural depressions. In western Binghamton, probably occurs only as lenses of limited extent.	Silt; silt with abundant thin layers of clay; silty very fine sand with layers of silt; calcareous; no plant fragments.	A significant barrier to vertical ground-water flow over large areas.
Ice-contact deposits	Exposed at land surface locally along valley sides, at altitudes up to 910 feet; commonly underlies younger units in central part of valley; thickness varies from zero to about 100 feet.	Sandy pebble to cobble gravel and pebbly sand with slight to more commonly abundant silt; typically characterized by layers of dense silt-bound sandy gravel; locally includes silt layers many feet thick.	Moderately to highly permeable; tapped by several large-capacity wells but too thin or too silty for large-capacity wells in many places.
Till	Immediately overlies bedrock and is the only material over the bedrock in most upland areas. Only a foot or two thick in many places, particularly on steep slopes, but several tens of feet thick beneath some hillsides and rounded hills low on the valley sides.	Unsorted mixture of silt, clay, stones, and sand; tough and compact; commonly called hardpan; contains rare, scattered lenses of sand or gravel. Deposits mapped as "morainal till" (pl. 1) are relatively thick and largely till but may be interbedded with appreciable sand, gravel, and (or) silt locally.	Very poorly permeable. Severely limits infiltration on hillsides; prevents induced infiltration from Susquehanna River for 3 miles west of Chenango River.
Bedrock	Present everywhere beneath other units.	Shale and siltstone.	Bedrock walls and floor of valley serve as aquifer boundaries. Poorly permeable but will yield 100 to 300 gallons per minute of salty water to wells several hundred feet deep.

Beneath the extensive lake deposits in this reach of the valley are early ice-contact deposits. Wells and test borings compiled from several sources indicate that these deposits are relatively thick near the Susquehanna River but thin rapidly to the north and south and are absent or only a few feet thick over large areas.

Throughout this reach of the valley, the juxtaposition of thick ice-contact deposits and locally collapsed outwash creates an extensive zone near the Susquehanna River in which coarse sand and gravel predominate. A relatively large saturated thickness of coarse deposits in this zone has been inferred previously by Randall (1977, fig. 1), MacNish and Randall (1982, pl. 1), Holecek and others (1982, sheet 3), and Hollyday (1969, pl. 1).

Willow Point to the Chenango River

The glacial geology of Johnson City and western Binghamton is described in an earlier report (Randall, 1977). The most significant feature of this reach is that the southern half of the Susquehanna valley floor is underlain by a thin layer of till over bedrock, whereas the northern half of the valley floor is underlain by a thick stratified-drift aquifer. The aquifer extends along Clinton Street from the Chenango River to the former baseball park in Johnson City, hence the name Clinton Street-Ballpark aquifer.

The Susquehanna River flows close to the south wall of the valley in this reach and for 3 miles is separated from the gravel aquifer by a broad, low bench of bedrock capped by several knolls (Randall, 1977, pls. 2 and 6). These knolls may consist of (1) till deposited directly by the ice, (2) flow till, deposited as mudslides off the ice, and (3) silty gravels deposited by meltwater streams. The knolls reach altitudes of 900 to 920 feet, approximately the same altitude as early ice-contact deposits that are exposed locally along the sides of the valley throughout the Triple Cities, and may be of the same age as those deposits. Some of the saddles between the knolls may be mantled with slightly younger gravels carried in from the north. However, till predominates in the southern half of the valley, both north and south of the river; hence these areas are mapped as morainal till in plate 1.

Over most of the Clinton Street-Ballpark aquifer, terraces consisting of bright outwash reach altitudes of 870 to 880 feet (Randall, 1977, pl. 6). Drab ice-contact deposits underlie the outwash and make up a large part of the valley fill (Randall, 1978a). In Binghamton, the bulk of the stratified drift is sand and gravel, with only small lenses of lake sediment, although postglacial lake sediment occupies two large kettleholes (pl. 1). In Johnson City, aquifer geometry is not well known and may be complex. Geologic logs of several test holes were available, but because pebble lithology has been determined for only a few holes, correlation remains uncertain.

At the east end of the Clinton Street-Ballpark aquifer, bright outwash lies near land surface along the Chenango River but is downwarped to the west near Front Street and extends beneath a kettlehole filled with postglacial lake sediments, as shown in cross section in figure 3. The collapsed outwash rests upon older ice-contact deposits, and together these units form a permeable basal aquifer. This aquifer configuration was demonstrated by correlation of gravel lithology and confirmed by water-level and temperature data (Randall, 1977, appendix F).

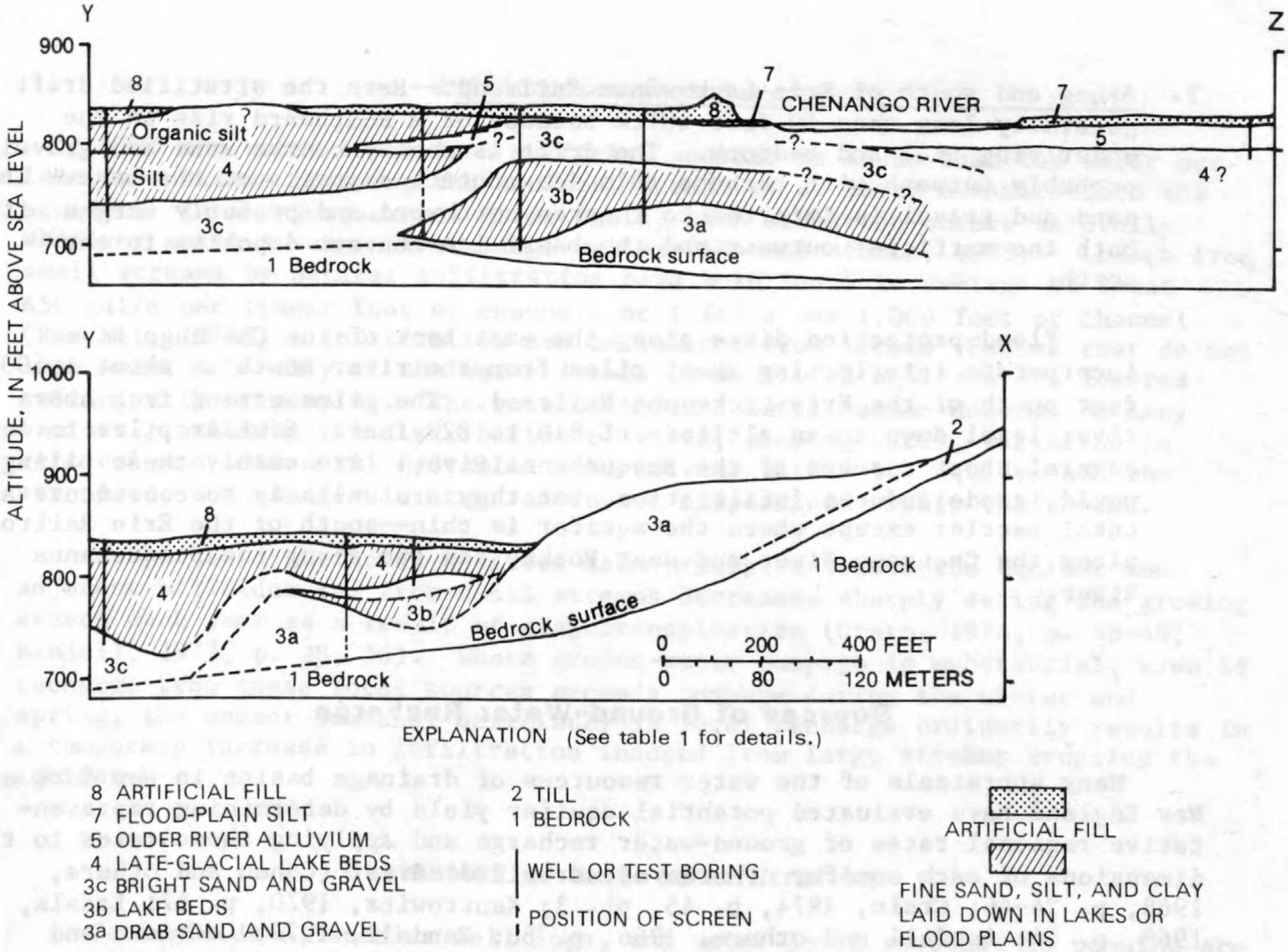


Figure 3.--Geologic sections in the Susquehanna River valley near Front Street in Binghamton. Above, view northward. Below, view westward. (Location shown on pl. 1. Modified from Randall, 1977.)

Eastern Binghamton

The valley floor within the town of Dickinson and the eastern part of the city of Binghamton can be classified into three areas:

1. <u>East of Chenango River and north of Erie-Lackawanna Railroad</u>.--Broad outwash terraces reaching altitudes of 850 to 870 feet are underlain by 20 to 45 feet of bright outwash atop extensive, apparently continuous lake deposits that are mostly silt. Beneath the lake deposits are unconsolidated deposits whose thickness and lithology are poorly known (Randall, 1972) but may consist of till and silty ice-contact sediment.

2. <u>West of Chenango River and north of Erie-Lackawanna Railroad</u>.--Several borings in this area within Binghamton (Randall, 1972) indicate thick variably silty sand and gravel from land surface to bedrock. The coarse deposits decline in altitude and may become thinner eastward but continue at least to the vicinity of Chenango Street beneath lake deposits. This band of sand and gravel is inferred to extend northward across the town of Dickinson (pl. 1) on the assumption that it was deposited by a river of considerable headwater extent that flowed in or under decaying ice.

3. Along and south of Erie-Lackawanna Railroad.--Here the stratified drift is generally less than 50 feet thick because of a southward rise in the underlying till and bedrock. The drift is mostly coarse sand and gravel, probably outwash with perhaps thin ice-contact deposits at the base. The sand and gravel is inferred to thicken northward and probably merges with both the surficial outwash and the basal ice-contact deposits to the north.

Flood-protection dikes along the east bank of the Chenango River incorporate interlocking sheet piles from the river mouth to about 1,500 feet north of the Erie-Lackwanna Railroad. The piles extend from above river level down to an altitude of 810 to 820 feet. Similar piles border several short reaches of the Susquehanna River. Presumably these pilings would impede induced infiltration, but they are unlikely to constitute a total barrier except where the aquifer is thin--south of the Erie Railroad along the Chenango River and near Rockbottom Dam along the Susquehanna River.

Sources of Ground-Water Recharge

Many appraisals of the water resources of drainage basins in New York and New England have evaluated potential aquifer yield by determining representative regional rates of ground-water recharge and applying those rates to the dimensions of each aquifer or area of stratified drift (Cohen and others, 1968, p. 24-46; Crain, 1974, p. 45, pl. 3; Kantrowitz, 1970, p. 67; LaSala, 1968, p. 54; Randall and others, 1966, p. 66; Randall, 1977; Cervione and others, 1972, p. 46-47). Several components of recharge were considered in these studies and are explained in the following paragraphs.

Precipitation on the Aquifer

Most of the rain and snowmelt in areas of sand and gravel infiltrates the soil. The Susquehanna River valley is such an area; of the 36 inches of average annual rainfall at Binghamton, probably 12 to 15 in/yr returns to the atmosphere by evapotranspiration; the rest reaches the water table as recharge (Randall, 1977). The annual volume of recharge from precipitation to a surficial aquifer depends principally on the extent of the surficial sand and gravel deposits and on the annual precipitation rate.

Precipitation on Hills Bordering the Aquifer

The Susquehanna River valley is bordered by hills blanketed by till containing a large percentage of silt and clay, which permits only a small percentage of rain or snowmelt to infiltrate beyond the top foot or two; the excess flows downslope on or just below land surface until it reaches the valley floor (Ku and others, 1975; MacNish and Randall, 1982), whereupon it seeps into the stratified drift and becomes recharge. Also, a small continuous flow of ground water moves through the bedrock from the upland toward the Susquehanna valley, where it, too, seeps into the stratified drift.

Infiltration from Streams

When the water level in a surficial aquifer is lower than the water surface in a stream crossing the aquifer, stream water moves downward into the aquifer. In the Susquehanna River basin, this situation occurs naturally wherever small streams enter large valleys (Ku and others, 1975). Losses from small streams by natural infiltration have been found to average at least 650 gal/d per linear foot of channel, or 1 ft^3/s per 1,000 feet of channel (Randall, 1978b). Infiltration can be induced from stream reaches that do not lose water naturally if the water levels in surficial aquifers are lowered sufficiently by pumping. The rate of induced infiltration depends on many factors, including the distribution of wells, pumping rates, variation in vertical and horizontal hydraulic conductivity within the aquifer and the streambed, and changes in stage and water temperature within the stream.

Recharge to surficial aquifers from precipitation on the aquifer and adjacent hillsides and from small streams decreases sharply during the growing season each year as a result of evapotranspiration (Crain, 1974, p. 40-46; Randall, 1977, p. 28, 56). Where ground-water pumpage is substantial, even if recharge from these local sources exceeds pumpage during the winter and spring, the summer seasonal deficiency in local recharge ordinarily results in a temporary increase in infiltration induced from large streams crossing the aquifer.

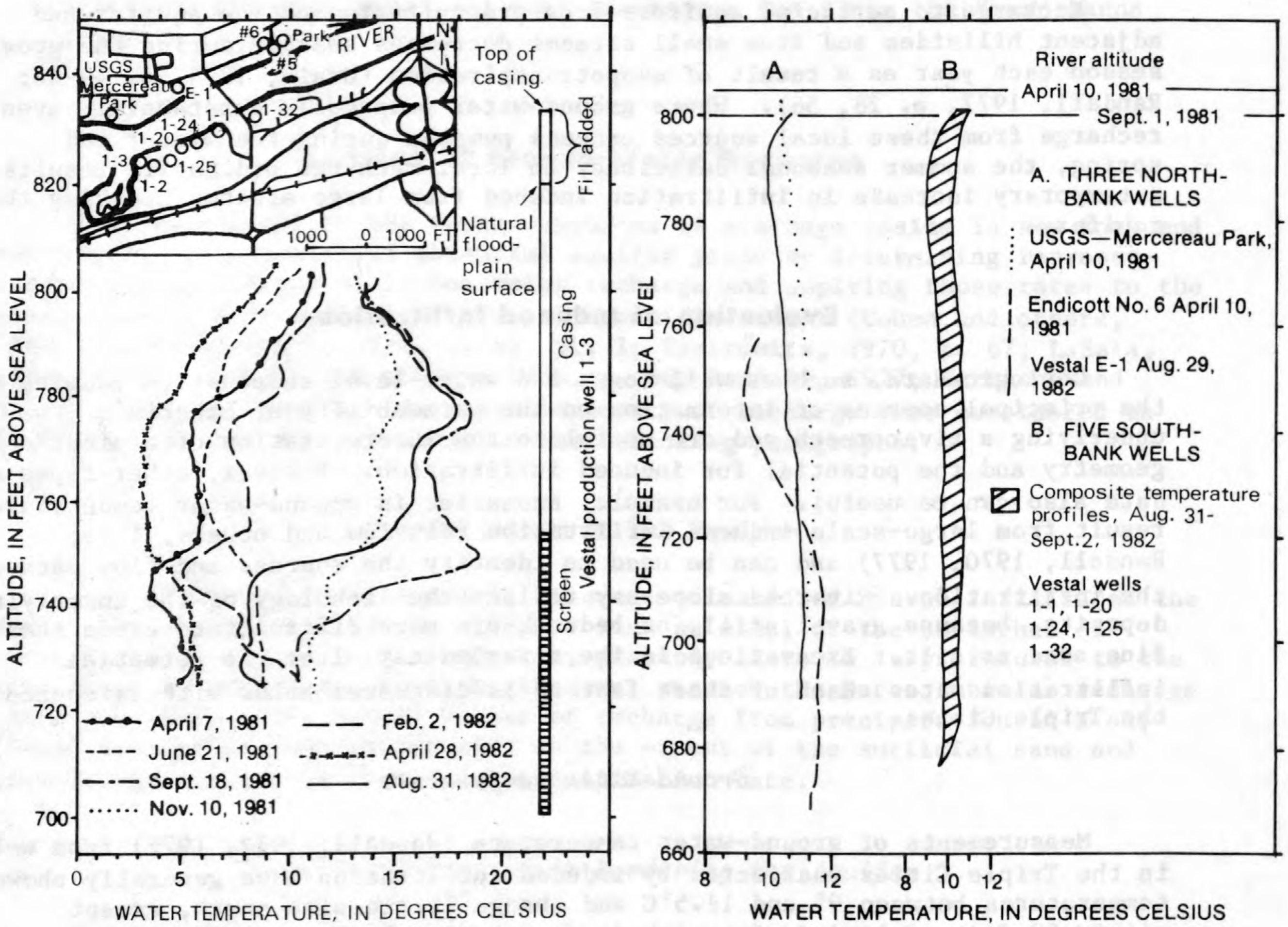
Evaluation of Induced Infiltration

Geologic data, such as well logs, and water-level response to pumping are the principal sources of information on the permeability of materials underlying a river reach and are useful to the interpretation of aquifer geometry and the potential for induced infiltration. However, other types of data also can be useful. For example, anomalies in ground-water temperature result from large-scale induced infiltration (Winslow and others, 1965; Randall, 1970, 1977) and can be used to identify the sources and flow paths of the infiltration. Riverbed slope may reflect the lithology of the underlying deposits, because gravel, till, or bedrock are more difficult to erode than fine sand or silt. Excavations in the riverbed may alter the potential infiltration rate. Each of these factors is discussed below with reference to the Triple Cities.

Ground-Water Temperature

Measurements of ground-water temperature (Randall, 1972, 1977) from wells in the Triple Cities unaffected by induced infiltration have generally shown temperatures between 9° and 11.5°C and change little with depth, except within 20 feet of land surface, where some seasonal fluctuations occur. Induced infiltration, however, can cause the temperature at depths greater than 20 feet to reflect the seasonal temperature changes in the infiltrated river water. The largest annual fluctuations are in the most permeable layers, and the temperature of water in permeable layers close to the riverbank may fluctuate widely--from near 0°C in March to 22°C in September. Further from the river, temperatures fluctuate less, and the seasonal extremes in temperature are delayed.

Vertical temperature profiles of several wells along the banks of the Susquehanna River between Vestal and Endicott illustrate the use of water-temperature data in locating areas of connection between riverbed and aquifer. The wide seasonal fluctuation in temperature at Vestal well 1-3 (fig. 4A) is unequivocal evidence of large-scale induced infiltration through the nearby riverbed. In contrast, temperatures in eight nearby wells were nearly constant with depth (fig. 4B), which indicates little or no induced river water moving past these wells even though Vestal wells 1-1, 1-2 or 1-3, and Endicott well 5 and Park well had all been pumped regularly at rates exceeding 1 Mgal/d before the measurements. Presumably, therefore, impermeable materials beneath parts of the riverbed prevent appreciable induced infiltration near these wells, and(or) ground-water flow past some wells moves toward or parallel to the river in response to pumping on the far bank.



A. *Test well 5 feet south of Vestal public-supply well 1-3, April 1981 to August 1982.*

B. *Three north-bank wells and five south bank wells north and east of Vestal well 1-3.*

Figure 4.--Water-temperature profiles of wells along the Susquehanna River in Vestal and Endicott, N.Y. (Location of inset map is shown in fig. 1.)

The temperature profiles in figure 4A also indicate hydraulic connection between Vestal wells 1-2 and 1-3. Several of these profiles were measured when well 1-3 had been in regular use for municipal water supply and well 1-2 had been mostly idle. However, well 1-3 was not pumped from mid-1980 through May 1981 nor to any great extent from May to August 1982. During these periods, well 1-2 was used to meet the demand for water. Temperature anomalies measured at well 1-3 when well 1-2 was in use (profiles for April 7, 1981 and August 31, 1982 in fig. 4A) are consistent in magnitude and depth with those measured when well 1-3 was in use, which demonstrates that river-water infiltration induced by well 1-2 originates northwest of well 1-3 and is drawn past well 1-3 toward well 1-2 through some continuous body of sand and gravel.

Other examples of the use of ground-water temperatures to locate induced infiltration in the Triple Cities are given by Randall (1970; 1977, p. 62-65; 1977, p. 67). Water-temperature data helped establish the continuity of aquifers near Front Street in Binghamton (shown in cross section in fig. 3 and block diagram in fig. 5) and further demonstrated that 78 percent of the water reaching GAF well 7 in 1977 (fig. 5) was river water that had seeped into the upper aquifer. The induced recharge then moved horizontally, followed the bright outwash where it is downwarped to the base of the valley fill slightly northeast of GAF well 7, and was withdrawn as well water (Randall, 1977, p. 62-65).

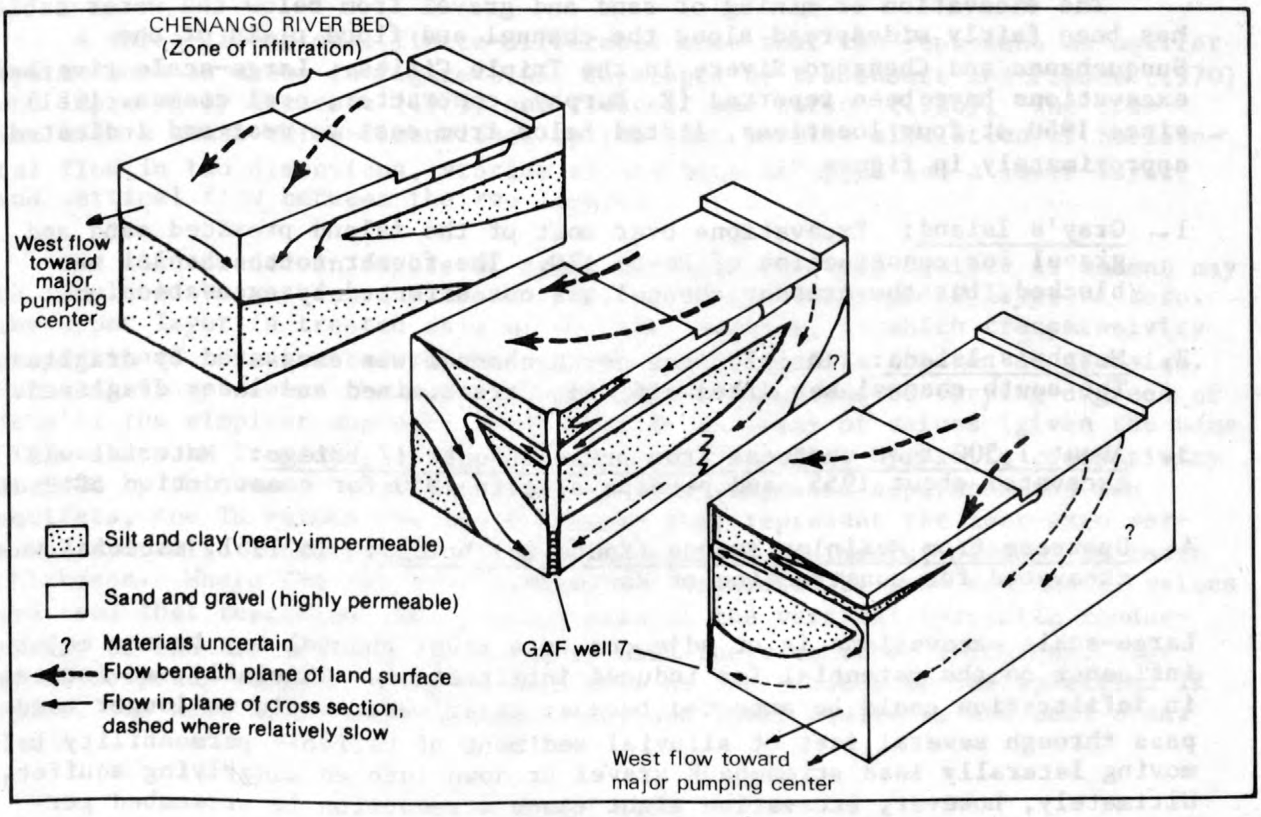


Figure 5.--Aquifer geometry and movement of induced recharge from Chenango River near Front Street in Binghamton. The three blocks represent adjacent sections of stratified drift, sliced and separated to reveal the geologic configuration. Geologic cross sections of area are shown in figure 3. (Modified from Randall, 1977.)

Riverbed Slope

The locations of riffles and pools along the Susquehanna River from Johnson City to Nanticoke Creek were sketched or plotted on topographic maps in 1967 and 1981. The river-surface altitude at several points downstream from Binghamton was determined by leveling. Riffles were relatively abundant and river-surface profiles relatively steep in three reaches: (1) from just above the confluence of Chenango and Susquehanna Rivers to Roundtop Hill in Vestal, a 3-mile reach; (2) north of Willow Point in Vestal, a 0.9-mile reach; and (3) from Bridge Street in Vestal westward past Round Hill in Endicott, a 1.4-mile reach. In the first reach, channels are incised in till and bedrock; in the latter two reaches, channels cross areas inferred from well and test boring records to be underlain by gravel and coarse sand (pl. 1). The relatively steep river-surface profiles in these three reaches may reflect slower erosion of gravel, till, or bedrock than of the sandy outwash or lake sediment that underlie the river alluvium elsewhere. If so, river slope could provide one means of identifying reaches underlain by coarse or resistant materials. Till and bedrock are rare in major valleys; therefore, most steep reaches may be underlain by gravel of relatively high infiltration potential.

Riverbed Excavations

The excavation or mining of sand and gravel from below the water table has been fairly widespread along the channel and flood plain of the Susquehanna and Chenango Rivers in the Triple Cities. Large-scale riverbed excavations have been reported (R. Murphy, contractor, oral commun. 1981) since 1950 at four locations, listed below from east to west and indicated approximately in figure 1.

1. Gray's Island: Excavations over most of the island provided sand and gravel for construction of Route 17C. The former north channel was blocked, but the present channel was not affected by excavation.

2. Murphy's Island: In 1970, the north channel was excavated by dragline. The south channel was diked off and surface mined and later draglined.

3. About 1,500 feet upstream from present Route 17 bridge: Material was excavated about 1955, and perhaps also in 1970 for construction of Route 17.

4. Upstream from McKinley Avenue (Route 26) bridge: In 1961, material was excavated for construction of Route 26.

Large-scale excavations in or adjacent to a river channel may have a major influence on the potential for induced infiltration. Initially, an increase in infiltration could be expected because river water would no longer need to pass through several feet of alluvial sediment of variable permeability before moving laterally into streambank gravel or down into an underlying aquifer. Ultimately, however, excavation might cause a reduction in streambed permeability because the quiet water in the unnaturally deep excavated pools might trap a substantial thickness of silt, thus sealing a section of the aquifer from the river. (Alternatively, if enough sand and gravel were being transported along the riverbed to refill the excavations with permeable materials, the streambed permeability would not be decreased.)

The U.S. Geological Survey has not documented the location or extent of riverbed gravel mining nor investigated the rate of refilling or its effects on riverbed permeability. The subject seems sufficiently important, however, to warrant mention, even if no quantitative analysis can be provided.

GROUND-WATER FLOW MODEL

Model Design

A diagram illustrating the stratigraphic relationships described in preceding sections is shown in figures 6A and 6B. The gravel layers may be thought of as forming two aquifers--a surficial aquifer that extends throughout the valley, although varying somewhat in thickness, and a deeper aquifer that is discontinuous and commonly separated from the surficial aquifer by a layer of impervious lake sediment. A continuous gravel deposit of essentially uniform age may be part of the surficial aquifer in one locality and part of the buried aquifer in another. Figure 6C shows how a two-layer model can represent this aquifer system.

Model Code

A three-dimensional finite-difference code that can represent an aquifer system such as shown in figure 6 was developed by Bredehoeft and Pinder (1970) and improved by Trescott (1975) and Trescott and Larson (1976). The code includes a quasi-three-dimensional option that permits simulation of horizontal flow in two dimensions, storage within both an upper and a lower layer, and vertical flow between the two layers.

As illustrated in figure 6, areas in which the deep aquifer is absent may be simulated by setting the transmissivity of the lower model layer at zero. The upper layer is treated as a water-table aquifer, in which transmissivity proportional to saturated thickness; the lower layer is treated as artesian. Vertical flow between the two aquifers may be simulated to varying degrees of detail; the simplest approach is to specify an array of values (given the name "TK" values by Trescott, 1975) that represent vertical hydraulic conductivity divided by thickness. Where a layer of lake sediments separates the two aquifers, the TK values are small because they represent the near-zero vertical hydraulic conductivity of the fine-grained sediments, divided by their thickness. Where the two aquifers are not separated by lake sediments, values are used that represent the harmonic mean of the vertical hydraulic conductivity of the two aquifers, divided by half their combined thickness. The model computes vertical flux in each cell as the product of the specified TK value, the head difference between upper and lower aquifers, and cell area.

Model Grid

To adapt this code to the Triple Cities, it was necessary to design a model grid and to specify boundary conditions, recharge distribution, and aquifer properties. The model grid, illustrated on plate 3, contains 63 rows and 159 columns of cells representing an area of 66 mi^2. Most of this area is

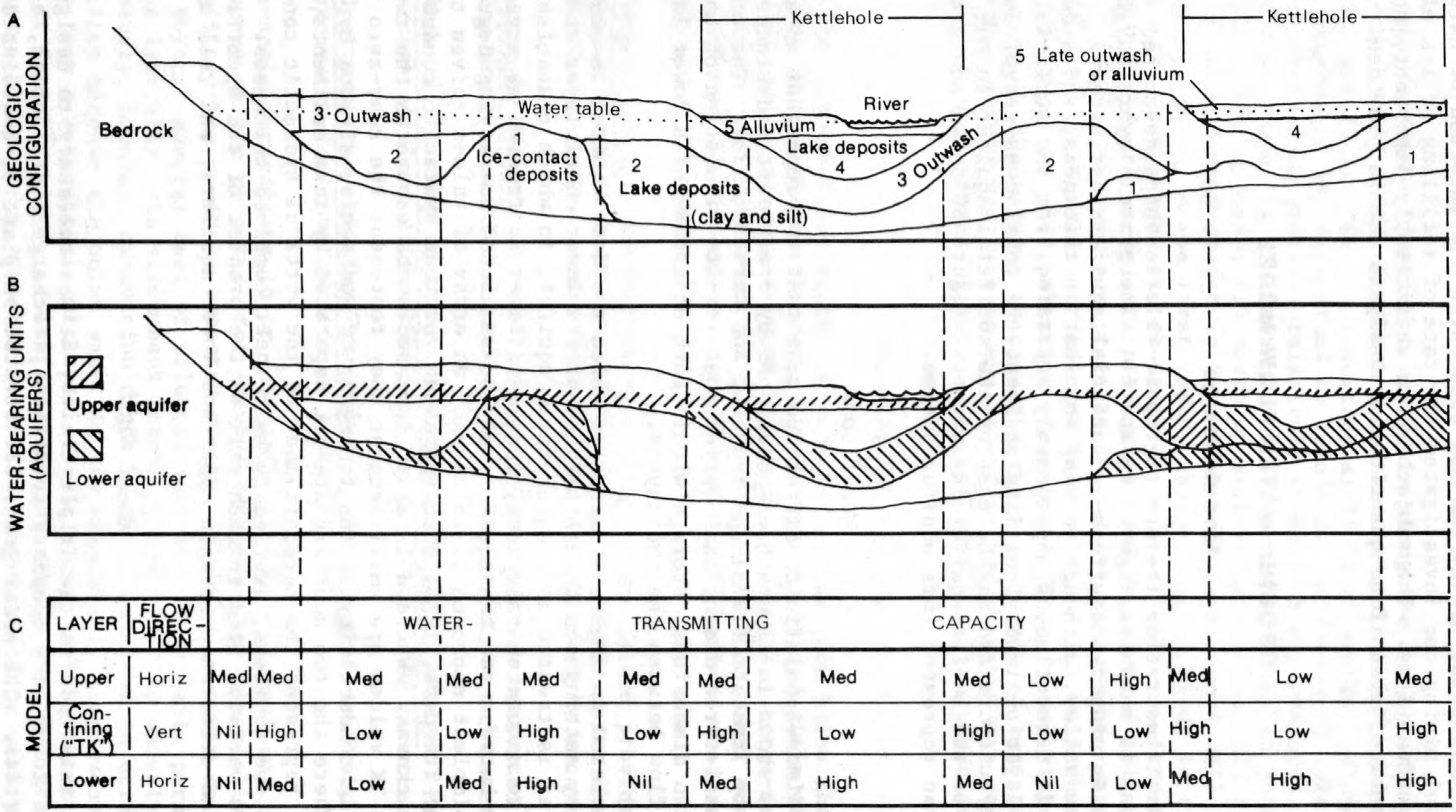


Figure 6.--Idealized aquifer in southwestern Broome County, N.Y., and corresponding aquifer model geometry: A. Geologic units. B. Classification of geologic units into aquifers and confining layers. C. Representation of aquifers and confining layers by a digital model. Dashed lines indicate alignment of equivalent segments of A, B, and C.

outside the major valleys that contain stratified-drift aquifers and was therefore excluded from model computations. The active part of the model grid, outlined on plate 3 by a heavy line, represents the entire valley floor within the City of Binghamton and the towns of Dickinson, Vestal, and Union (including, within Union, the villages of Endicott and Johnson City). The grid is oriented parallel to the general trend of the Susquehanna valley to minimize the number of cells needed.

Over most of the model area, cell lengths and widths range from 150 to 700 feet, but dimensions of 800 to 1,800 feet are used near the boundaries. The variable cell spacing permits location of most production wells and observation wells at the center of a cell. Also, the relatively fine grid facilitates detailed analysis, which is advantageous for two reasons:

1. The Triple Cities model is intended as a general-purpose model to demonstrate how the aquifer system functions and to answer questions that may arise in the future rather than to solve specific current problems. The more cells within a given area, the greater the flexibility of the model in representing any postulated stress.

2. The process of model calibration could include simulation of the shape of the cone of depression around several large-capacity wells shortly after startup or shutdown and after long periods of steady pumping. Small cells were necessary to accomodate the close spacing of observation wells and abrupt geologic variations in some of these localities.

Model Boundaries

Boundaries are established to define the part of the grid within which the model will operate. They are usually placed either (1) to coincide with the location of known hydrologic features, or (2) far enough from the area of interest that any error in representing flow across the boundary will not distort model simulations in the area of interest. The type of boundary is selected to best represent the rate at which water is expected to enter or leave the active part of the model grid.

<u>Valley Walls</u>.--Till and bedrock along the sides of the major valleys are treated as impermeable boundaries because very little ground water moves through them. Cells representing this area are outside the model and are assigned zero hydraulic conductivity, which results in zero flow intc or out of the model area across the valley walls. In some areas, the upper model layer extends slightly farther up the valley wall than the lower layer to conform to actual stratigraphy. Unsaturated ice-contact deposits high on the sides of the valley were ignored because they are not part of the aquifer.

<u>Boundaries within the valleys</u>.-Arbitrary impermeable boundaries were drawn across the major valleys near the edge of the model grid. These boundaries do not represent any known abundance of impermeable material in the stratified-drift aquifers at these locations. Rather, their placement was intended to be reasonably far from the area of interest. Most of these arbitrary no-flow boundaries are short and approximately parallel to inferred directions of natural ground-water flow, and simulated high vertical water-transmitting capacity causes heads in both model layers to closely approach river level,

which was specified as explained in the next section. The upstream boundaries were placed at the narrowest part of the Susquehanna River valley near the State Hospital, which nearly coincides with the Binghamton city line, and at the Dickinson-Fenton town line in the Chenango valley. The downstream boundary was placed far enough into Tioga county to include municipal wells serving the Crestview Heights development. If some model simulations were to result in water-level changes at one of these arbitrary boundaries, it would be necessary to specify a constant head or head-dependent flux along the boundary to more closely reproduce real conditions.

Rivers

Model cells that represent the upper aquifer beneath the Susquehanna or Chenango Rivers or bordering Nanticoke Creek were assigned constant heads that corresponded to water-surface altitude for the time period simulated. Altitude of the water surface at 11 sites on these three streams was determined on one or more dates by spirit leveling or by measurement from a reference point whose altitude was known. The water-surface altitude at each site was correlated with the concurrent altitude at the U.S. Geological Survey gaging station on the Susquehanna River at Bridge Street in Vestal (pl. 1). From the correlations, river-surface altitude at each site could be estimated for any date or period of interest. For cells between these sites, river-surface altitude was interpolated in a nonlinear manner that took into account the distribution of pools and riffles noted during river traverses in 1967 and 1980.

Seepage between the river and the lower model layer is computed by the model as a function of (1) vertical hydraulic conductivity of the confining layer or, where no confining layer is simulated, a vertical hydraulic conductivity representative of the aquifer, and (2) any difference between the variable head in the lower-layer cells and the fixed head in the upper-layer cells. Seepage between the river and the upper layer is computed from head differences and horizontal hydraulic conductivity within that layer near the river's edge. In the real upper aquifer, however, water must move vertically as well as horizontally to enter the river. Accordingly, calibration of the model as now designed may cause the hydraulic conductivity of the upper model layer to be less than the horizontal hydraulic conductivity of the shallow aquifer it represents because the model value must incorporate the effect of resistance to vertical flow.

Recharge from Precipitation on the Aquifer

Recharge from direct precipitation on the aquifer was calculated as precipitation minus evapotranspiration for the period of interest, as follows:

Precipitation.--Records of daily precipitation for 1981 and 1982 are available (National Weather Service, 1981-82) for two stations near the Triple Cities but outside the Susquehanna River valley--at the Broome County Airport, 6 mi north of the river, and at a site in Vestal, 3 mi south of the river. In addition, precipitation during most of 1981-82 was recorded by weighing rain gages at one to three locations in Union in connection with a sewer project (V. Shumaker, engineer, written commun., 1982) and at the government center in downtown Binghamton (D. Gardner, Binghamton Engineering Department, written commun., 1982). Data from the six stations were averaged to obtain monthly precipitation over the model area. Generally most weight was given to measurements made within the Susquehanna valley.

Evapotranspiration.--Computation of evapotranspiration involved three steps:

1. An estimate of average annual evapotranspiration was selected from previous studies. Ku and others (1975, p. 5) found that if average annual evapotranspiration in southwestern Broome County were about 19.5 inches, they could draw mutually consistent regional maps of precipitation and runoff for 1931-60. However, this regional analysis was based on runoff from large basins, all of which consist predominantly of till-mantled uplands in which most runoff remains within a foot or two of land surface, within the reach of plants. Many of these basins also contain large wetlands on the valley floor. In contrast, the water table in most areas of stratified drift within the Triple Cities remains many feet below land surface throughout the year. Calculations that use 19.5 inches of evapotranspiration yield recharge rates of zero during some months in which water-level rises were observed. This suggests that evapotranspiration from stratified drift in the Triple Cities may be generally less than the regional average. Water budgets for small areas of stratified drift in Binghamton and in southeastern Cortland County (Randall, 1977, p. 60; written commun., 1981) yielded annual evapotranspiration estimates of 12 to 15 inches for areas that seem generally typical of the Triple Cities. Therefore, an annual evapotranspiration rate of 14.5 inches was selected to develop initial estimates of recharge from precipitation.

2. The percentage of the annual evapotranspiration that occurred each month was calculated by a method developed by Olmsted and Hely (1962, p. A-13). This method treats evapotranspiration as proportional to the product of the mean temperature (°C) and the percentage of annual daylight during each month. For this study, because monthly mean temperatures for November through March were near or below 0°C, a value of zero was substituted for each daily mean temperature below zero, and the monthly means were recomputed to obtain a set of values that approached zero in very cold weather but remained positive.

3. The percentage of annual evapotranspiration for each month (step 2) was multiplied by the estimated average annual evapotranspiration (step 1) to yield monthly evapotranspiration, in inches of water.

Recharge in inches was calculated as the difference between precipitation and evapotranspiration for each month in 1981 and 1982. Evapotranspiration exceeded precipitation in some summer months; the resulting negative recharge values were considered to represent soil-moisture deficits and were carried over from month to month until cancelled by positive recharge values in succeeding months. Some positive values in winter were considered to represent surface snow and ice and were carried over until the next major warm spell. Occasionally, the entire computation was run for half-months rather than months to reflect erratic precipitation distribution. Monthly or half-monthly recharge values could then be averaged over any period of interest and the result converted to feet per second for use in the model.

Recharge from Hills Bordering the Aquifer

Recharge from till-covered hillsides that slope toward stratified-drift aquifers comes from the same climatic events as direct recharge to the aquifer and should occur at nearly the same time; therefore, both were incorporated into the same array of recharge data for the upper layer of the model. This data array has a value of 1 in most cells, which the computer multiplies by the recharge rate specified for precipitation on the aquifer. To incorporate recharge from hillsides, drainage divides were sketched on maps up the hillside from the corners of each model cell that lay on the perimeter of the active cells. In undeveloped localities, the area between these small divides was compared by inspection with the area represented by the cell. If the hillside area was twice that of the cell, a value of 2 for the hillside plus 1 for the cell itself, or a total of 3, was entered for that cell in the data array. If the hillside area was 5 times that of the cell, a value of 5 + 1 = 6 was entered. Thus, a constant fraction of precipitation on both the cell and the adjacent hillside is treated as recharge to that cell. It was assumed that much of the storm runoff from hillsides that are crossed by one or more roads would be intercepted by roadside ditches and storm sewers. Therefore, the hillside area in such localities was divided by an arbitrary factor of 3; for example, if the hillside area were 6 times the cell area, a value of 3 was entered for that cell (6/3 = 2, plus 1 for the cell itself = 3). This is an intuitive adjustment without theoretical or empirical basis; future field observations or model analysis may provide a basis for revising estimates of ditch and storm-sewer interception of recharge. Where the Susquehanna River or Nanticoke Creek abut till-covered hillsides with no intervening aquifer, runoff from the hillsides was ignored as a source of recharge.

Recharge from Tributary Streams

Loss of water by seepage from upland tributary streams where they cross a surficial aquifer was simulated by specified constant fluxes in the layer of the model in cells along the courses of the streams. Stream losses were computed in several ways:

1. For the few stream reaches that cross the aquifer in culverts or concrete-lined channels, loss was considered to be zero.

2. For Little Choconut Creek and tributaries in Johnson City, loss was estimated from Randall (1977, fig. D3) and distributed along the channel in proportion to loss rates measured on various dates (U.S. Geological Survey, 1970). This method requires an estimate of average flow duration for the period of interest, which was obtained by calculating the average flow recorded at a nearby gaging station during that period and obtaining its duration from a flow-duration table (Ku and others, 1975).

3. For most small streams, a general method presented by MacNish and Randall (1982) was used:

   (a) The length of tributary channel lying within the major valley and having a slope greater than 1 percent was measured on a topographic map and multiplied by a loss rate of 1 ft^3/s per 1,000 feet of channel. The product is a conservative estimate of potential loss.

(b) Flow of the stream during the period of interest was computed from graphs in MacNish and Randall (1982, figs. 14-15) after estimating flow duration in the same manner as in step 2.

(c) Seepage loss was taken as equal to either potential loss or streamflow, whichever was smaller, and distributed in cells along the channel.

For a few tributaries, the general method was modified to fit known stream characteristics or stratigraphy.

Aquifer Properties

Horizontal water-transmitting capacity. Three types of data were used to estimate transmissivity of the lower model layer and (or) hydraulic conductivity of the upper model layer at individual sites. They are: (1) geologic logs of about 260 wells and test borings, (2) specific capacities of about 80 large-capacity municipal or industrial wells or test wells, and (3) results of pumping tests of about 20 large-capacity wells, generally conducted for about 24 hours at the time of well completion and including measurements in some observation wells. These data were analyzed by methods described by Randall (1977, p. 39). For the lower model layer, individual values were plotted, and contours of equal transmissivity were drawn to reflect the plotted values and the inferred thickness of the aquifer. For the upper model layer, individual values of hydraulic conductivity showed no obvious areal pattern, so an initial average value of 500 feet per day was used everywhere except for a few localities with abundant data or where geologic interpretations indicated fine-grained surficial materials.

Vertical water-transmitting capacity. The conceptual design of the Triple Cities model was identical to that of a model developed earlier for an area in southwestern New York. The values for uninterrupted sandy gravel and for silty confining layers in that model, all originally based on typical vertical hydraulic conductivity or anisotropy values from the literature, were used in the Triple Cities model.

Model Calibration

The process of comparing measured hydrologic phenomena with model simulations of the same phenomena, and adjusting estimated model parameters to minimize any differences between the two, is known as model calibration. It is an essential step in verifying or modifying the concepts of aquifer geometry and hydraulics used in designing the model and in appraising the reliability or limitations of the model as a means of predicting aquifer response to postulated future stresses. Data collected by the U.S. Geological Survey from the Triple Cities have included water levels, pumpage, river stage, precipitation, and stream-seepage measurements that collectively make possible several types of calibration for a digital computer model. On the following pages, a data set especially suitable for steady-state calibration is evaluated and results of model calibration presented. Other data sets that could be used for three different types of transient-state calibration are also described.

Steady-State Calibration for April 1981

The process of model calibration is customarily started with a steady-state simulation incorporating water levels, recharge, and discharge rates for a period during which these factors were reasonably constant or at least exhibited no net change. A steady-state simulation permits adjustments to be made in the arrays of data for horizontal hydraulic conductivity and transmissivity, vertical water-transmitting capacity, and bottom altitude of the upper layer (all required by the Triple Cities model) without the complications of simulating storage, changing pumping rates, and fluctuating river stage at the same time.

The hydrologic setting in April 1981 was representative of steady-state conditions. Ordinarily, early April is a time of high and rapidly changing water levels in the stratified-drift aquifers of the Susquehanna River basin because periodic snowmelt and rain on saturated upland soils produce peaks in river stage, and infiltrating rain and river water cause sharp rises in ground-water levels. In 1981, however, most of the snow melted in late February, and precipitation from March through May was subnormal. As a result, both ground-water levels and river stage were subnormal in April and showed only modest fluctuations with little net change throughout April and May, as shown in the hydrographs in figure 7. Recharge from precipitation during April and May was estimated by the method described in the section on model design to average 1.55 in/mo. Calibration was improved by reducing estimated steady-state recharge to 1.28 in/mo. Further evidence that the April-May precipitation and recharge are approximately what is required to maintain the water levels observed in early April indefinitely was obtained by applying the same computational method to data for November-December 1980 and January-March 1981. Larger recharge rates of 1.8 and 2.4 in/mo were obtained for these periods, respectively, and indeed, both river stage and ground-water levels show proportional net rises (fig. 7).

Depth to water was measured between April 7 and 10, 1981 in all wells known to be accessible in the Triple Cities area. (See table 3, at end of report.) Measurements were made by steel tape below a measuring point of known altitude. Level lines were run by the U.S. Geological Survey in 1980-82 or 1967-68 from nearby benchmarks to each well, except for a large group of wells owned by International Business Machines Corp. in Endicott. (Level lines had been run to all wells in that group by a consultant, and duplicate levels to a few wells provided a basis for a small adjustment to make the two sets of altitudes consistent.) A water-table map drawn from the water-level data is presented as plate 4. In some localities, principally near major pumping centers, the potentiometric surface in wells tapping the lower aquifer was significantly lower than the water table in the upper aquifer; these data, although not shown in plate 4, were important in calibrating the model.

Records of pumpage from most large municipal or industrial wells during 1981 were obtained from the owners, and estimates of pumpage from several unmetered production wells were made from reports of typical hours of operation and withdrawal rates.

A steady-state calibration for April 1981 was achieved by adjusting estimated model parameters in successive steps between model runs. Adjustment was made in horizontal and vertical water-transmitting capacity, altitude of the base of the upper aquifer, and, to a lesser extent, recharge from precipitation

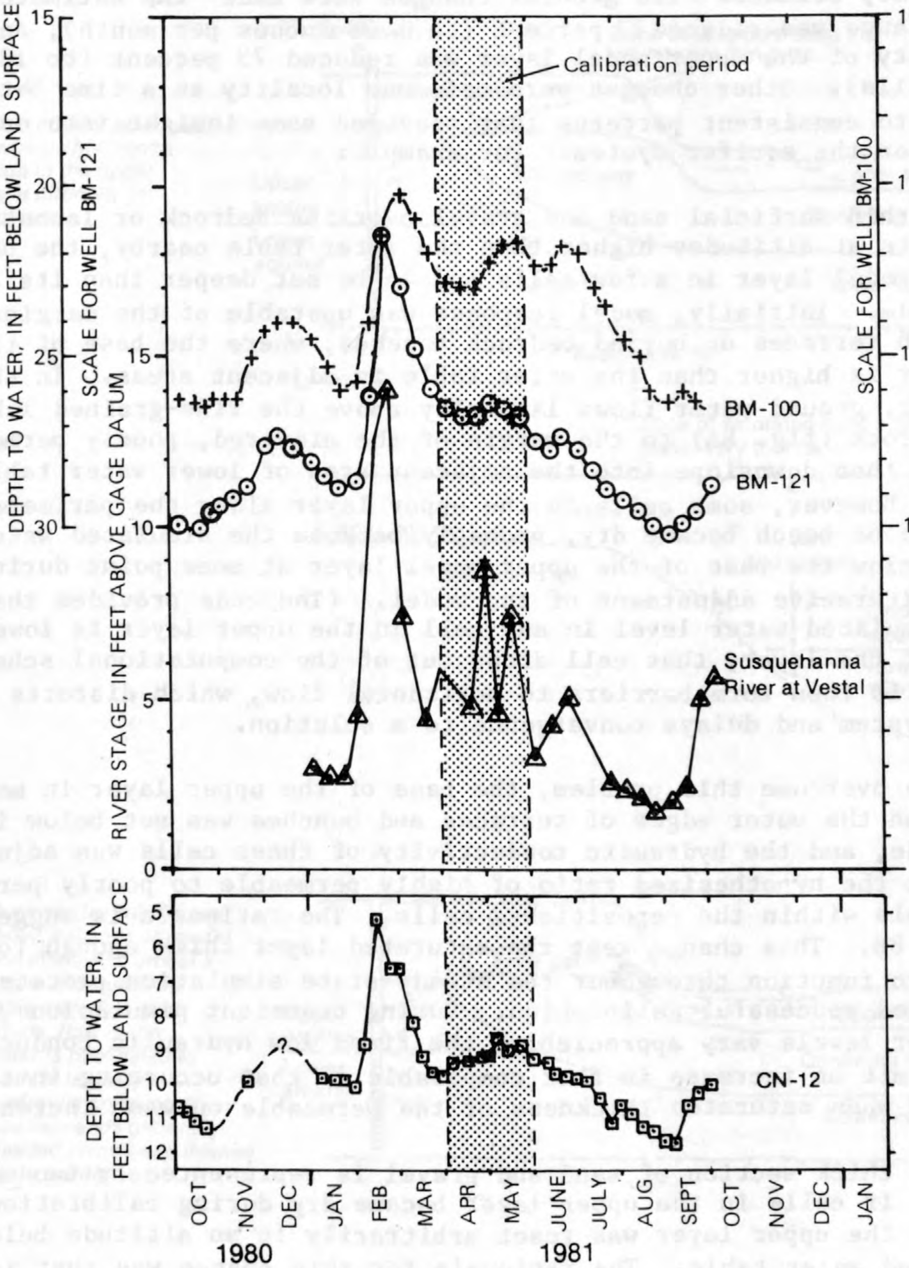


Figure 7.--Water levels and river stage in the Susquehanna River and in three wells, October 1980 through September 1981. Well Bm-100 is in eastern Binghamton (model row 30, column 152); well Bm-121 is in Johnson City (model row 12, column 99); river stage was measured near Bridge Street in Vestal (model row 39, column 41). Well locations and grid are shown on plate 3. River-stage hydrograph depicts approximate weekly average stages; brief peaks are not shown. Well Cn-12 is near the Susquehanna River in Bainbridge, 25 miles northeast of Binghamton.

and tributary streams. Two general changes were made--the estimate of steady-state recharge was reduced 17 percent (to 1.28 inches per month), and hydraulic conductivity of the upper model layer was reduced 75 percent (to 130 ft/d in typical cells). Other changes were made one locality at a time but commonly according to consistent patterns that provided some insight into operation of the model or the aquifer system. For example:

1. Where thin surficial sand and gravel overlies bedrock or lacustrine deposits at altitudes higher than the water table nearby, the base of the upper model layer in a few cells had to be set deeper than its actual altitude. Initially, model response was unstable at the margins of some outwash terraces or buried bedrock benches, where the base of the upper aquifer is higher than the water table in adjacent areas. In the real aquifer, ground water flows laterally above the fine-grained lake sediment or bedrock (fig. 8A) to the margin of the elevated, poorly permeable layer, then downslope into the adjacent area of lower water table. In the model, however, some cells in the upper layer along the perimeter of the terrace or bench became dry, probably because the simulated water level fell below the base of the upper model layer at some point during the early iterative adjustment of the model. (The code provides that whenever the simulated water level in any cell in the upper layer is lower than the base of the layer, that cell drops out of the computational scheme.) The dry cells then form barriers to horizontal flow, which distorts the model flow system and delays convergence to a solution.

   To overcome this problem, the base of the upper layer in many of the cells on the outer edges of terraces and benches was set below its actual altitude, and the hydraulic conductivity of these cells was adjusted to reflect the hypothesized ratio of highly permeable to poorly permeable materials within the repositioned cells. The rationale is suggested by figure 8B. This change kept the saturated layer thick enough for the cells to function throughout the steady-state simulation process and permitted successful calibration. During transient simulations, however, if water levels vary appreciably, the fixed low hydraulic conductivity may not permit an increase in flux comparable to that occurring in the real aquifer when saturated thickness of the permeable outwash increases.

2. Where a thick section of sand and gravel is represented by two model layers, if cells in the upper layer became dry during calibration, the base of the upper layer was reset arbitrarily to an altitude below the simulated water table. The rationale for this change was that in such areas, the base of the upper model layer does not represent a real hydraulic boundary, and subsequent transient simulations might be more realistic if two layers were represented everywhere. Then, transient calibration could test the hypothesis of assigning to the lower layer a small storage coefficient (as commonly suggested by water-level response to short-term pumping) and assigning to the upper layer a large storage coefficient typical of unconfined sand and gravel.

3. Transmissivity of the lower model layer, initially estimated from data from individual sites, had to be reduced by 20 to 50 percent near most pumping centers to reproduce observed water levels. An earlier study (Randall, 1977, p. 41) found a similar contrast between transmissivity estimated from point data and values derived from flow-net analysis. The

A

Idealized sketches of actual settings where thin zones of saturation in the upper layer caused difficulty in model calibration.

B

Flow-path geometry substituted for the above settings in model. Rather than attempt to model a very thin zone of saturation in permeable materials at its actual altitude, flow in the upper aquifer is postulated to follow a path of laterally variable hydraulic conductivity whose base descends gradually. Double-headed arrows and dashed lines indicate position of flow path.

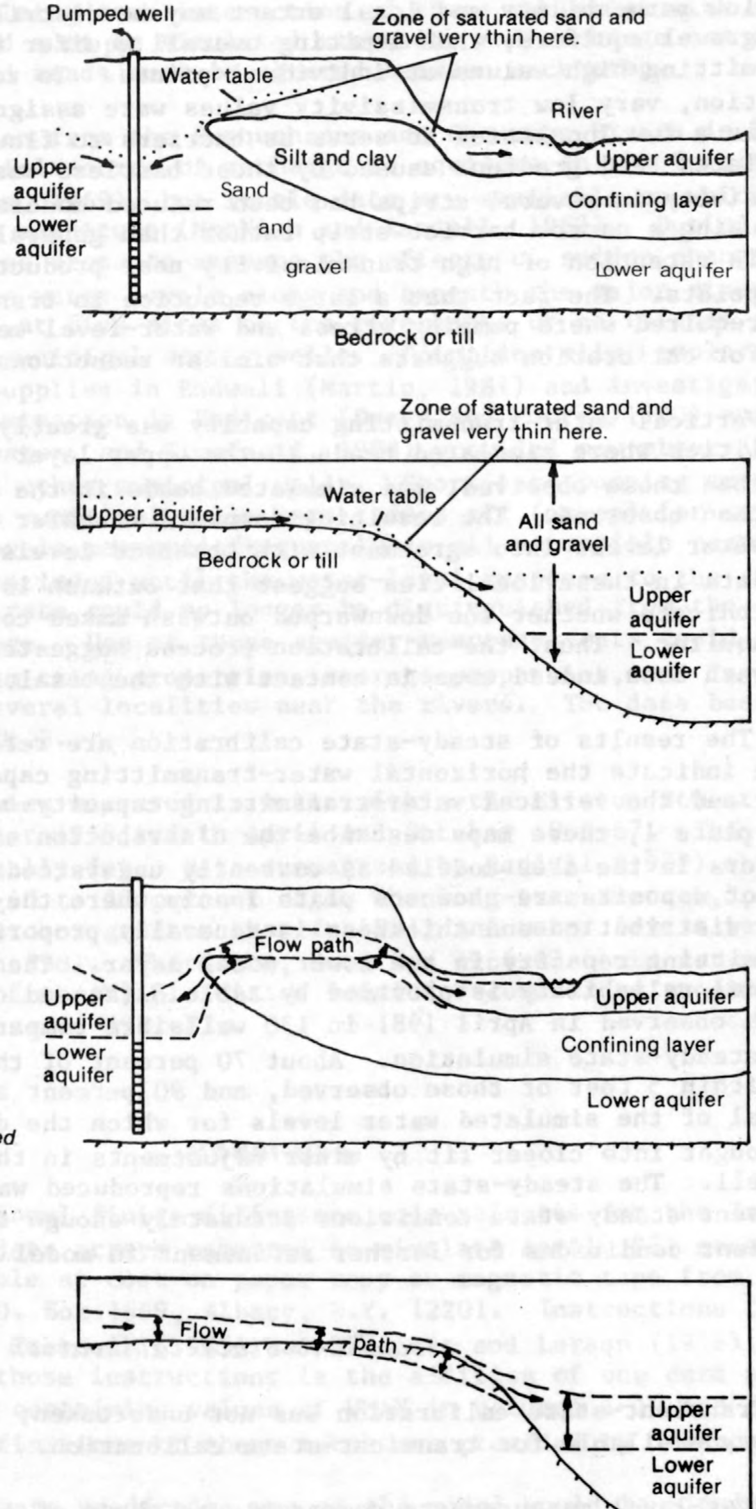


Figure 8.--Rationale for simulation of thin zones of saturation in water-table aquifer.

discrepancy suggested that thin, poorly sorted or fine-grained layers of low permeability and local extent may be distributed within the sand-and-gravel aquifers, thus limiting overall aquifer transmissivity while permitting high values at individual points. To represent this interpretation, very low transmissivity values were assigned to strips one cell wide in a few localities to serve as barriers to flow. The net change in simulated head gradient caused by these barriers was the same as if transmissivity in several strips had been reduced moderately. The advantage of using a narrow barrier strip rather than general reduction of transmissivity is retention of high transmissivity near production wells and other data points. The fact that a large reduction in transmissivity was commonly required where pumping stress and water-level measurements provided control for calibration suggests that similar reductions may be needed elsewhere.

4. Vertical water-transmitting capacity was greatly increased in a few localities where simulated heads in the upper layer were consistently higher than those observed, and simulated heads in the lower layer were lower than observed. The resulting downward transfer of water brought simulated water levels into agreement with observed levels. Geologic or topographic data in these localities suggest that outwash is downwarped but do not indicate whether the downwarped outwash makes contact with a deeper aquifer. Thus, the calibration process suggested that the downwarped outwash does indeed come in contact with the basal aquifer.

The results of steady-state calibration are reflected in plates 2 and 3, which indicate the horizontal water-transmitting capacity of the lower model layer and the vertical water-transmitting capacity between layers. Together with plate 1, these maps describe the distribution and interconnection of aquifers in the area modeled as currently understood. Although the early ice-contact deposits are shown on plate 1 only where they occur at land surface, their distribution and thickness is generally proportional to water-transmitting capacity in the lower model layer. The only quantitative guide to model reliability is provided by table 3 (at end of report), in which water levels observed in April 1981 in 130 wells are compared with those obtained from steady-state simulation. About 70 percent of the simulated water levels are within 5 feet of those observed, and 90 percent are within 8 feet. Several of the simulated water levels for which the deviation was large could be brought into closer fit by minor adjustments in the immediate vicinity of the well. The steady-state simulations reproduced water levels judged to represent steady-state conditions accurately enough to justify turning to transient conditions for further refinement in model calibration.

Transient-State Calibration

Transient-state calibration was not undertaken; however, several sets of data are available for transient-state calibration:

1. Water-level measurements comparable to those of April 1981 (table 3) were made at all accessible wells in October 1981, April 1982, and October 1982. Pumpage and precipitation records for the successive 6-month periods between these dates are also available. The code used for the Triple Cities model does not allow changes during a model simulation in the head in cells that represent river stage nor in recharge rates and pumping rates

in other cells. Therefore, to achieve realistic simulation of the water-level changes resulting from interaction of these variables with storage, each 6-month period must be simulated stepwise, and recharge and discharge rates and specified heads must be altered between each step.

2. Induced infiltration from the Susquehanna and Chenango Rivers is by far the largest potential source of recharge to aquifers in the Triple Cities (MacNish and others, 1969), but little data are available on which to base an estimate of its magnitude (MacNish and Randall, 1982). During this study, opportunities arose to measure the effects of sudden changes in pumping stresses on water levels along and beneath the major rivers. Test wells were drilled at five sites on the riverbanks or on islands in the channel, all near municipal supply wells. Coincidentally, exploration for additional water supplies in Endwell (Martin, 1981) and investigations of ground-water contamination in Endicott (Dames and Moore, 1980) and in Vestal (Martin, Coates, and Timofeeff, 1983) provided new observation wells near several other municipal wells. Short-term pumping tests were conducted in which municipal wells were turned off (or on) for several hours and water levels measured frequently in all accessible nearby wells. Most tests were continued until the water-level response to the controlled change in pumping rate could no longer be distinguished from the effect of changing river stage. Use of these aquifer-response tests could help define water-transmitting properties, storage properties, and induced infiltration in several localities near the rivers. The data base is summarized in table 2.

3. Water levels had been measured in wells within the Clinton Street-Ballpark aquifer in September 1958 and in April and October 1966-67. The years 1966-67 were unusually dry. Data summarized by Randall (1977) show that precipitation was 10 to 20 percent below the long-term average, pumpage was at least 30 percent greater than in 1981, and water levels were many feet lower than in 1981. Therefore, data for 1966-67 could serve to calibrate a part of the Triple Cities model for conditions considerably different from those of 1981-82.

Model Use

The three-dimensional finite-difference code selected for the Triple Cities model and the data arrays required to simulate April 1981 steady-state conditions are available at cost on paper copy or magnetic tape from the U.S. Geological Survey, P.O. Box 1669, Albany, N.Y. 12201. Instructions for use of the code are given in Trescott (1975) and Trescott and Larson (1976). The only modification to those instructions is the addition of one card after group 3, data set 11, containing values of WMAX in columns 1-10 and HMAX(B') in columns 11-20. Definitions of these terms are provided by Trescott (1975).

The most appropriate predictive use of the model would be simulation of average steady-state conditions because the only data used for calibration thus far approximate long-term average conditions. Until the steady-state calibration is refined, manual adjustment (based on data in table 3) of water-level altitudes predicted by the model may be worthwhile in localities where simulated and observed water levels differ significantly (table 3). East of

Table 2.--Pumping tests of riverbank wells in the Triple Cities

[Well locations are shown in plate 3. Only tests observed by U.S. Geological Survey are listed. Additional tests of some of these wells were supervised by the drillers upon well completion, commonly at multiple pumping rates. Data from such tests are available in files of well owners and (or) U.S. Geological Survey, Albany, N.Y.]

Well owner and name or number	Well location in model grid (row, column)	Date of test	Test duration (hours)[a]	Number of observation wells: Same side of river	Number of observation wells: Across river or on island	Pumping rate (gallons per minute)
Crestview Heights 2	56,7	June 22, 1965	12	5	0	395
Endicott Ranney	44,27	Feb. 10, 1948	5 b	3	0	1,420
		June 11, 1981	5 c	2	1	3,700
		April 28, 1982	5 c	0	1	3,700?
Vestal 1-1	39,41	Aug. 31-Sept. 1, 1982	25	22	3	850
1-2	41,38	Sept. 1-2, 1982	22	22	3	670
1-3	40,37	Sept. 2, 1982	8	22	3	1,270
Endicott 5	33,44	June 20, 1981	23	11	1	2,500
Park	32,45	June 21, 1981	12	11	1	1,500
30	27,69	Aug. 3, 1981	23	0	1	700?
		Oct. 5, 1981	7	5	1	700?
36	27,69	Aug. 26-27, 1981	46	4	3	1,300?
Vestal 4-4	15,87	Sept. 21, 1981	8	3	2	1,050
Johnson City 3	16,98	Aug. 1-2, 1981	24	8	2	2,400
GAF 7	34,136	Sept. 3, 1967	22	13	0	575
8, 9	32,132-4	Sept. 4, 1967	10	13	0	600, 950
Binghamton Stow Flats No. 1	23,145	Oct. 10, 1946	4	4	0	870?

a Number of hours pump ran (if normally idle) or shut down (if normally running). Most tests include observations during both recovery and drawdown periods.

b Pumping continued at higher rate, no data.

c Brief well shutdown for repairs of electrical facilities.

the Chenango River and west of Nanticoke Creek, calibration is based on relatively few wells and on natural stresses only, and thus may be less precise than elsewhere.

New production wells are commonly constructed at especially permeable sites selected through test drilling. If a new well were drilled in a part of the aquifer now represented by some average value of transmissivity, simulated pumpage from the new well may result in excessive simulated drawdown. In such cases, model transmissivity could safely be increased in the cell containing that well without requiring recalibration of the entire model.

SUMMARY

1. A digital model of ground-water flow in a 14-mile reach of the Susquehanna River Valley in the Triple Cities of southwestern Broome County, N.Y. has been constructed. This model is available at cost from the U.S. Geological Survey office in Albany, N.Y.

2. In much of the area modeled, extensive deposits of silt and clay separate near-surface coarse sand and gravel aquifers from buried, variably silty gravel aquifers.

3. The model simulates horizontal flow in two dimensions in an upper and a lower layer and vertical flow between the layers.

4. Where the silt and clay are absent and the two aquifers are contiguous, large vertical water-transmitting capacity was simulated; consequently, in these areas the two layers generally have the same heads and function as a single unit. Many areas in which the two aquifers are connected were identified from geologic data; other areas of connection were inferred from adjustments required to make the model simulations conform to observed water levels.

5. The model was calibrated for average steady-state conditions from data for April 1981, when water levels remained fairly stable for nearly 2 months. Recharge from precipitation on the aquifer during this period was determined to be about 1.28 in/mo.

6. Transmissivity of the lower aquifer, as determined by calibration of the model for areas influenced by pumping, was significantly less than transmissivity values calculated from pumping records or geologic logs of individual wells. This difference is attributed to scattered silty layers that reduce average transmissivity of generally permeable materials.

7. Data are available for transient calibration, including (1) semiannual water-level measurements that reflect seasonal changes in river stage, pumping, and recharge; (2) short-term pumping tests at several sites, and (3) measurements in the Clinton Street-Ballpark aquifer before 1967, when pumpage was 30 percent greater than in 1981. Transient calibration would improve the accuracy and predictive capability of the model.

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Table 3.--Comparison of measured and simulated steady-state water levels in the Susquehanna River valley, Broome County, N.Y.

Well owner and number or name	Model layer[a]	Model cell[b] (row,column)	Water-level altitude April 7-10, 1981 (feet above sea level) Observed	Simulated	Difference
		West Endicott			
IBM Glendale GR2	1	29,21	815.62	806.32	-9.30
GR4	1	22,22-3	824.78	808.85	-15.93
Rose/Batch Plant	1?	32,23	805.25	806.06	+.81
Oliver Main	1	20,29	809.50	812.34	+2.84
Endicott Kelly well	1	43,27	768.78	795.75	+26.97
		Endicott			
Endicott Motor Inn	1	25,48	803.95	802.83	-1.12
Endicott Trust	1	20,49	816.31	813.01	-3.30
IBM END-1	1	25,46	805.69	803.49	-2.20
END-2	1	23,50	803.87	804.01	+.14
END-3	1	28,46	802.12	800.28	-1.84
Cafeteria	1	18,51	823.32	817.37	-5.95
Endicott 6	1	33,44	758.30	758.54	+.24
23	2?	32,44	794.25	804.55	+10.30
Stable	2	30,50	804.5 est.	805.24	+.74
Stable 8" TW	1	do.	802.54	802.99	+.45
USGS Mercerau Park	1	37,36	803.64	801.90	-1.74
IBM EN59	2	23,43-4	819.48	817.71	-1.77
EN61	2	25,46	820.89	818.61	-2.28
EN63	2	25-6,47-8	819.79	821.89	+2.10
EN62	2	22,47-8	822.65	827.64	+4.99
EN64	2	23,49	823.09	827.22	+4.13
EN22	2	23,50	822.89	827.30	+4.41
EN95	2	19,44	820.71	824.62	+3.91
EN60	2	20-1,46	822.29	827.10	+4.81
EN94	2	19,47	822.30	829.20	+6.90
EN91	2	19,48	823.91	830.01	+6.10
EN92	2	19-20,49	822.98	830.30	+7.32
EN93	2	20-1,49	822.68	829.69	+7.01
EN29	2	19-20,49-50	823.07	830.49	+7.42
EN6	2	19-20,50-1	822.09	830.77	+8.68
EN77	2	17-8,50-1	827.68	832.69	+5.01
EN78	2	18-9,51	827.50	831.86	+4.36
EN79	2	21,51-2	822.97	829.25	+6.28
EN80	2	19-20,51-2	827.34	830.65	+3.31
EN83	2	14,52	833.29	838.68	+5.39
EN84	2	12,52	839.82	842.27	+2.45
EN23	2	17-8,51-2	827.89	832.62	+4.73

[a] 1, lower; 2, upper.
[b] Cell locations are shown in plate 3. Hyphenated numbers indicate well is close to boundary between two rows or columns.

Table 3.--Comparison of measured and simulated steady-state water levels in the Susquehanna River valley, Broome County, N.Y.--continued.

Well owner and number or name	Model layer[a]	Model cell[b] (row,column)	Water-level altitude April 7-10, 1981 (feet above sea level) Observed	Simulated	Difference
		Endicott (continued)			
IBM EN65	2	19,53	828.79	829.74	+.95
EN30	1	16,55-6	829.52	823.00	-6.52
EN89	2	14-5,56-7	835.26	834.38	-.88
EN87	2	16,53-4	829.96	831.40	+1.44
EN86	2	15,53-4	833.62	835.03	+1.41
EN82	2	16,52	828.63	834.32	+5.69
EN35	2	16-7,51	829.40	833.91	+4.51
EN81	2	16-7,51	830.01	833.91	+3.90
EN32	2	15,51	831.64	835.74	+4.10
EN58	2	14-5,51	831.34	836.21	+4.87
EN34	2	14,50-1	830.98	836.25	+5.27
EN21	2	16,50	831.60	834.18	+2.58
EN19	2	17,50	829.95	833.07	+3.12
EN24	2	17,48-9	827.07	832.33	+5.26
EN76	2	17-8,47-8	826.68	830.54	+3.86
EN74	2	15-6,48	829.63	834.38	+4.75
EN27	2	15,50	832.11	835.21	+3.10
EN98	2	14,48	835.70	835.67	-.03
EN72	2	14,47-8	829.09	835.99	+6.90
EN26	2	15,47	828.69	835.33	+6.64
EN71	2	14,46-7	828.57	836.88	+8.31
EN70	2	15-6,46	827.25	834.75	+7.50
EN66	2	16,44-5	822.12	832.41	+10.29
EN96	2	16-7,43-4	819.23	829.02	+9.79
EN69	2	14-5,45-6	827.75	836.74	+8.99
IBM Well Field 2	1	26,56	796.81	804.35	+7.54
3	1	do.	796.88	804.35	+7.28
Endicott B2	2	27,65-6	805.25	806.69	+1.44
B7	1	28,65	809.31	806.92	-2.39
		Vestal			
Castle Gardens	1	49,26	802.07	803.21	+1.14
James Dittrich	2	47,27	802.1 est.	802.25	+.15
Vestal 1-3 (1-14)	1	40,37	803.90	801.94	-1.96
Vestal 4-1	1	36,71	814.96	811.36	-3.60
H. J. Russell	1	32,70	807.10	808.74	+1.64
Vestal B-3	1	17,85	794.44	802.68	+8.24
B-6	1	16,85	803.64	804.03	+.39
B-2	1	16-7,86	796.03	802.99	+7.02
Barney Dickenson	1	19,83	799.19	802.75	+3.56
DEC TH-1	1	21,82	801.68	804.51	+2.83
TH-2	1	do.	799.12	804.51	+5.39

Table 3.--Comparison of measured and simulated steady-state water levels in the Susquehanna River valley, Broome County, N.Y.--continued.

Well owner and number or name	Model layer[a]	Model cell[b] (row,column)	Water-level altitude April 7-10, 1981 (feet above sea level) Observed	Simulated	Difference
		Vestal (continued)			
Monarch A	2	do.	799.72	804.59	+4.88
B	2	21-2,82	800.23	804.59	+4.36
R. Murphy shop	1	21-2,84-5	799.88	803.72	+3.84
Nichols	1	24,81	804.38	807.41	+3.03
NYG & E 1	2	24-5,86-7	820.32	825.99	+5.67
2	2	23-24,88	839.56	833.23	-6.33
3	2	23,86	802.78	812.19	+9.41
USGS Prentice Road Island 2 1/2"	2	14,85	808.4 est.	808.4	0
6"	1	do.	808.0 est.	805.45	-2.55
		Clinton Street-Ballpark Aquifer			
IBM Country Club 2	2?	6,94	814.96	810.83	-4.13
Johnson City 1	1	16,98	799	793.13	-6
2	1	do.	794	793.13	-1
USGS Hill Park 1"	2	15,96	812.81	810.98	-1.83
6"	1	do.	807.43	804.15	-3.28
USGS Bm-121	2	12,99	809.09	807.02	-2.07
Bm-120	2?	5,99	820.41	811.80	-8.61
Johnson City Fifth Street	1	11,102	806.79	807.20	+.41
6	1	12,107	811.42	809.06	-2.36
4	1	13,109	811.60	807.84	-3.76
General Electric 2	1	16,100	808.32	803.70	-4.62
Goudey Station 1	2	20-1,100	815.68	811.93	-3.75
2	2	20-1,101-2	818.27	811.57	-6.50
3	2	17-8,100-1	811.33	805.90	-5.43
Wilson Hospital	1	19-20,108-9	822.33	816.42	-5.91
USGS Bm-119	2?	18-9,102-3	818.98	811.86	-7.12
Bm-118	1	21,107	823.91	817.04	-6.87
Bm-117	2	39,122	850.8	841.57	-9.23
Bm-116	2	26-7,115	824.02	824.23	+.21
Bm-101	2	33,117	844.97	844.58	-.39
Bm-111	2	20-1,118-9	818.54	820.79	+2.25
Bm-114	1	38,128	813.95	817.35	+3.40
Bm-113	2	28,129	811.09	816.57	+5.48
Bm-115	1	32,135	813.80	817.46	+3.66
Well K	1	35,134	822.03	822.38	+.35
Well I	1	35,137	823.28	824.76	+.48
Well J 6"	1	35,139	826.10	826.59	+.49
Well J 2 1/2"	2	do.	825.89	826.65	+.76
Fairbanks Co.	1	29,119	816.12	817.65	+1.53

Table 3.--Comparison of measured and simulated steady-state water levels in the Susquehanna River valley, Broome County, N.Y.--continued.

Well owner and number or name	Model layer[a]	Model cell[b] (row,column)	Water-level altitude April 7-10, 1981 (feet above sea level) Observed	Simulated	Difference
			Clinton Street-Ballpark Aquifer (continued)		
Anitec 27T	1	25,117	822.0	818.60	-3.40
24T	1	23,119	814.1	817.53	+3.43
23T	1	23,120	813.3	816.95	+3.65
21T	1	24,121	813.4	816.16	+2.86
25T	1	26,119	812.2	817.41	+5.21
3	1	28,124	812.2	816.34	+4.14
4	1	27,125	812.0	816.28	+4.28
Anitec 2A	1	28,125	813.1	816.25	+3.15
10	1	32,129	797.30	804.89	+7.59
6	1	32,130	810.32	810.34	+.02
9	1	32,132	804.11	808.92	+4.81
8	1	32,134	812.42	816.34	+3.92
7	1	34,136	822.93	821.26	-.67
33T	1	37,135-6	824.07	824.5	+.43
USGS Harry L. Drive	2	9-10,114-5	839.79	833.07	-6.72
			East of Chenango River		
USGS Bm112	2?	40,138	826.35	826.99	+.64
Bm107	1	36,148	830.77	832.45	+1.68
Bm106	1	47-8,157	840±	834.94	-5
Bm100	2?	30,152	839.45	841.38	+1.93

Randall AQUIFER MODEL OF THE SUSQUEHANNA RIVER VALLEY IN SOUTHWESTERN BROOME COUNTY, NEW YORK USGS/WRI 85-4099