Prepared in cooperation with the Livingston County Department of Health

Simulated Effects of 1994 Salt-Mine Collapse on Ground-Water Flow and Land Subsidence in a Glacial Aquifer System, Livingston County, New York

Professional Paper 1611

Courtesy of Ron Pretzer/LUXE, May 1994

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

Availability of Publications of the U.S. Geological Survey

Order U.S. Geological Survey (USGS) publications by calling the toll-free telephone number 1-888-ASK-USGS or contacting the offices listed below. Detailed ordering instructions, along with prices of the last offerings, are given in the current-year issues of the catalog "New Publications of the U.S. Geological Survey."

Books, Maps, and Other Publications

By Mail

Books, maps, and other publications are available by mail from—

USGS Information Services
Box 25286, Federal Center
Denver, CO 80225

Publications include Professional Papers, Bulletins, Water-Supply Papers, Techniques of Water-Resources Investigations, Circulars, Fact Sheets, publications of general interest, single copies of permanent USGS catalogs, and topographic and thematic maps.

Over the Counter

Books, maps, and other publications of the U.S. Geological Survey are available over the counter at the following USGS Earth Science Information Centers (ESIC's), all of which are authorized agents of the Superintendent of Documents:

- Anchorage, Alaska–Rm. 101, 4230 University Dr.
- Denver, Colorado–Bldg. 810, Federal Center
- Menlo Park, California–Rm. 3128, Bldg. 3, 345 Middlefield Rd.
- Reston, Virginia–Rm. 1C402, USGS National Center, 12201 Sunrise Valley Dr.
- Salt Lake City, Utah–2222 West, 2300 South
- Spokane, Washington–Rm. 135, U.S. Post Office Building, 904 West Riverside Ave.
- Washington, D.C.–Rm. 2650, Main Interior Bldg., 18th and C Sts., NW.

Maps only may be purchased over the counter at the following USGS office:

- Rolla, Missouri–1400 Independence Rd.

Electronically

Some USGS publications, including the catalog "New Publications of the U.S. Geological Survey" are also available electronically on the USGS's World Wide Web home page at **http://www.usgs.gov**

Preliminary Determination of Epicenters

Subscriptions to the periodical "Preliminary Determination of Epicenters" can be obtained only from the Superintendent of Documents. Check or money order must be payable to the Superintendent of Documents. Order by mail from—

Superintendent of Documents
Government Printing Office
Washington, DC 20402

Information Periodicals

Many Information Periodicals products are available through the systems or formats listed below:

Printed Products

Printed copies of the Minerals Yearbook and the Mineral Commodity Summaries can be ordered from the Superintendent of Documents, Government Printing Office (address above). Printed copies of Metal Industry Indicators and Mineral Industry Surveys can be ordered from the Center for Disease Control and Prevention, National Institute for Occupational Safety and Health, Pittsburgh Research Center, P.O. Box 18070, Pittsburgh, PA 15236–0070.

Mines FaxBack: Return fax service

1. Use the touch-tone handset attached to your fax machine's telephone jack. (ISDN [digital] telephones cannot be used with fax machines.)
2. Dial (703) 648–4999.
3. Listen to the menu options and punch in the number of your selection, using the touch-tone telephone.
4. After completing your selection, press the start button on your fax machine.

CD-ROM

A disc containing chapters of the Minerals Yearbook (1993–95), the Mineral Commodity Summaries (1995–97), a statistical compendium (1970–90), and other publications is updated three times a year and sold by the Superintendent of Documents, Government Printing Office (address above).

World Wide Web

Minerals information is available electronically at **http://minerals.er.usgs.gov/minerals/**

Subscription to the catalog "New Publications of the U.S. Geological Survey"

Those wishing to be placed on a free subscription list for the catalog "New Publications of the U.S. Geological Survey" should write to—

U.S. Geological Survey
903 National Center
Reston, VA 20192

Simulated Effects of Salt-Mine Collapse on Ground-Water Flow and Land Subsidence in a Glacial Aquifer System, Livingston County, New York

By **Richard M. Yager, Todd S. Miller, and William M. Kappel**

Professional Paper 1611

Prepared in cooperation with the Livingston County Department of Health

U.S. DEPARTMENT OF THE INTERIOR

GALE A. NORTON, Secretary

U.S. GEOLOGICAL SURVEY
Charles G. Groat, Director

Reston, Virginia 2001

Library of Congress Cataloging-in-Publications Data

Yager, Richard M.
Simulated effects of salt-mine collapse on ground-water flow and land subsidence in a glacial aquifer system, Livingston County, New York / by Richard M. Yager, Todd S. Miller, and William M. Kappel.
p. cm. – (Professional paper ; 1611)
Includes bibliographical references (p.).
ISBN 0-607-95033-1
1. Groundwater flow–New York (State)–Livingston County–Comp;uter simulation. 2. Subsidences (Earth movements–New York (State)–Livingston County–Computer simulation. I. Miller, Todd S. II. Kappel, William M. III. Geological Survey (U.S.) IV. Livingston County (N.Y.). Dept. of Health. V. Title. VI. U.S. Geological Survey professional paper ; 1611.

QE75 .P9 v. 1611
[GB1025.N]
557.3 s
[551.49'0974785] 00-060049

For sale by U.S. Geological Survey,
Branch of Information Services
Box 25286, Federal Center
Denver, CO 80225

CONTENTS

FIGURES

TABLES

APPENDIX

CONVERSION FACTORS AND ABBREVIATIONS

Multiply	By	To obtain
Length		
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
Area		
acre	0.4047	hectare
square mile (mi^2)	2.590	square kilometer
Flow rate		
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
gallon per minute (gal/min)	0.06309	liter per second
Temperature		
degrees Farenheit (˚F)	5/9 (˚F-32)	degrees Centigrade (˚C)
Density		
pound per cubic foot (lb/ft^3)	16.02	kilogram per cubic meter
Pressure		
ton per square foot (ton/ft^2)	95.77	kilopascal
pound per square inch (psi)	6.895	kilopascal
Hydraulic conductivity		
foot per day (ft/d)	0.3048	meter per day
Transmissivity		
foot squared per day (ft^2/d)	0.09290	meter squared per day
Specific storage		
per foot (ft^{-1})	3.281	per meter
Other abbreviations		
picocurie per millimeter	pCi/ml	
microsiemens per centimeter	µS/cm	

Sea level: In this report, “sea level” refers to the National Geodetic Vertical Datum of 1929—a geodetc datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

Simulated Effects of 1994 Salt-Mine Collapse on Ground-Water Flow and Land Subsidence in a Glacial Aquifer System, Livingston County, New York

By Richard M. Yager, Todd S. Miller, and William M. Kappel

ABSTRACT

The bedrock ceiling in parts of the Retsof salt mine in the the Genesee Valley, Livingston County, N.Y. collapsed on March 12, 1994 and water from overlying aquifers began to flow into the mine at a rate of 5,500 gal/min (gallons per minute), and the rate increased to as much as 20,000 gal/min after a second collapse 3 weeks later in April 1994. Efforts to save the mine were abandoned by the end of 1994, and the mine was completely flooded by January 1996.

The Genesee Valley (including the tributary Canaseraga Valley) is about 40 miles long and contains as much as 750 feet of glacial sediment and recent alluvium, which rest on mostly carbonate bedrock along the valley axis near and north of the collapse site, and on shale in areas farther south. The mined salt bed lies within Silurian shale and is overlain by 600 feet of Devonian shale and limestone. Rock-rubble zones and fractures that formed within units overlying the collapse resulted in about 45 feet of subsidence at the bedrock surface and the formation of sinkholes as much as 70 feet deep at land surface.

The glacial aquifer system in the Genesee Valley generally consists of an unconfined alluvial aquifer (upper) and two confined glacial aquifers (middle and lower) separated by two confining layers. The underlying carbonate bedrock also contains water-bearing fracture zones in which water flowed northward before the collapse. The salinity of ground water generally increases with depth; specific conductance ranges from 430 to 1,390 μS/cm (microsiemens per centimeter) in the middle aquifer and from 2,980 to 39,300 μS/cm in the lower aquifer. Concentrations of tritium in the lower aquifer indicate the water to be generally more than 45 years old.

Subsidence at land surface ranges from 15 feet over some parts of the mine (through shrinkage of the mine cavity in response to deformation of salt beds) to more than 70 feet over two areas within the mine, and is as much as 0.8 ft outside the mined area as a result of compaction of fine-grained sediments in the confining layers. By January 1996, when the mine was completely flooded, drawdowns in the lower aquifer were as much as 400 feet near the collapse area, as much as 50 feet in an area 7 miles to the north, and as much as 135 feet in an area 8 miles south of the collapse area. The northward direction of ground-water flow in the confined aquifers north of the collapse was reversed, and five domestic wells 7 miles north of the collapse began to produce saline water from the lower aquifer, apparently because the drawdown either induced the inflow of saline water from bedrock, or allowed saline water from further north to flow southward. Ratios of chloride to bromide concentrations in ground-water samples indicate two sources of salinity in the lower aquifer: (1) saline water from carbonate bedrock and (2) halite-bearing water from an unknown source. Both thermogenic and biogenic gases (methane and sulfide) were produced by wells screened in confined aquifers as a result of exsolution as water-level declines lowered the pore pressure.

A five-layer ground-water flow model based on generalized geologic sections was used to simulate flow conditions within the unconsolidated sediments before and after the mine collapse. The model represents a 61.3 square-mile section of the Genesee Valley that extends from the Village of Avon in the north to the Village of Dansville, 40 miles in the south. Nonlinear regression was used to calibrate the model to (1) assumed water-table altitudes and estimated base flows in steady-state simulations representing precollapse conditions, and (2) measured water levels in wells and estimated ground-water discharges to the mine in transient-state simulations that represent the flooding of the mine and water-level recovery thereafter.

The computed water budget indicated that ground water released from storage in the lower aquifer accounted for about 11 percent of the total inflow to the model, and most of the water discharged to the mine. Values of specific storage estimated by regression for the two confined aquifers (2.9×10^{-4} and 6.9×10^{-5} ft^{-1}) were much larger than values typically estimated for other sand and gravel aquifers affected by land subsidence (1×10^{-6} to 3×10^{-6} ft^{-1}); this could be explained by the exsolution of natural gas (methane) from ground water during the period of drawdown; exsolution also could account for part of a bias in the head distribution computed in model simulations. A gas-partitioning equation developed to compute specific storage in the presence of a gas phase indicates that specific storage was not constant, as assumed in the flow model, but varied temporally and spatially as a function of pressure. Computations of specific storage in the lower aquifer near the collapse area, where the maximum drawdown occurred, included the effects of gas exsolution and ranged from 2.4×10^{-5} to 1.8×10^{-4} ft^{-1}; the maximum pore volume occupied by gas was estimated to be 7 percent. Including (1) gas exsolution and (2) aquifer compressibility typically associated with aquifer-system compaction, decreased model error by about 15 percent and slightly reduced model bias.

Hydraulic heads computed in 12.4-year transient-state simulations indicate that water levels will return to precollapse conditions by about the year 2006, but the recovery would be completed by 1999 if the specific storage were smaller (2.3×10^{-6} ft^{-1}). Ground-water flow paths computed from model results indicate that the reversal of the northward hydraulic gradient in the northern part of the valley allows saline water to migrate southward toward the collapse at a velocity of 1 to 2 ft/d (feet per day). The time between the collapse and the onset of salinity in two domestic wells indicates a greater velocity (8 ft/d), however, and suggests that the saline water could have traveled along discrete flowpaths with a larger hydraulic conductivity (2,000 ft/d) than the 300 ft/d estimated by regression for the lower aquifer. The present southward gradient (1.2×10^{-3} ft/ft) is about 3 times larger than the northward gradient (3.9×10^{-4} ft/ft) that will develop once the natural hydraulic gradient is reestablished. The calculated northward velocity indicates that, if saline water migrates southward for 10 years after the mine collapse, about 30 years will be required for freshwater to flush the area into which the saline water has intruded.

Transient-state simulations with one- and three-dimensional models were conducted to estimate the extent of land subsidence due to dewatering of fine-grained sediments outside the mined area, where subsidence measurements were not made. The one-dimensional models represent vertical flow and compression in the upper and lower confining layers at two locations—one near a borehole with pressure transducers screened in the upper confining layer near the collapse area, and the second at a survey monument 2,800 ft to the south. Vertical hydraulic conductivity of the upper confining layer, and specific storage of both confining layers, were estimated through nonlinear regression from observations of pore pressure and subsidence. Both elastic and inelastic storage were represented in both confining layers. Elastic storage was assumed, from an observed relation between compressibility and depth computed from consolidation tests of confining-layer sediments, to be inversely proportional to effective stress (or depth); inelastic storage was assumed, from published values, to be 30 times greater than elastic storage.

Computed pressure changes in the upper confining layer near the collapse were in close agreement with measured pressures changes, and the maximum computed subsidence was about 94 percent of the observed subsidence (0.78 feet). About 92 percent of the computed subsidence was the result of drainage and compression in the lower confining layer; 90 percent of this compression was estimated to be inelastic, nonrecoverable compaction. Representing drainage of confining-layer sediments by elastic storage gave pressure and maximum-subsidence values that matched observed values as closely as those based on inelastic storage; elastic storage also simulated land-surface rebound after water-level recovery, which had not occurred by 1999. Hydraulic properties of the confining-layer sediments estimated by the one-dimensional models that represented only elastic compression were incorporated in the three-dimensional model and resulted in close agreement between computed subsidence and subsidence measured along a survey line. Simulated subsidence closely matched the actual subsidence in Mt. Morris, 3 mi south of the collapse, where as much as 0.3 ft of subsidence was measured. Model results indicate that at least 0.1 ft of subsidence occurred by February 1996 over the 16 square-mile area that extended 5 miles north of the collapse area and 7 miles south of the collapse.

INTRODUCTION

A collapse of the bedrock ceiling in parts of the Retsof salt mine in Livingston County, N.Y. (fig. 1) in March and April 1994 allowed water from overlying aquifers to flood the mine. All mining activity was halted in September 1995, and the mine shafts were sealed. The mine was completely flooded by January 1996, eliminating a major commercial source of rock salt. The flooding of the mine caused widespread water-level declines in the overlying aquifer system, which lies within the valley drained by the Genesee River and Canaseraga Creek in Livingston County; this is referred to herein as the Genesee Valley study area.

The Retsof salt mine is about 30 mi southwest of the City of Rochester (fig. 1) and adjoins the western

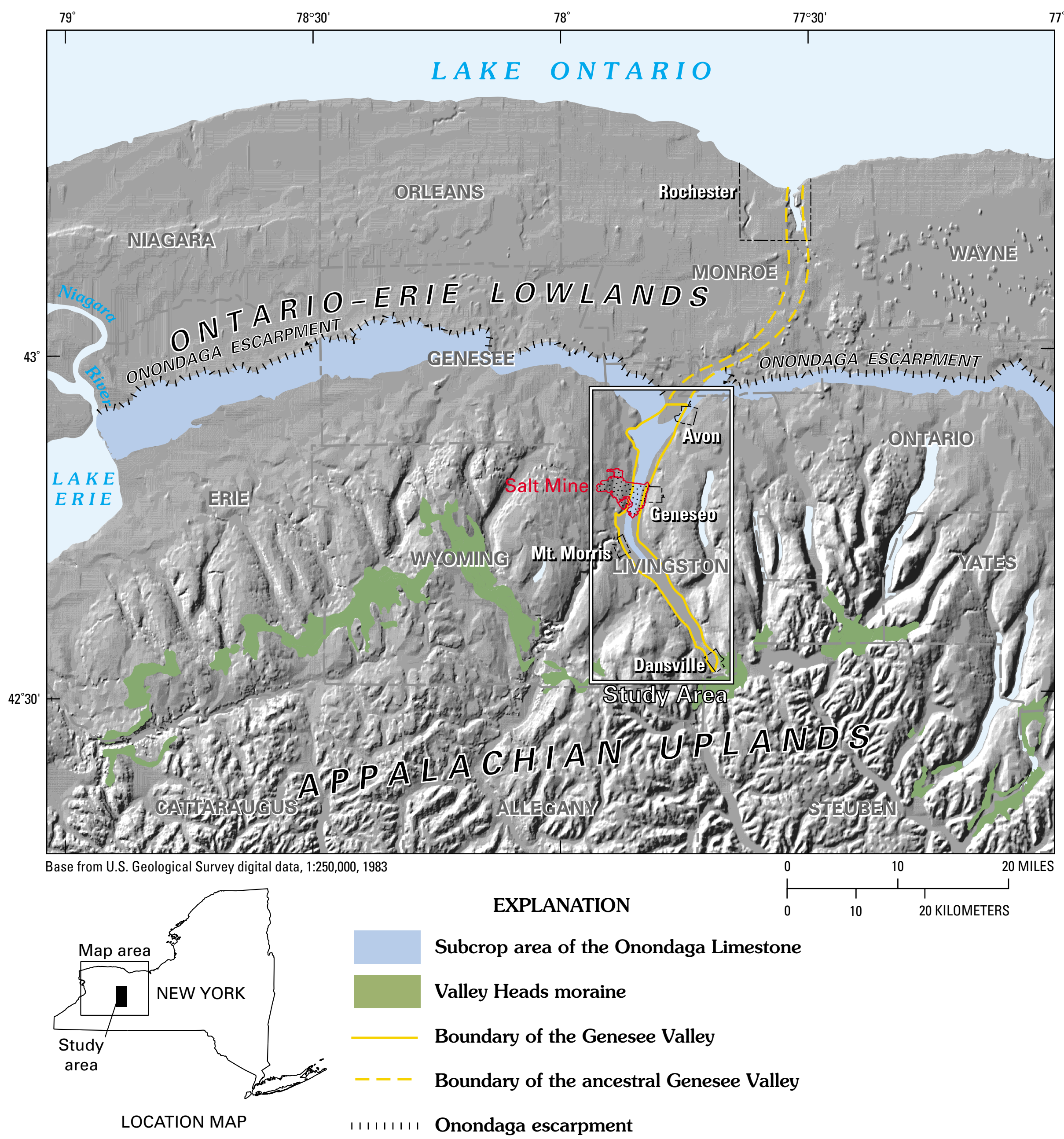


Figure 1. Locations of selected geologic features in western New York and of study area.

edge of the Village of Geneseo (fig. 2). The mine began operations in 1885 and became the largest rock-salt mine in the western hemisphere; it now underlies an area of more than 6,000 acres (about 11 mi^2). Most of the mined salt was used for road deicing in 14 northeastern States. The abandoned Sterling mine (fig. 2) was connected to the Retsof mine to improve ventilation.

On March 12, 1994, the mine ceiling collapsed in a 650- by 650-ft mining panel (2 Yard South) which consists of 40 ft-wide rooms separated by 25-ft-square pillars located 1,100 ft below land surface (fig. 2). The collapse allowed water from overlying aquifers to drain into the mine at a rate of about 12 ft^3/s (7.9 Mgal/d). Less than a month later, on April 8, 1994, the bedrock overlying a nearby panel (11 Yard West) collapsed, and total ground-water inflow to the mine increased to 47 ft^3/s (30.2 Mgal/d), which is equivalent to 5.6 percent of all ground water pumped by municipalities (535 Mgal/d) in New York State (Solley and others, 1993).

Subsidence of rock and sediment above the collapsed areas propagated to land surface, causing two 300-ft-diameter sinkholes (fig. 3) to develop, around which ring fractures developed with vertical offsets ranging from 1 to 10 ft. Land surface subsidence was as much as 20 ft in the first sinkhole and as much as 70 ft in the second. The sinkholes caused the closure of a highway bridge over Beards Creek on State Route 20A (fig. 2); an alternate route was established to detour traffic around the collapse area.

Water levels in domestic and industrial wells within the valley declined rapidly within eight weeks after the collapses, and at least a dozen wells went dry. By January 1996, when the mine was completely flooded, ground-water levels had declined about 400 ft near the collapse area and more than 50 ft in areas 7 mi to the north and 8 mi to the south within the valley. Since 1994, the mine owner has supplied water to residents whose wells went dry, has monitored water levels in a network of more than 100 wells and has measured land subsidence at monument surveys. A Memorandum of Understanding among the mine owner, Livingston County, and the State of New York in December 1994 provided funds for test-well drilling to characterize the aquifer system and to support the development of a ground-water flow model.

In 1995, the U. S. Geological Survey (USGS), in cooperation with the Livingston County Department of Health, began a 2-year study to define the effect of the ceiling collapse on the Genesee Valley aquifer system. The study used information from field surveys, including (1) stratigraphic logs from test drilling (Alpha Geoscience, 1995a,b,c and 1996a,b,c); (2) seismic-reflection and refraction surveys, (3) drillers' logs and records of ground-water levels; (4) measurements of land-surface altitude; and (5) chemical analyses of ground-water samples to develop a three-dimensional ground-water-flow model of the aquifer system. The model results were compared with actual water-level declines to simulate (1) the time required for water levels to return to precollapse conditions and (2) the extent of land subsidence caused by compression of fine-grained sediments.

Purpose and Scope

This report describes the hydrogeology of the Genesee Valley and the effects of the ceiling collapse on the aquifer system. It discusses (1) the origin and character of glacial sediments, (2) the occurrence, flow directions, and chemical quality of water in the aquifer system before the collapse, (3) the effects of the collapse on the aquifer system in terms of land subsidence, water-level declines, changes in water quality, and exsolution of natural gas, and (4) design and calibration of the ground-water-flow model; it also presents (1) results of flow-model simulations, including an estimated ground-water budget and graphs showing the simulated water-level recovery, and (2) results of subsidence simulations, including maps and graphs showing the extent of land subsidence.

Previous Studies

Ground-water resources of the Genesee River Basin were described by Kammerer and Hobba (1967) in a report that includes drillers' logs and chemical analyses of water. The history of glacial deposition in the area has been described by Muller and others (1988) and by Brennan (1988) and Young and Sitkin (1994).

Various engineering studies addressed leakage of ground water into an access shaft to the abandoned Sterling mine. Langill and Associates (1981) traced the leakage to joints and bedding-plane fractures at the contact between the Onondaga and Bertie Limestones and identified the source of water as the glacial aquifer underlying the Genesee River. Dunn Geoscience Corporation (1992) discussed the hydrogeology of uncon-

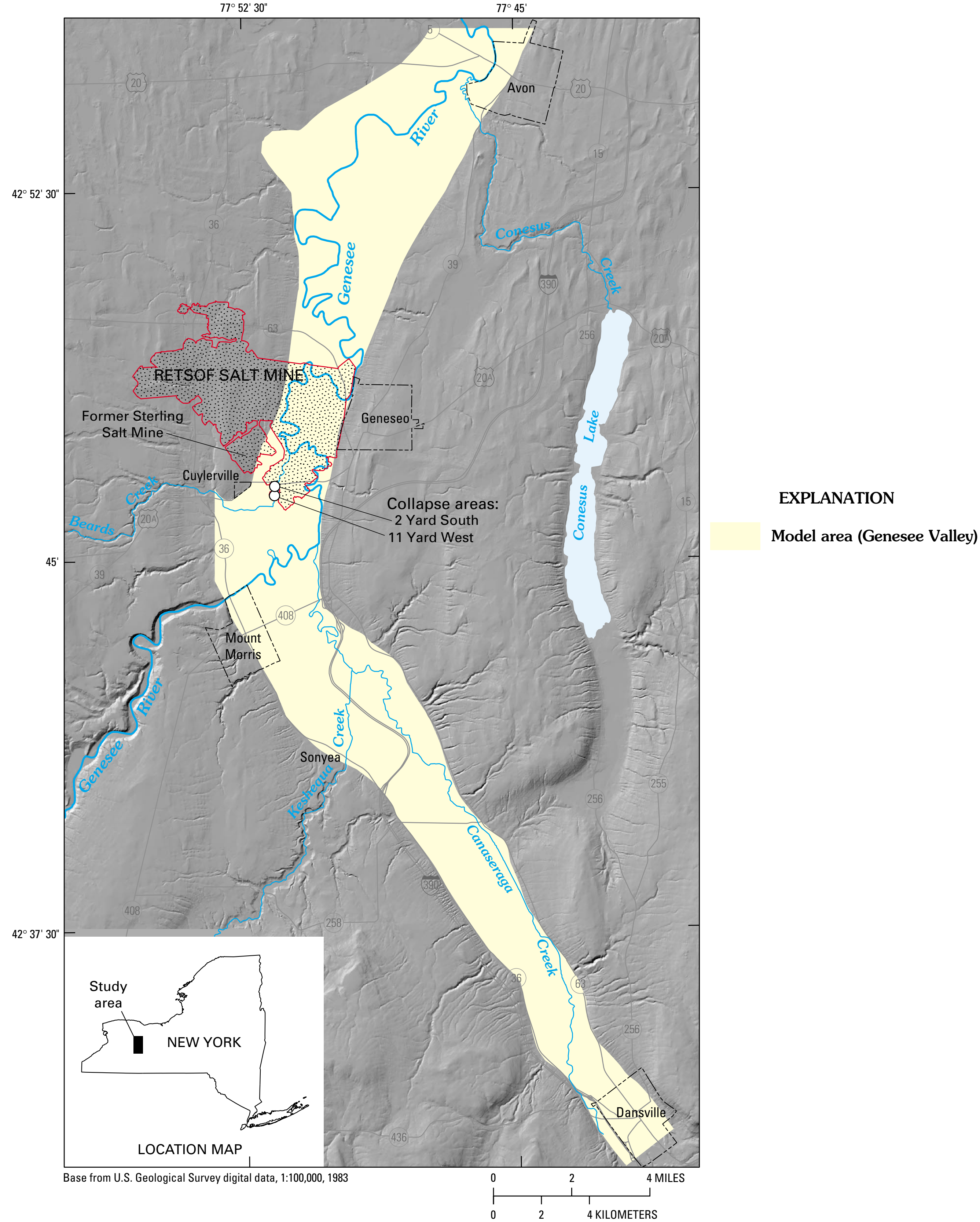


Figure 2. Principal geographic features of Genesee Valley study area, Livingston County, N.Y., and location of mined area and of collapse. (Location is shown in fig. 1.)

solidated deposits and bedrock in the vicinity of the Retsof and Sterling mines as part of a proposal for an incinerator-ash-processing facility.

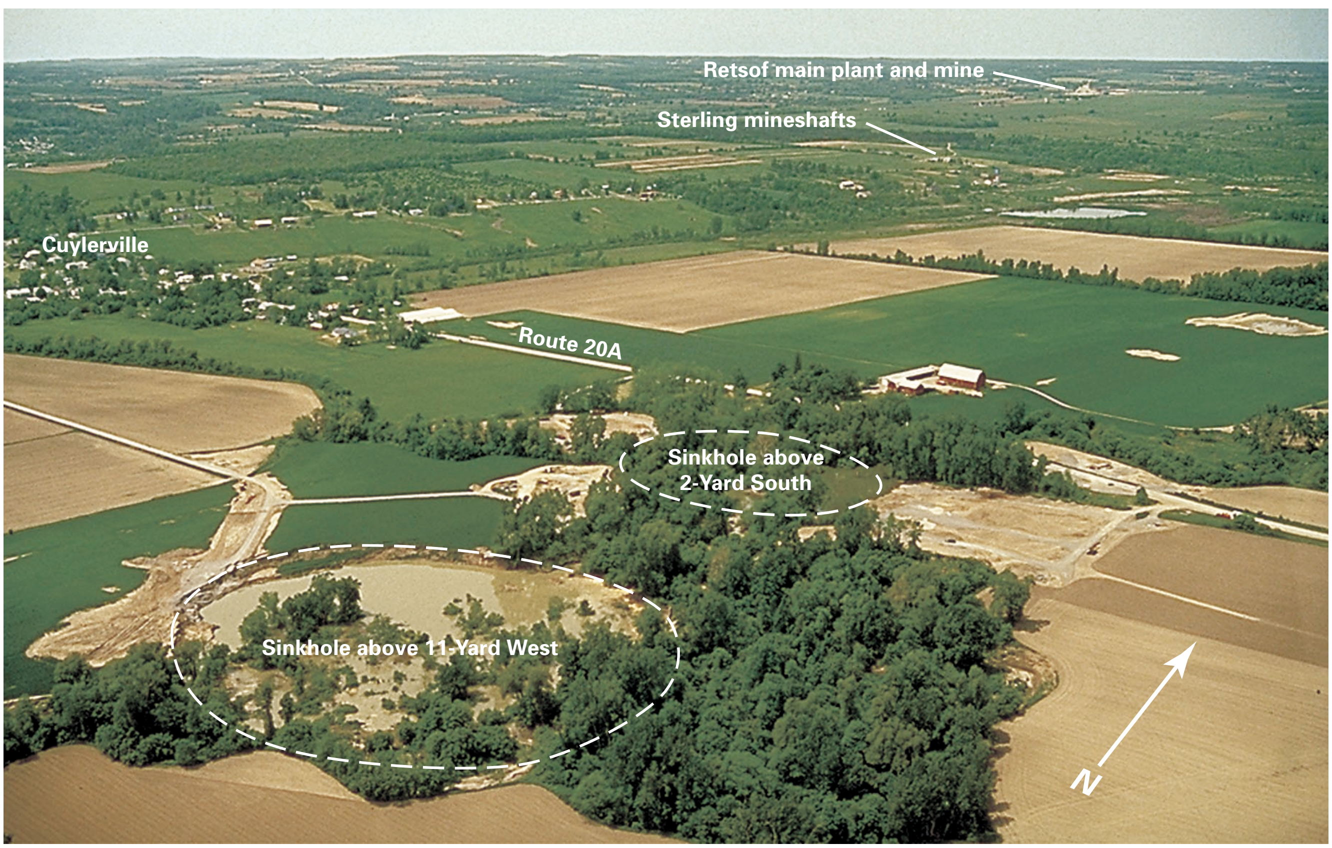
After the roof collapses, Nittany Geoscience (1995) summarized ground-water levels and chemical quality in the vicinity of the collapse in an effort to identify which wells were affected by drainage of ground water to the mine. Ground-water levels measured by the mine owner and Livingston County in the monitoring network from March 1994 through May 1996 are reported in Akzo Nobel Salt Salt, Inc. (ANSI, 1996), and land-surface-subsidence measurements from August 1994 to September 1996 are given in monthly reports by the mine owner (Kenneth Cox, Akzo Nobel Salt Salt, Inc. (ANSI), written commun., 1996). Stratigraphic logs from test drilling at five locations in the Genesee Valley, and results of hydraulic tests performed in the field and in the laboratory, are included in Alpha Geoscience (1995a,b,c and 1996a,b). The hydrogeology of the Genesee Valley, as indicated by test drilling and by records of gas wells and drilling logs prepared in previous studies, is summarized in Alpha Geoscience (1996c).

Methods of Investigation

Data from (1) seismic-reflection and seismic-refraction surveys, (2) water-level measurements, and (3) analyses of ground-water quality in this study were used with hydrogeologic information from previous studies to develop a conceptual model of the aquifer system and to design a three-dimensional ground-water-flow model.

Seismic Surveys

Seismic-reflection profiling was done by boat in the Genesee River to define the stratigraphy of unconsolidated deposits within the Genesee Valley. The survey began downstream from a small hydroelectric dam at the west side of the valley, where the Genesee River exits a bedrock gorge (Letchworth State Park), 1 mi



Courtesy of Ron Pretzer/LUXE, May 1994

Figure 3. View from southeast of sinkholes that formed in the collapse area, Retsof salt mine, Genesee Valley, Livingston County, N.Y., 1994. (Location is shown in fig. 2.)

north of the Village of Mt. Morris (fig. 2), and proceeded eastward across the valley. A description of the equipment, field technique, and theory is given by Reynolds and Williams (1987), and the application of the method to the delineation of hydrogeologic boundaries is described by Haeni (1986).

A seismic-refraction survey was conducted in the northern part of the study area to supplement bedrock-elevation data from well records. A continuous record of depth to bedrock was obtained along the survey line. Seismic-refraction techniques used in this study are described by Haeni (1988). A two-layer analysis of the saturated unconsolidated sediments and the bedrock (Scott and others, 1972) was used to calculate depths to bedrock.

Water-Level and Streamflow Measurements

Water levels were measured in more than 100 wells bimonthly or monthly (appendix 1 and fig. 4) from April 1994 to the present (1998) with electric tapes or chalked steel tapes. Water levels in eight wells were measured continuously with float-driven graphical recorders or pressure-transducers with data loggers. The altitude of measuring points was measured through transit surveys on some wells; the rest were estimated from 1:24,000-scale topographic maps. Wells listed in appendix 1 are included in the U. S. Geological Survey's Ground-Water Site Inventory data base; aquifers screened by wells for which no stratigraphic logs were available were assigned on the basis of well depth, estimated depth to bedrock and ground-water level. Streamflow measurements were made in Beards Creek and the Genesee River by standard USGS techniques (Rantz and others, 1982).

Ground-Water Quality

Ground-water samples were collected by Alpha Geoscience (1996c) from the 14 wells installed in 1994. A centrifugal pump was used to purge the wells and collect the samples. pH, specific conductance, and alkalinity were measured in the field; analyses for principal inorganic ions were done by Lozier Laboratories, Inc., of Rochester, N.Y. through standard techniques. Splits of the same samples were analyzed by the USGS for hydrogen-isotope (δD) values with a hydrogen-equilibration technique at 30°C (Coplen and others, 1991), and for oxygen-isotope ($\delta^{18}O$) values by the CO_2-equilibration technique of Epstein and Mayeda (1953) at 25°C, followed by mass spectrometry. Stable-isotope ratios of deuterium to hydrogen and of ^{18}O to ^{16}O are reported in per mil relative to Vienna Standard Mean Ocean Water (V-SMOW) and normalized such that the δD and $\delta^{18}O$ values of Standard Light Antarctic Precipitation (SLAP) are -428‰ and -55.5‰, respectively (Coplen, 1994). The uncertainty of δD and $\delta^{18}O$ values, expressed as twice the standard deviation of the measurements, is 2‰ and 0.2‰, respectively.

Acknowledgments

Ralph VanHouten of the Livingston County Department of Health provided water-level and water-use data. ANSI staff provided water-level and geologic data, and access to the collapse area and to the rest of the mine. Dr. Richard Young and Dr. William Brennan of the Geology Department at State University of New York (SUNY) at Geneseo provided geologic data and advice. Dr. John Fountain of the Geology Department at SUNY at Buffalo helped derive the analytical expression used to describe the effect of gas exsolution on specific storage of the aquifer material. Many home-owners provided access to their water wells for measurement of water levels and collection of water samples.

GEOLOGIC SETTING

The Genesee Valley was formed through several geologic processes, including: (1) tectonic uplift of Paleozoic sedimentary rocks and subsequent fluvial downcutting, (2) glaciation that resulted in erosion of bedrock and subsequent deposition as much as 750 ft of glacial sediments, and (3) erosion and deposition by postglacial streams. Structural control within bedrock may account for the present alignment of the valley, which bends 60° from N30°W in the south to N30°E in the north (see fig. 2) (Isachsen and McKendree, 1977).

The stratigraphy and composition of the glacial deposits are more complex than described in this report. The data available are insufficient to describe the geologic framework of the Genesee Valley in detail, but the ground-water flow model (described further on) was based on the geologic framework as described herein, and seems adequate to simulate the generalized ground-water flow conditions in the valley.

Bedrock Units

The mined salt bed and overlying rocks are sedimentary units of Silurian and Devonian age. The age, formation names, lithologic description and average thickness of bedrock units in the vicinity of the mine are summarized in figure 5. Bedrock units affected by the ceiling collapses include the mined B6 salt unit within the Vernon Shale and the following overlying units, in ascending order: the Syracuse Formation, Camillus Shale, Bertie Limestone, and Onondaga Limestone (fig. 6). The younger Devonian units (Marcellus Shale, Skaneateles Shale, Ludlowville Shale, and Moscow Shale), which form the valley walls and uplands (fig. 6), were probably not greatly affected by the collapses. All bedrock units dip less than 1° to the south. A regional unconformity separates the late Silurian Bertie Limestone from the Middle Devonian Onondaga Limestone (figs. 5, 6). The Onondaga Limestone subcrop forms most of the buried bedrock surface in the northern and central part of the Genesee Valley (figs. 4 and 7).

Silurian Units

The Vernon Shale is typically 420 ft thick in the Genesee Valley (Dunn Geoscience Corporation, 1992) and consists of calcareous shales interlayered with salt beds. The mine owner mined a salt bed within the Vernon Shale known as the B6 salt unit, which ranges from 13 to 30 ft in thickness (figs. 5 and 6). Salt unit B6 slopes southward from an altitude of -260 ft in the northern part of the mine to -540 ft in the southern part. This unit is about 1,100 ft below land surface in the collapse area at the south end of the mine.

The Syracuse Formation typically is 185 ft thick in the study area; it consists mainly of dolomitic shale and shale interbedded with some salt and anhydrite beds and rests conformably on the Vernon Shale. It was referred to as the "Syracuse Salt" by Newland and Leighton (1910). It is conformably underlain by the Camillus Shale, a mostly dolomitic and calcareous shale that typically is 70 ft thick in the study area (Dunn Geoscience Corporation, 1992). Gamma and drillers' logs indicate that the upper 15 to 20 ft of the Camillus shale contains some dolomite and evaporite beds, but the lower 50 ft is almost entirely shale.

The uppermost Silurian unit is the Bertie Limestone, which in the Genesee Valley is typically 72 ft thick and, despite its name, consists mainly of dolomitic shale and dolomite with some limestone.

Devonian Units

The Silurian Bertie Limestone is unconformably overlain by the Devonian Onondaga Limestone. This unit is mostly limestone, but some layers are as much as 50 percent insoluble, hard chert. The massive beds of chert-bearing limestone make this formation one of the most erosion-resistant sedimentary units in New York. The Onondaga Limestone is a major aquifer in parts of western and central New York where it crops out at land surface or underlies thin glacial drift (fig. 1). Large amounts of ground water can be pumped from the Onondaga wherever dissolution along vertical joints and bedding planes has resulted in major water-bearing zones (Staubitz and Miller, 1987). The Onondaga Limestone averages 137 ft thick beneath the uplands where it has not been eroded, but is thinner in the north part of the Genesee Valley, where the top was eroded by glaciers.

The Onondaga Limestone is overlain by about 750 ft of younger Devonian rock in the uplands and along the sides of the Genesee Valley; these units, in ascending order, are the Marcellus Shale, Skaneateles Shale, Centerfield Limestone, Ludlowville Formation, Tichenor Limestone, and Moscow Shale (fig. 6, section A-A′). The contact between the Moscow Shale and the Ludlowville Formation was formerly placed 50 ft or more above the Tichenor Limestone (Cooper, 1930; Bradley and Pepper, 1938; Rickard, 1975), but more recently has been placed at the base of the Tichenor Limestone (Dunn Geoscience Corporation, 1992; Mayer ,1994; Baird, 1979). All of these units have been removed by erosion along the valley axis near and north of the collapse area. The three shale units, while capable of yielding a few gallons per minute to wells in most places, are probably less permeable than the thin limestone units, which contain solution-widened joints and openings. Water levels in wells that tap these units were generally unaffected by the flooding of the mine. Few data on water levels in bedrock wells prior to the collapse are available, however.

Glacial and Postglacial Geology

The glacial history of the Genesee River Basin is described by Muller and others (1988); discussion in this report is limited to glacial sediments in part of the valley that was affected by the mine collapse. The interpretation of glacial geology in the Genesee Valley is based on recent subsurface information collected during this investigation and results of seismic-reflection

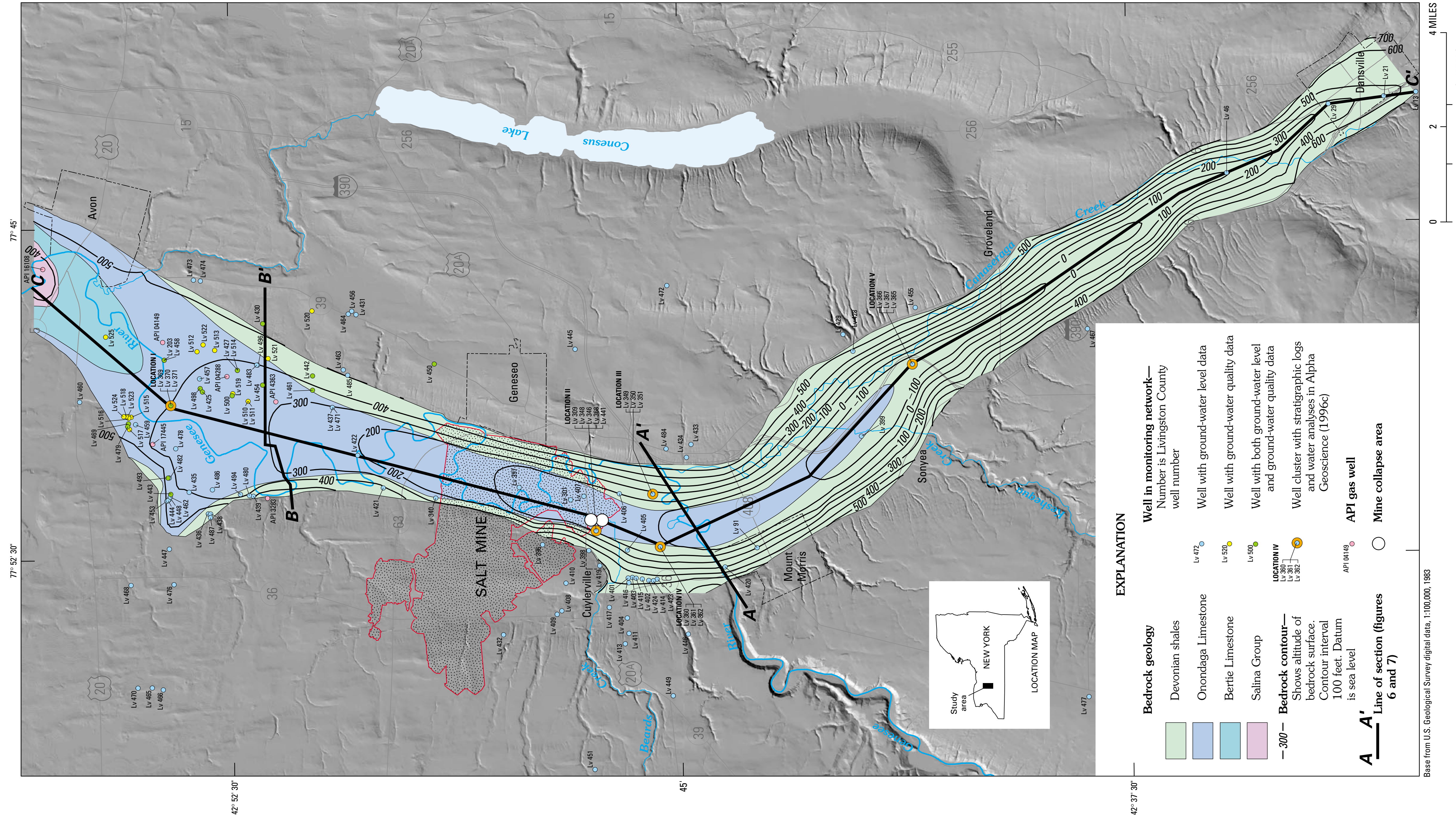


Figure 4. Bedrock geology and bedrock surface altitudes in modeled Genesee Valley aquifer area, Livingston County, N.Y., and locations of wells in ground-water-monitoring network.

FOLD FOLD

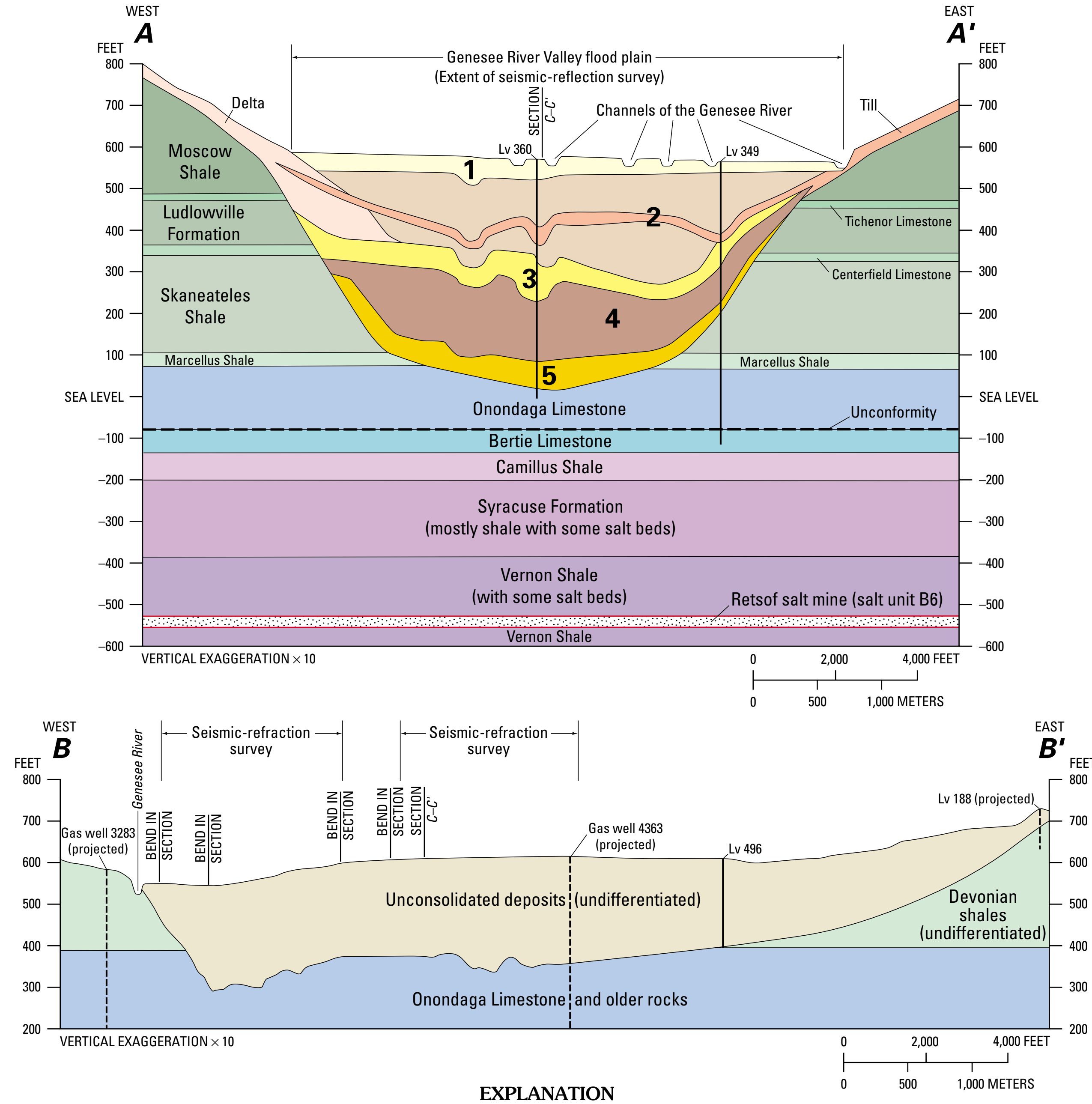
studies conducted by Mullins and others (1991) in parallel Finger Lake valleys to the east that were similarly deepened by glacial scour.

Glaciation altered stream courses in western New York, removed preglacial sediments, deepened valleys, and deposited large amounts of sediment. Glaciers typically transport large amounts of sediment (1) at the base of the ice as subglacial material, (2) within the ice as englacial material, and (3) on top of the ice as supraglacial material. Test drilling revealed as much as

SYSTEM	GROUP	FORMATION	LITHOLOGY	AVERAGE THICKNESS, IN FEET	DESCRIPTION
DEVONIAN	Hamilton	Moscow Shale		120	Shale and siltstone, gray and black; Tichenor Limestone (8 feet thick) at base of unit
		Ludlowville Formation		118	Shale and siltstone, gray, calcareous; Centerfield Limestone (8 feet thick) at base of unit
		Skaneateles Shale		237	Shale and siltstone, gray, brown, and black, calcareous; commonly contains thin limestone beds
		Marcellus Shale		27	Shale and limestone, gray
		Onondaga Limestone		137	Limestone, gray, massive bedding; contains abundant chert
				Unconformity	
SILURIAN	Salina	Bertie Limestone		72	Dolomite and dolomitic shale, gray to grayish brown
		Camillus Shale		70	Shale, gray and green, dolomitic; contains anhydrite concretions
		Syracuse Formation		185	Shale, salt, gypsum, and anhydrite gray shale, white and pink to yellowish-brown salt, gypsum, and anhydrite
		Vernon Shale		420	Shale, interbedded with salt, brown, green and gray shale; clear white, gray, and orange salt Retsof mine in B6 salt unit (13-30 feet thick)

Geology modified from Dunn Geoscience Corporation , 1992

Figure 5. Generalized columnar geologic section of Paleozoic rocks in Retsof salt mine vicinity, Livingston County, N.Y.



EXPLANATION

Unconsolidated deposits—Numbers are geohydrologic units referenced in text (Geologic Setting)

- **1** Alluvial-fan and flood-plain deposits—Silt, sand, and gravel
- **2** Lacustrine sand, silt, and clay
- **2** Till—Silt and clay embedded with pebbles
- Deltaic deposits—Pebbly sand
- **3** Supraglacial debris—Silty sand and gravel
- **4** Englacial meltout material and subaqueous-fan deposits—Mix of clay, silt, and sand and gravel
- **5** Subglacial meltwater deposits—Sand and gravel

Lv349 **Well and U.S. Geological Survey well number—** Dashed where projected. Numbers with Lv prefix were assigned by USGS. Well data are given in appendix 1. Numbers for gas wells were assigned by New York State Department of Environmental Conservation

Figure 6. Generalized geologic sections A–A′ across the valley near collapse area, and B-B′ 6 miles north of collapse area at Retsof salt mine, Livingston County, N.Y. (Lines of sections are shown in fig. 4.)

750 ft of gravel, sand, silt, and clay in the Genesee Valley (Alpha Geoscience, 1996c). These sediments were deposited in (1) glaciofluvial environments (where meltwaters discharged at the base of the ice when ice occupied the valley), (2) glaciolacustrine environments (where proglacial lakes occupied the valley in front of the retreating ice), and (3) flood-plain environments (along the riverbank during recent postglacial time).

Erosion of Bedrock by Ice

Glacial erosion deepened the preglacial Genesee Valley by as much as 600 ft and widened it, creating a concave, north-south bedrock profile (fig. 7). The eroded bedrock surface (Onondaga Limestone) slopes southward from Fowlerville at the north end of the valley where it subcrops at an altitude of 400 ft. The deepest part of the bedrock basin (177 ft below sea level) is estimated to be in the central part of the valley, near Sonyea, 9.5 mi south of Fowlerville. The bedrock surface rises from Sonyea southward and reaches an altitude of about 580 ft at Dansville.

The erosion-resistant Onondaga Limestone limited the depth to which ice could scour, and thus, controlled to a large extent the resulting bedrock-surface configuration in the valley. Ice in the northern part of the valley, where depth to the Onondaga Limestone was relatively shallow (150-500 ft below land surface), eroded the soft overlying shales but could not erode through the entire limestone unit; therefore it slid southward atop the limestone to the point at which this unit was too deep to reach (750 ft below land surface near Sonyea) or to where it encountered harder bedrock units to the south, where the Appalachian Plateau begins. The ice was deflected upward through a bedrock saddle about 5 miles south of Dansville.

Much of the published literature describes glaciated valleys in central and western New York as being U-shaped. Seismic-reflection work by Mullins and others (1991) in the Finger Lakes of New York indicates, however, that the major north-south valleys change from broad and relatively shallow in the north to U-shaped in the middle, and are narrower and V-shaped in the south. The mine collapse area is in the U-shaped middle reach of the Genesee Valley (fig 6, section A–A′). A seismic-refraction survey conducted 6 mi north of the collapse area indicated that the bedrock valley floor is broad and relatively shallow (fig. 6, section B–B′), a finding consistent with that of Mullins and others (1991). South of Sonyea, the buried bedrock valley appears to become V-shaped, but subsurface data are insufficient to verify this conclusion.

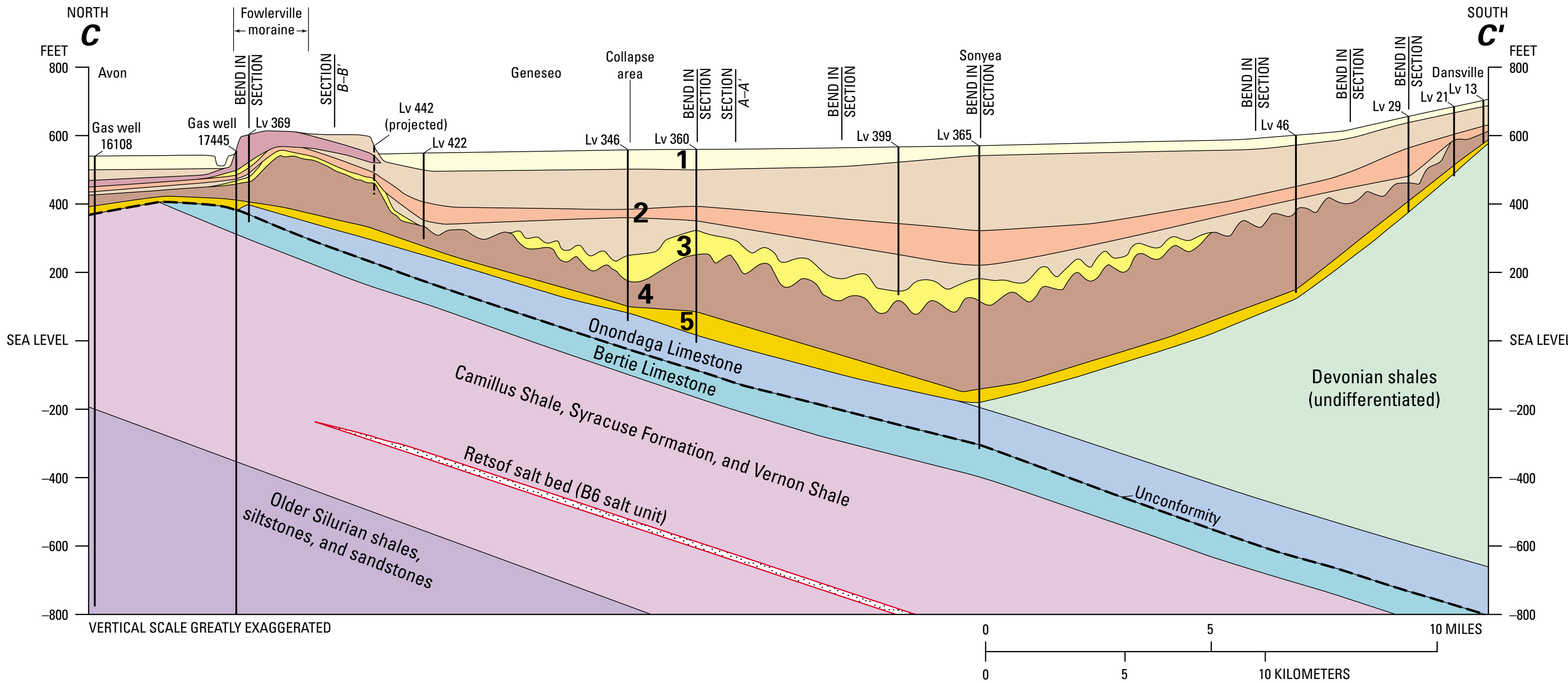
Glaciation probably removed much of the weathered bedrock that had presumably formed at the bedrock surface before the the Wisconsinan glaciation. Subsequent weathering of the bedrock surface (after glaciation) was probably limited because the bedrock was protected by thick overlying deposits of glacial sediments.

Unconsolidated Deposits

The unconsolidated sediments that partly fill the Genesee Valley were deposited during glacial advances and retreats and, to a lesser extent, during postglacial time. The distribution of surficial deposits in the Genesee Valley from Avon to Dansville is generalized in figure 8. Although the valley was covered by continental glaciers several times during Pleistocene time, each glaciation removed most evidence of prior glaciations. Most of the glacial sediments in the region are probably from the late Wisconsinan stage (Muller and others, 1988), but remnants of older glacial deposits are found north of the study area (Young and Sirkin, 1994).

Subglaciofluvial Sand and Gravel

A layer of coarse sand and gravel overlies the bedrock valley floor, as indicated by drillers' stratigraphic logs, gamma logs from wells drilled by the mine owner and by water-well drillers in the Genesee Valley. This basal layer is 10 to 40 ft thick in the collapse area and 460 to 500 ft below land surface (fig 6). Drillers reported that the clasts were predominantly limestone, presumably derived from the underlying Onondaga Limestone. All deep wells in the valley that extended to, or close to, bedrock penetrated a basal coarse-grained deposit, but the permeability of this unit varies from place to place. This unit was probably deposited by subglacial meltwaters (fig. 9*B*, unit 5) that formed eskers and subglacial outwash fans at the base of the ice. Eskers were found in several Finger Lake valleys by Mullins and others (1991), who hypothesize that major north-south valleys between Lake Ontario and the Allegheny Plateau were principal routes of sediment transport by subglacial meltwater. Some of the subglacial sediment that was transported southward through the Genesee Valley near the base of the ice was deposited south of Dansville where it forms the Valley Heads Moraine, which plugs the valley (fig. 8).



EXPLANATION

Unconsolidated deposits—Numbers are geohydrologic units referenced in text (Geologic Setting)

1 — Alluvial-fan and flood-plain deposits—Silt, sand, and gravel

2 — Lacustrine sand, silt, and clay

2 — Till—Silt and clay embedded with pebbles

Deltaic deposits—Pebbly sand

3 — Supraglacial debris—Silty sand and gravel

4 — Englacial meltout material and subaqueous-fan deposits—Mix of clay, silt, and sand and gravel

5 — Subglacial meltwater deposits—Sand and gravel

Lv349 — **Well and U.S. Geological Survey well number—** Dashed where projected. Numbers with Lv prefix were assigned by USGS. Well data are given in appendix 1. Numbers for gas wells were assigned by New York State Department of Environmental Conservation

Figure 7. Generalized geologic section C–C′ along Genesee valley axis, Livingston County, N.Y. (Line of section is shown in fig. 4.)

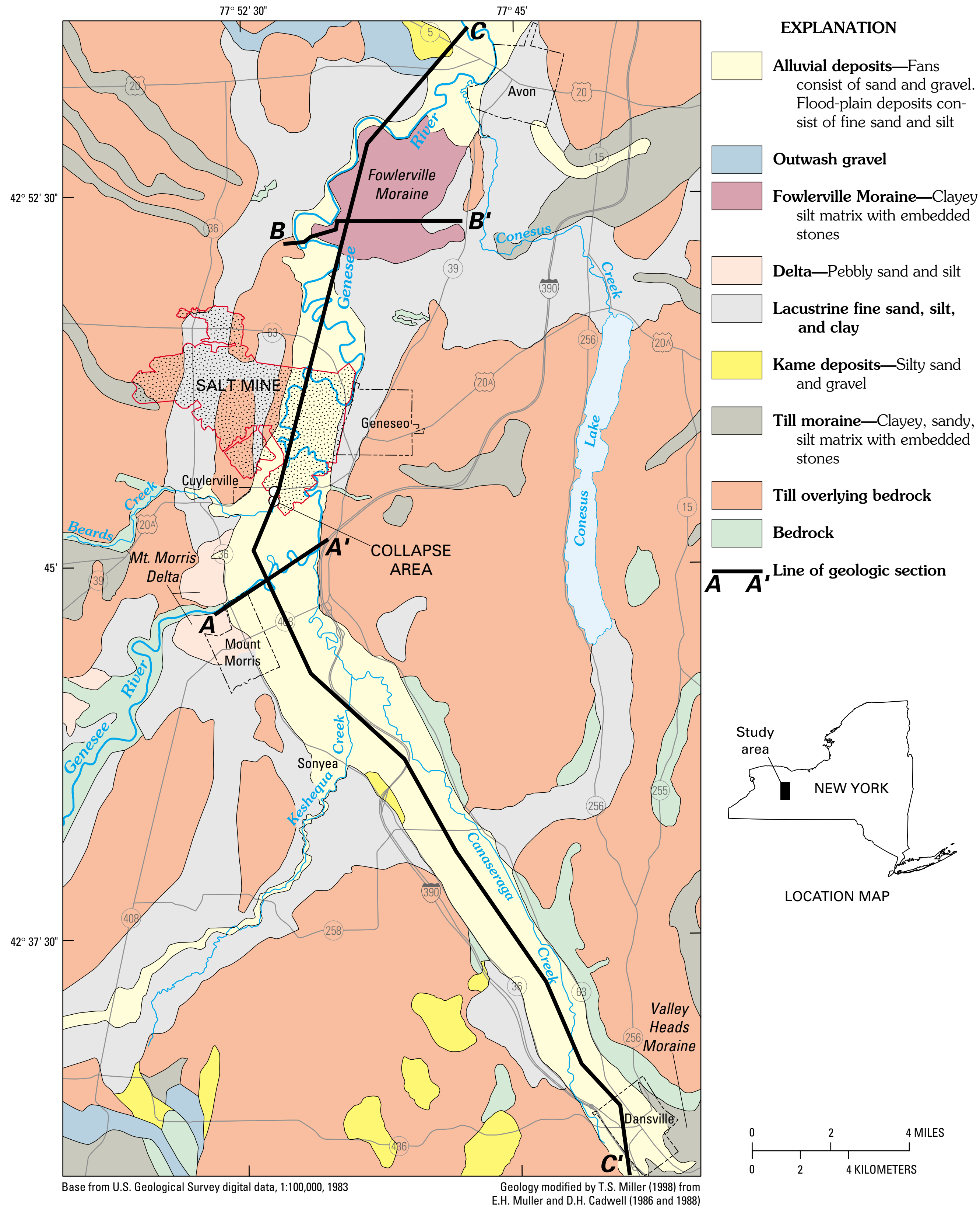
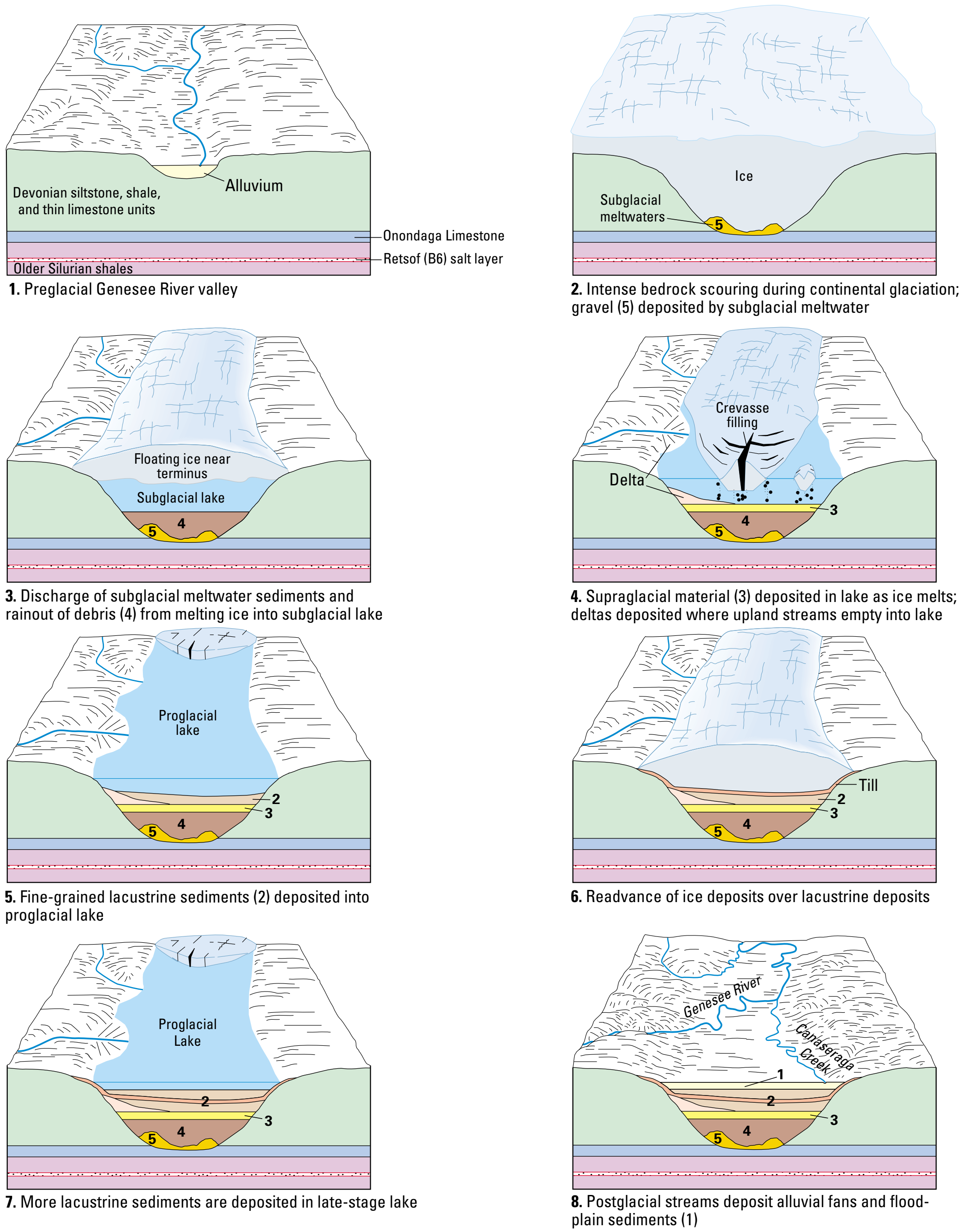


Figure 8. Surficial geology of Genesee Valley from Avon to Dansville, N.Y. (Location is shown in fig. 1.)



EXPLANATION

Unconsolidated deposits—Numbers are geohydrologic units referenced in text (Geologic Setting)

- **1** Alluvial-fan and flood-plain deposits—Silt, sand, and gravel
- **2** Lacustrine sand, silt, and clay
- **2** Till—Silt and clay embedded with pebbles
- Deltaic deposits—Pebbly sand
- **3** Supraglacial debris—Silty sand and gravel
- **4** Englacial meltout material and subaqueous-fan deposits—Mix of clay, silt, and sand and gravel
- **5** Subglacial meltwater deposits—Sand and gravel

Figure 9. Glacial-deposition processes in Genesee Valley, Livingston County, N.Y.

Englacial and Glaciolacustrine Sediments

As the ice retreated northward from the Valley Heads moraine near Dansville, it formed a temporary dam that impounded drainage to form a proglacial lake that declined through several stages in the valley between the retreating ice and the moraine. The proglacial lakes were sufficiently deep to allow the thinning ice front to float. As much as 250 ft of sediment was deposited in the proglacial lakes from several sources, which included (1) subaqueous fans deposited by subglacial meltwater at the submerged part of the ice front, and (2) englacial sediment (referred to as flowtill and rainout material) that dropped from floating ice to form a poorly sorted deposit of variable thickness (fig. 9*C*, unit 4). Core samples collected by Alpha Geoscience (1995a,b,c and 1996a,b) indicate that these sediments consist of interbedded coarse sand and gravel, silty gravel, till, and lacustrine fine sand, silt and clay. The precise geometry of all these sedimentary units is more complex than shown in figures 5 and 6.

Supraglacial Sand and Gravel

Water-well drillers' logs and results of a 1994 marine-seismic-reflection survey (fig. 6) conducted by the USGS in the Genesee River northeast of Mt. Morris indicate a relatively thin layer of sand and gravel (typically less than 10-ft-thick but as much as 60-ft-thick) that overlies the englacial and glaciolacustrine sediments in many places. The extent of this deposit is uncertain, but several domestic wells in the Fowlerville Moraine area tapped similar sand and gravel zones before they went dry after the collapse. These wells were subsequently deepened to tap the lower aquifer. The seismic-reflection record south of the collapse contains strong reflectors that clearly delineate an undulating surface across most of the valley; this surface is assumed to correspond to the sand-and-gravel layer that lies 250 to 310 ft below land surface near the collapse area. This layer probably consists of (1) supraglacial sediments deposited by meltwater, (2) upland inwash that settled in the proglacial lake during the recession of the ice margin, and (3) material derived from coarse-grained subaqueous fans (fig. 9*D*, unit 3).

Glaciolacustrine and Deltaic Sediments and Till

A series of proglacial lakes formed in the Genesee Valley in front (south) of the receding glacier margin. The lake levels were controlled by natural spillways at topographic lows on the watershed perimeter. A detailed account of the various lake levels and spillways is given in Muller and others (1988). The oldest and highest spillway was across the Valley Heads Moraine, south of Dansville (fig. 8); several successively lower spillways to the north, between 1,400 and 900 feet, were exposed as the ice margin retreated northward. Lake levels probably dropped abruptly as each successive spillway was exposed. As much as 250 ft of fine-grained sediment from the receding ice margin and upland streams (fig. 9*E*, unit 2) was deposited above the supraglacial sediments in the middle and southern parts of the valley, and deltas were deposited along the edges of the valley where upland streams flowed into the lakes. The largest delta in the valley (locally known as the Mt. Morris Delta) was deposited near Mt. Morris (fig. 8), where the Genesee River flowed from the Letchworth Gorge into the Genesee-Canaserga valley. The Mt. Morris Delta consists mostly of silt, sand, and pebbly sand. Other deltas are either buried by flood-plain sediment or are small and above the present water table and, therefore, are not shown in figure 8.

Drillers' logs and seismic-reflection surveys indicate an apparent till layer within the glaciolacustrine sediments (fig. 6*A*). The presence of till suggests that the ice readvanced southward into the lake after it had retreated to the northern part of the valley, and overrode the glaciolacustrine sediments (fig. 9*F*). The till layer consists of compacted and reworked lacustrine sediments with some stones embedded in a fine-grained matrix. The hydraulic properties of the till are similar to those of the lacustrine sediments, and the till is considered part of the lacustrine deposit (unit 2) in this report. The subsequent northward retreat of the ice left the till submerged in the proglacial lake, and additional lacustrine sediments were deposited on top of the till. Several drill logs indicate a 5- to 15-ft-thick sand-and-gravel layer beneath the till in some locations. These small, discontinuous sand-and-gravel layers could have been deposited by subglacial meltwater at the readvancing ice margin before deposition of the till.

The Fowlerville Moraine (figs. 7 and 8) was deposited during the final and lowest proglacial lake stage and fills the valley in the northern part of the study area. A lake formed in the 27-mi reach between this moraine and the Valley Heads Moraine near Dansville to the south (Young, 1975), and glaciolacustrine sediments were deposited over the till layer described above (fig. 9*G*). The glaciolacustrine sediments above the till are finer grained and better sorted than those below the till and consist mostly of rhythmically bedded silt and clay, whereas the glaciolacustrine sediments below the till consist of fine sand, silt, and clay, and contain some ice-rafted sand and gravel and dropstones that probably dropped into the lake when the ice melted. The glaciolacustrine and till deposits are 200 to 250 ft thick and extend to depths of 250 to 310 ft below land surface. The most recent proglacial lake drained as the outlet stream gradually eroded a spillway through the Fowlerville Moraine.

Alluvial Sediments

The upper 20 to 60 ft of the valley-fill sediments are Holocene deposits (fig. 9*H*, unit 1) that consist of mixed coarse- and fine-grained sediments deposited by the Genesee River, Canaseraga Creek, and small upland tributaries that flow into the valley. The coarse fraction of the alluvial sediment consists of sand and gravel deposited as fans and within stream channels; the fine-grained fraction includes fine sand, silt, and clay laid down as point bar and overbank deposits during floods, which were frequent in the Genesee Valley until the Mt. Morris Dam was completed in 1954.

HYDROLOGIC SETTING

The shallow unconsolidated deposits in the Genesee Valley are recharged by precipitation on the valley floor and by runoff from upland areas. Prior to the mine collapse, ground water discharged to streams or flowed northward. The salinity of ground water generally increases with depth in the valley, and the low tritium concentrations indicate that most of the water in the two confined aquifers is more than 45 years old.

Aquifers and Confining Units

The glacial aquifer system in the Genesee Valley is underlain and bounded laterally by bedrock and consists of three aquifers separated by two confining layers. The degree of hydraulic connection among these three aquifers in areas along the edges of the valley, where the confining layers are thin or absent, is unknown because the stratigraphy of glacial sediments in the Genesee Valley is complex. (The general conceptualization of aquifer geometry presented in this section may not conform to actual aquifer conditions at all locations.) The underlying carbonate bedrock contains water-bearing zones that are bounded on the north by the Onondaga escarpment, where the outcrop ends (fig. 1), and extend southward (downdip) beyond the end of the valley, as well as eastward and westward (cross-strike).

Upper Aquifer

The unconfined (upper) aquifer consists of sand-and-gravel layers within the 20-to-60-ft thick Holocene alluvial sediments at land surface (fig. 10). The bottom of the upper aquifer conforms to the nearly flat-lying lacustrine sediments (upper confining layer) below. Several domestic wells are screened in the upper aquifer.

Upper Confining Layer and Deltaic Deposits

The upper confining layer separates the upper aquifer from the middle aquifer (fig. 10) and consists of glaciolacustrine sediment and till. It is about 250 ft thick along the axis of the valley and could be locally absent or coarse-grained near the valley walls. The hydraulic conductivity of the lacustrine sediment, calculated from the geometric mean of 11 permeameter test samples, is about 9×10^{-5} ft/d (Alpha Geoscience, 1996c).

The confining layer is assumed to be bounded laterally by permeable deposits of coarse materials that in some places mantle the sides of the bedrock valley, and provide vertical hydraulic connections between the upper aquifer and the middle aquifer. For example, a wedge-shaped deltaic deposit of mostly sand mantles the west valley wall near Mt. Morris and slopes eastward to the center of the valley (figs. 6, 8 and 10). Several domestic and commercial wells along the valley wall are screened in presumed deltaic or ice-marginal deposits and have yields ranging from 20 to 400 gal/min. A kame deposit mapped at land surface near Sonyea (fig. 8) also might extend to the top of the middle aquifer, but no stratigraphic logs are available to support this hypothesis.

Middle Aquifer

The middle aquifer underlies the upper confining layer and consists of sand and gravel, typically less than 10 ft thick, that was deposited in a proglacial lake. The upper surface is undulating as a result of the varied depositional processes that formed this aquifer. Several domestic wells are screened in this aquifer beneath the Fowlerville Moraine; a well screened in this aquifer near Sonyea (Lv366), where it is 60 ft thick and has a

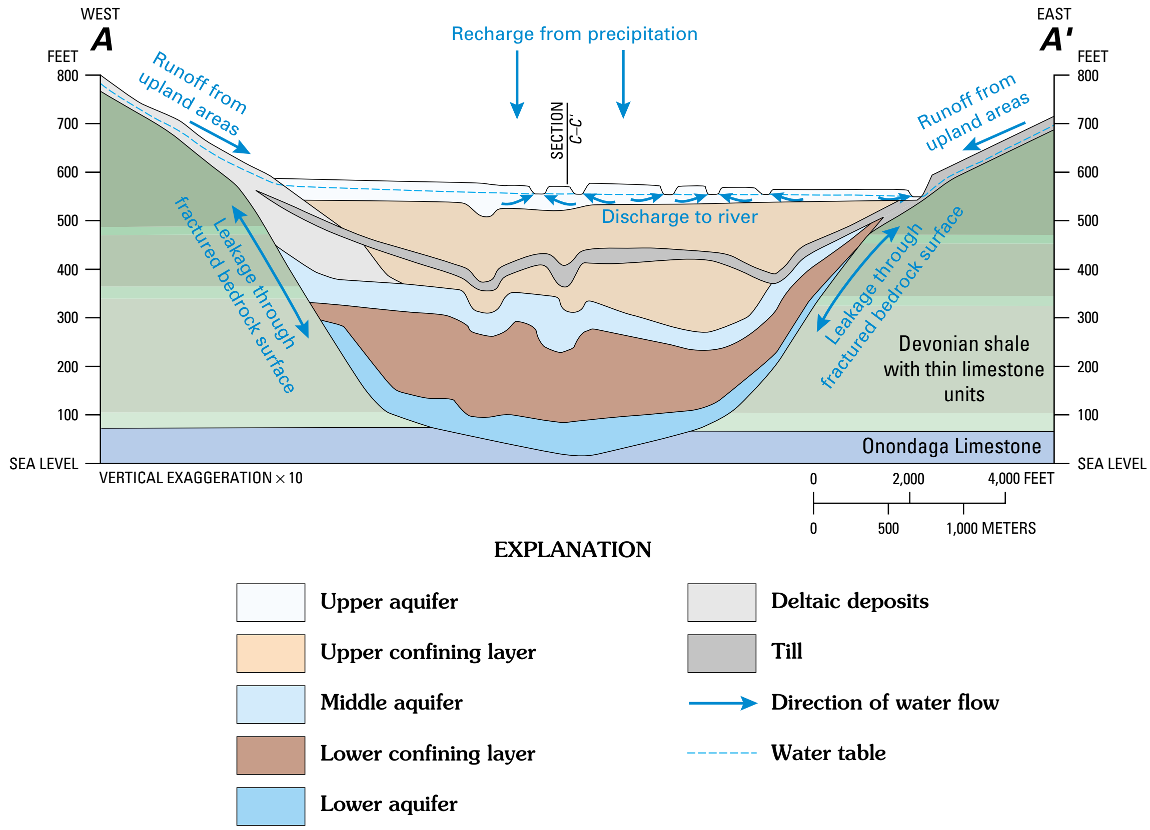


Figure 10. Aquifers and confining layers along geologic section A–A′ in Genesee Valley, Livingston County, N.Y. (Line of section is shown in fig. 8.)

hydraulic conductivity of 80 ft/d, yielded 300 gal/min (Alpha Geoscience, 1996c).

Lower Confining Layer

A lower confining layer of fine-grained sediments as much as 250 ft thick separates the middle aquifer from the lower aquifer (fig. 10). The lower confining layer consists of glaciolacustrine sediment and till with lenses of coarse materials that were deposited in a proglacial lake from the floating ice margin. The hydraulic conductivity of fine-grained sediments in the lower confining layer, as calculated from the geometric mean of four permeameter test samples, was 3×10^{-4} ft/d (Alpha Geoscience, 1996c). Kame deposits, if present, would provide hydraulic connections between the middle and lower aquifers, but no stratigraphic logs are available to confirm their presence.

Lower Aquifer

The lower aquifer overlies the bedrock valley floor and consists of glaciofluvial sand and gravel about 25 ft thick that was deposited in tunnels beneath the ice and on fans where subglacial streams flowed into the proglacial lake at the ice front. Two industrial wells and several domestic wells are screened in the lower aquifer; their yields range from 20 to 500 gal/min. Industrial well Lv91 in Mt. Morris (fig. 11) yields 500 gal/min resulting in a drawdown of about 15 ft. Aquifer transmissivity at this location is about 10,000 ft^2/d as estimated from the specific capacity of 33 (gal/min)/ft, calculated from a relation given in Todd (1980, eq. 4.70). This transmissivity value corresponds to a hydraulic conductivity of 500 ft/d. Well Lv46, which was drilled 2 mi north of Dansville (fig. 11), yielded about 3,000 gal/min from this aquifer under flowing artesian conditions.

Fractures at the bedrock surface where the aquifer overlies the Onondaga Limestone increase the effective thickness of the lower aquifer and increase ground-water flow through it. Well Lv346, 900 ft west of the collapse, is finished from 7 to 17 ft below the Onondaga surface and yielded 100 gal/min during a pumping test in November 1994, from which the hydraulic conductivity was estimated to be 75 ft/d (Alpha Geoscience, 1996c). If the Onondaga Limestone at this location is highly fractured from the collapse, this value is higher than in in other areas. Fractures in the shales that form the valley walls could provide hydraulic connections between the lower aquifer and bedrock aquifers in uplands that border the Genesee Valley (fig. 10).

Bedrock Aquifers

Several water-bearing zones in the bedrock underlying the Genesee Valley were identified through borehole geophysical surveys (Williams, 1996). Borehole geophysical logs indicate that most of these zones are nearly horizontal bedding planes at stratigraphic contacts. The principal water-bearing zone that subcrops within the Genesee Valley is a fracture zone near the contact between the Onondaga and Bertie Limestones, previously identified as the source of leakage into a shaft drilled to the Sterling salt mine (Langill and Associates, 1981).

Few wells are finished in bedrock beneath the Genesee Valley because the water quality is generally poor, except in subcrop areas near the Onondaga Escarpment, where recent fresh recharge enters the bedrock. Well yields outside the Genesee Valley range from 10 to 160 gal/min in subcrop areas but are generally less than 10 gal/min in areas south of the subcrop, where the weight of overlying sediments compresses the fracture apertures. Leakage from the Onondaga/Bertie contact to the Sterling shaft has increased steadily since the 1930’s and was estimated to be 300 gal/min in 1986. This increase indicates that fracture apertures near the shaft have probably been widened by dissolution and are, therefore, not representative of fractures in other areas (Acres International, 1986).

Ground-Water Flow Patterns Before the Collapse

Ground water within the valley originates as precipitation on the valley floor and the surrounding uplands, and discharges to the Genesee River and Canaseraga Creek and their tributaries, or as underflow to downgradient areas. It generally flowed northward through the aquifer system prior to the mine collapse.

Upper Aquifer

Most of the ground-water levels recorded prior to the roof collapse in the Retsof mine were measured in wells screened in the upper aquifer (Kammerer and Hobba, 1967). These measurements, together with mapped elevations of springs and perennial streams, were used to construct a map showing the altitude of the water table prior to the collapse (fig. 11). This map is assumed to be representative of undisturbed, steady-state conditions with no change in ground-water storage, because no large withdrawals that could cause water-level fluctuations were being made prior to the collapse.

Water in the upper aquifer generally flows northward and to major streams in the valley. Recharge is derived from precipitation over the valley floor and runoff from the surrounding uplands (fig. 10), it also enters as lateral underflow from the surrounding bedrock uplands. The rate of recharge is largest in the north, where the aquifer is bounded by Onondaga Limestone, because fracture zones in the limestone probably transmit more water than fracture zones in the shales that border the rest of the valley. Recharge also enters the upper aquifer as upland runoff (Morrissey and others, 1988), which is potentially significant everywhere in the valley. Most of the water in the upper aquifer discharges to the Genesee River and Canaseraga Creek, although some discharges northward as underflow.

Middle and Lower Aquifers

Recharge enters the two confined aquifers through the Valley Heads moraine south of Dansville and along the sides of the valley in some areas (fig. 10). The rate of recharge along the sides of the valley is probably largest where locally permeable deposits, such as the Mt. Morris delta, provide a hydraulic connection with the upper aquifer. Water flowing through the confined aquifer in the Beards Creek Valley, west of Mt. Morris, also discharges to the middle aquifer. Ground water discharges from the confined aquifers primarily as underflow to the area north of the study area and possibly to the upper aquifer along the edges of the valley in some areas.

The hydraulic-head distribution in the confined aquifers under precollapse conditions was probably similar to that in the upper aquifer, but the heads in the confined aquifers beneath the valley floor were probably higher than the water table. This interpretation is supported by drillers’ logs that note confined conditions at several locations in the lower aquifer (see fig. 11), including well Lv46 near Dansville , well Lv421 at an industrial facility near Cuylerville (H&A of New York Consulting Engineers, 1988), and well

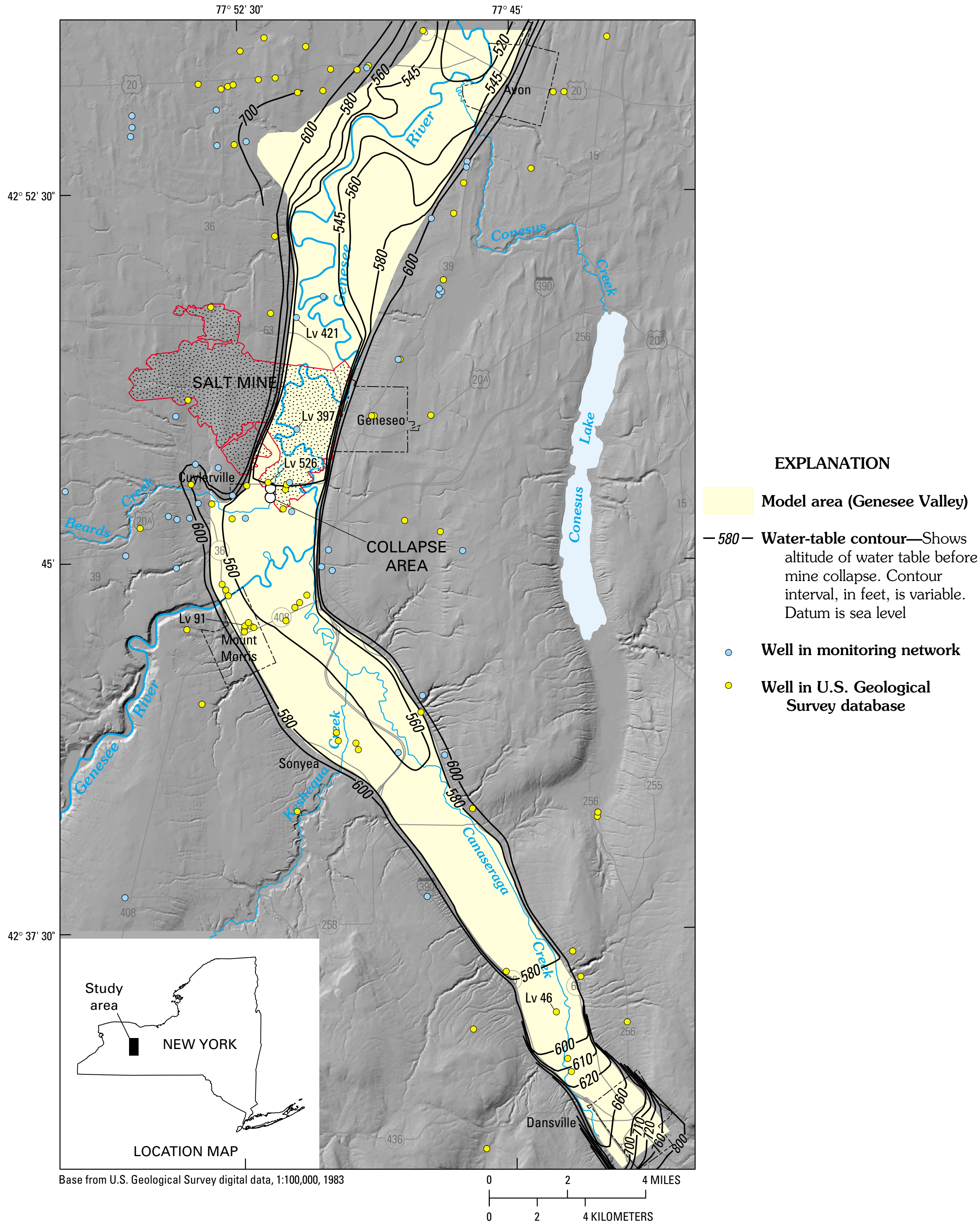


Figure 11. Water-table altitude in upper aquifer prior to mine collapse in March 1994, and locations of wells for which water-level meaurements are available. (Location is shown in fig. 1.)

Lv397 near the Geneseo airport (Dunn Geoscience Corporation, 1992). Heads at other wells screened in the lower aquifer, including well Lv526 near the collapse area and well Lv91 in Mt. Morris, were within a few feet of land surface before the collapse.

Bedrock Aquifers

Little information is available on ground-water flow in the bedrock underlying the valley. La Sala (1968) suggested that circulation of deep ground water flowing northward from the Appalachian uplands to the Erie-Ontario lowlands brought highly mineralized water into shallow aquifers in Erie County, west of Livingston County (fig. 1). The similarity of the bedrock stratigraphy and topographic setting in Livingston County to that in Erie County suggests northward flow in this area as well, but the rate of flow is unknown. Water in fracture zones near the Onondaga/Bertie Limestone contact probably discharges upward into the lower aquifer near the Bertie Limestone subcrop in the Genesee Valley north of the Fowlerville Moraine (fig. 4); water in these fracture zones also discharges to springs along the Onondaga escarpment east and west of the Genesee Valley. The Onondaga/Bertie contact along the escarpment is higher than the water table in the Genesee Valley and ground water in this area probably flows laterally (eastward or westward) to the Genesee Valley, rather than to the escarpment through the fracture zone.

Ground-Water Quality

The middle and lower aquifers in the Genesee Valley differ in chemical composition of ground water. Water in the middle aquifer is relatively fresh and has a low specific conductance (430 to 1,390 µS/cm, table 1). The principal chemical constituents are calcium and bicarbonate, which probably result from dissolution of calcite in glacial sediments derived from the underlying carbonate bedrock (fig. 12). The relatively high concentrations of sodium and potassium at some wells could reflect the exchange of calcium for sodium and potassium on the surface of clay particles through ion exchange.

Water in the lower aquifer generally is more saline than water in the middle aquifer—specific conductance values during 1995-96 ranged from 1,450 to 39,300 µS/cm, except near well Lv367 at location V near Sonyea (fig. 13), where it was 460 µS/cm (table 1). Water from this well was a calcium-bicarbonate type, similar to that in the middle aquifer (fig. 12). Water from other wells screened in the lower aquifer was a calcium/sodium-chloride type that probably reflects mixing with water from water-bearing zones in the underlying bedrock. Williams (1996) measured specific conductance values of 1,000 to 2,500 µS/cm in the lower aquifer during borehole conductance logging of five wells drilled near the collapse.

Fracture zones in the Onondaga and Bertie Limestones generally contain saline water with high specific conductance (72,000 to 106,200 µS/cm) (table 1). Williams (1996) measured specific conductance

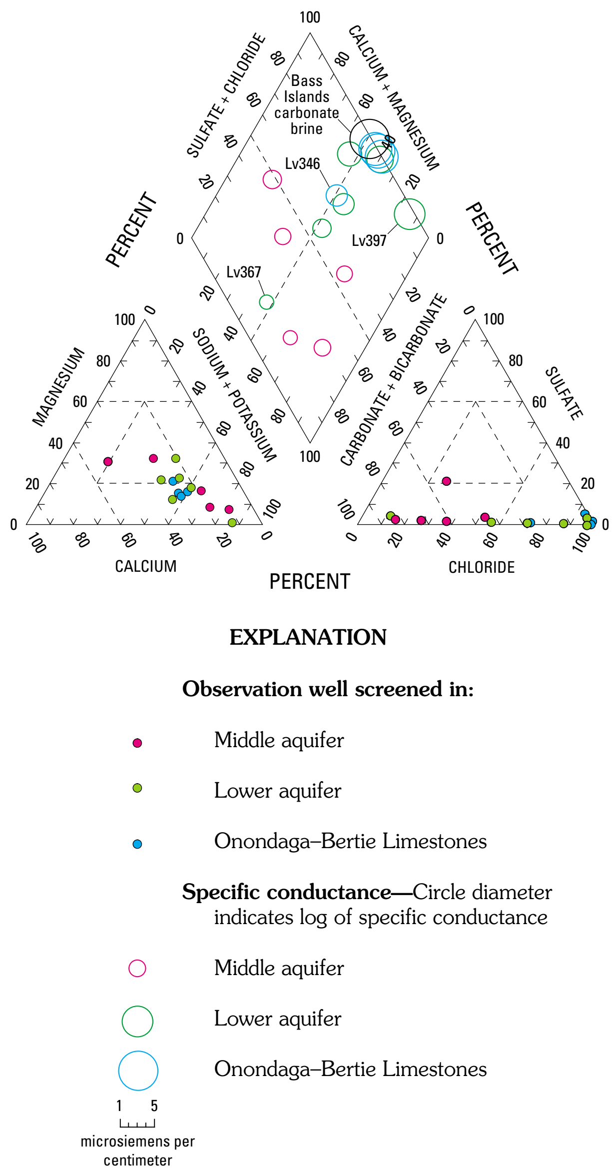


Figure 12. Percentage of major cations and anions in water from confined aquifers and Onondaga-Bertie Limestones in Genesee Valley, Livingston County, N.Y.

values of 10,000 to 50,000 µS/cm in these zones during borehole conductance logging of a well drilled near the collapse. The saline water is a calcium/sodium-chloride type (fig. 12) that is chemically similar to brines from oil-bearing zones in the Bass Island carbonates in western New York (New York State Department of Environmental Conservation (NYSDEC), 1992). The Bass Island carbonates, which are stratigraphically equivalent to the Rondout Formation (Rickard, 1969), overlie the Bertie Limestone in the southern part of the Genesee watershed but have been eroded within the study area. The lowest specific conductance value obtained from bedrock wells (3,040 µS/cm) was in a sample from well Lv346, screened in the upper part of the Onondaga Limestone at location II near the collapse (fig. 13). The water in this well probably originated from the overlying lower aquifer and, therefore, is not representative of other water-bearing zones in the bedrock.

Salinity

The ratio of chloride to bromide concentrations in saline water from the Onondaga and Bertie Limestones (120 to 180) is close to the average ratio of 110 reported for brines from oil-bearing zones in the Bass Island carbonates in western New York (NYSDEC, 1992). Samples from three of the wells screened in the lower aquifer (Lv370, Lv350 and Lv360) had chloride-to-bromide ratios of about 100, which indicates mixing with water from the Onondaga and Bertie Limestones. Water from well Lv370 at location I, near the subcrop of the Bertie Limestone near Avon, had the highest specific conductance (12,400 µS/cm) of these three wells (fig. 13). Water from Lv203, also screened in the lower aquifer near the Bertie subcrop, had a specific conductance of 12,100 µS/cm in 1966 (Kammerer and Hobba, 1967). The highest specific conductance measured in the lower aquifer (39,300 µS/cm) was in a sample from well Lv397, which had a chloride-to-bromide ratio of

Table 1. Concentrations of inorganic constituents of water from confined aquifers in the Genesee Valley, Livingston County, N.Y.

[Locations are shown in fig. 4. Concentrations are in milligrams per liter except as noted. µS/cm, microsiemens per centimeter at 25°C. Ca, calcium; Mg, magnesium; Na, sodium; K, potassium; SO_4, sulfate; Cl, chloride; Br, bromide; $CaCO_3$, calcium carbonate; <, less than.]

			Constituent or physical property								
Location	**Well number**	**Sampling date**	**Ca**	**Mg**	**Na**	**K**	**Alkalinity, as $CaCO_3$**	**SO_4**	**Cl**	**Br**	**Specific conductance (µS/cm)**
Middle aquifer:											
I	Lv371	11/95	180	66	58	36	370	150	140	< 1	1390
III	Lv351	11/95	22	8.6	97	8.7	150	11	130	1.7	700
IV	Lv361	10/95	44	29	61	4	240	8	100	1.6	790
IV	Lv362	8/96	15	6.8	74	110	260	8.2	67	< 1	870
V	Lv366	11/95	16	8.8	64	2	190	6.2	23	< 1	430
Lower aquifer:											
I	Lv370	8/96	450	230	1,400	67	86	182	4300	44	12,400
II	Lv368	8/96	180	44	280	140	460	< 14	840	2.2	2980
III	Lv350	10/95	530	220	880	27	430	19	2100	22	6670
IV	Lv360	11/95	62	34	150	7.2	320	7.7	300	2.8	1450
V	Lv367	8/96	19	18	45	7.9	240	13	24	< 1	460
—	Lv397	10/95	1000	30	8,100	64	460	120	14,300	32	39,300
Onondaga and Bertie Limestone											
I	Lv369	11/95	4600	1500	10,600	360	440	2100	29,400	230[a]	72,200
II	Lv346	10/95	150	71	310	13	410	7.3	750	7.6	3040
III	Lv349	8/96	6800	2100	16,000	780	270	320	47,000	260	106,200
V	Lv365	8/96	3900	1600	11,000	450	240	49	32,500	260	72,000

[a]Sampled 8/96

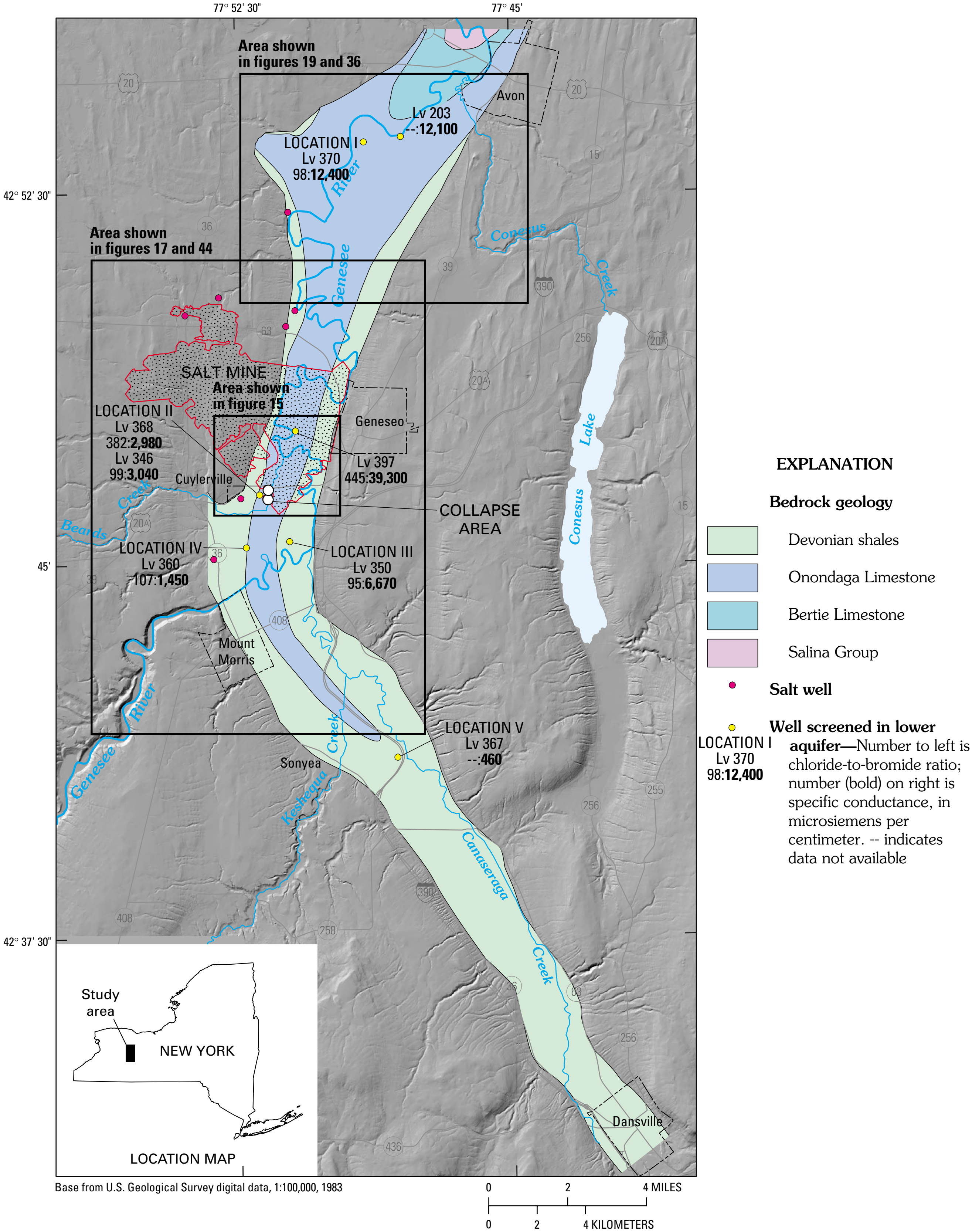


Figure 13. Locations of water-sampling sites I through V in Genesee Valley, Livingston County, N.Y., with (1) specific conductance, and (2) ratios of chloride-to-bromide concentrations, in wells screened in the lower aquifer.

445. This ratio and that of well Lv368 (382) at location II near the collapse is much higher than the ratio in water from the Onondaga and Bertie Limestones; this suggests that the ground water in this area has mixed with water containing dissolved halite, which has a chloride-to-bromide ratio of over 1,000 (Whittemore, 1988; Davis and others, 1998). Halite from the Salina salt beds has been mined by solution through several wells in the Genesee Valley in the past 100 years (fig. 13), and leakage from mining operations or brine disposal could have contaminated the lower aquifer in some areas. Improperly abandoned gas wells that penetrate the Salina salt beds could also allow upward leakage of saline water into the lower aquifer.

Isotopic Composition

Waters from the two confined aquifers have similar isotopic compositions of hydrogen (δD) and oxygen ($\delta^{18}O$) (table 2), and most samples plot along the global meteoric water line (GMWL) with a mean isotopic values of -75 ‰ δD and -10.8 ‰ $\delta^{18}O$ (fig. 14*A*). These values are near the mean composition for samples from a shallow dolomite aquifer in Niagara Falls (fig. 1) (-75 ‰ δD and -10.4 ‰ $\delta^{18}O$; Yager and Kappel, 1998) and, therefore, reflect the average values for recharge in western New York. Water samples from the Genesee River in January 1996 also plot near the GMWL, but the isotopic values are more negative (depleted) than those from the confined aquifers because are lower in winter than in summer. The δD and $\delta^{18}O$ values of samples from two wells screened in the middle and lower aquifers at location III, 1.2 mi southeast of the collapse (Lv350 and Lv351), are more depleted than in other wells in the valley; this suggests poor hydraulic connection with other parts of the confined-aquifer system; the depletion also suggests that recharge in this area occurred at lower temperatures than in other parts of the confined-aquifer system, but the reason is unknown.

The low tritium concentration in 9 of the 15 samples from the confined aquifers (less than 3 pCi/L, fig. 14*B*), indicates that most of the recharge entered the system before atmospheric testing of nuclear weapons began in 1953 (Michel, 1989). The slightly elevated tritium concentration in 5 of the 6 remaining samples (3 to 11 pCi/L) suggests either that (1) recharge in these areas entered the confined aquifer after atmospheric testing of nuclear weapons began in 1953 but before peak tritium concentrations were recorded in 1962 (Michel, 1989), or (2) post-1962 recharge, which had an elevated tritium concentration, has mixed with older water to yield the observed intermediate concentration. The sample with the highest tritium concentration (67 pCi/L) is from the middle aquifer at location I, near Avon (Lv371); this indicates that the most recent recharge entered this shallow part of the confined-aquifer system after 1962.

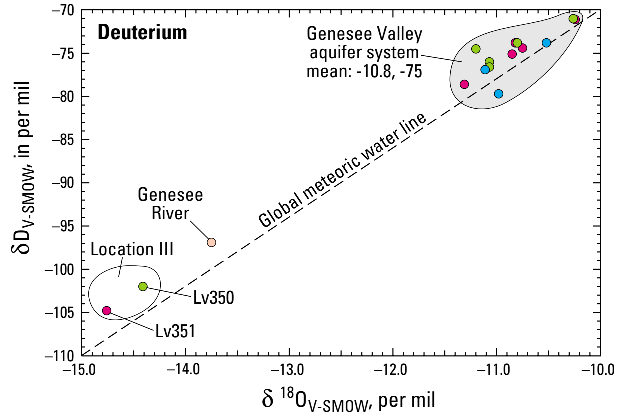
Table 2. Concentrations of stable isotopes and tritium in water from confined aquifers in the Genesee Valley, Livingston County, N.Y.

(‰, per mil; pCi/L, picocuries per liter; —, unnamed location. Locations are shown in fig. 4)

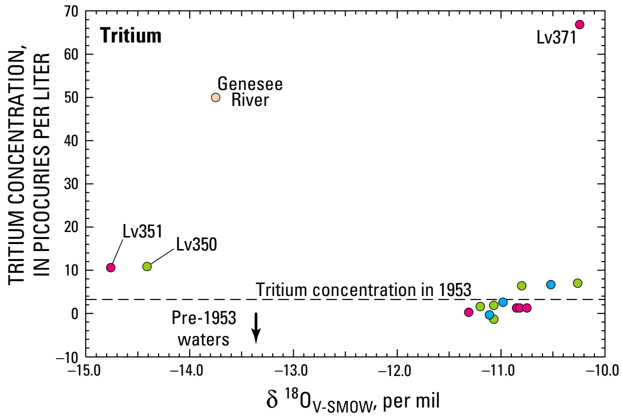
Location	Well number	Constituent		
		δD (‰)	$\delta^{18}O$ (‰)	Tritium (pCi/L)
Middle aquifer:				
I	Lv371	-71.1	-10.24	66.9
II	Lv348	-74.4	-10.75	1.3
III	Lv351	-104.8	-14.76	10.6
IV	Lv361	-73.8	-10.82	1.3
IV	Lv362	-75.1	-10.85	1.3
V	Lv366	-78.6	-11.31	.3
Lower aquifer:				
I	Lv370	-71	-10.26	7
II	Lv368	-76.6	-11.07	-1.3
III	Lv350	-102	-14.41	10.9
IV	Lv360	-73.8	-10.8	6.4
V	Lv367	-76	-11.07	1.9
—	Lv397	-74.5	-11.2	1.6
Onondaga and Bertie Limestones				
I	Lv369	-76.9	-11.11	-0.3
II	Lv346	-73.8	-10.52	6.7
III	Lv349	-79.7	-10.98	2.6

Natural Gas

Natural gas is commonly encountered during drilling in confined aquifers in the Genesee Valley (Ronald Hall, Dansville Water Wells, oral commun., 1996). Methane and hydrogen sulfide were reported during the sampling of several wells screened in confined aquifers in 1966 (Kammerer and Hobba, 1967). Gas from four wells sampled in 1994 had a $\delta^{13}C$ value of -55.1‰ and methane-to-ethane ratios of 54 to 135 (John C. Fountain, Geology Department, State University of New York, Buffalo, written commun., 1995). The relatively light isotopic composition of carbon in



A. Relation between deuterium and oxygen-18.



B. Relation between tritium and oxygen-18.

EXPLANATION

- **Middle aquifer**
- **Lower aquifer**
- **Onondaga/Bertie limestone**
- **Genesee River**

Figure 14. Concentrations of stable isotopes and tritium in confined aquifers in Genesee Valley, Livingston County, 1995-96: ***A.*** δD in relation to δ^{18}O. ***B.*** Tritium in relation to δ^{18}O.

the methane (δ^{13}C), and the low ethane content of gas from wells screened in the confined aquifers, indicate that the gas is produced by microbial decomposition of organic matter. In contrast, gas produced from thermal decompostion of organic matter in western New York typically has δ^{13}C values of -32 to -40.9‰, a methane-to-ethane ratio less than 20, and average methane and ethane contents of 85 percent and 7 percent, respectively (John C. Fountain, written commun., 1995; Mark Pearce, Gypsum Energy, written commun., 1995). The gas produced by microbial decomposition probably enters the ground water through diffusion from the Devonian shale bedrock that borders the confined-aquifer system.

EFFECTS OF MINE COLLAPSE AND FLOODING

The collapse within the salt mine affected hydrogeologic conditions not only in the sedimentary bedrock units overlying the mine (including the mined salt unit), but in the overlying glacial deposits. The collapses propagated upward through 600 ft of bedrock and 530 ft of unconsolidated sediments to land surface. Ground water began to flow into the mine from above after the collapses, and the fall of rock eventually caused a vertical zone or “chimney” of rock rubble to form within the overlying bedrock and glacial aquifer system. This rubble zone provided a major vertical pathway for ground-water flow from overlying water-bearing units.

The collapse and mine flooding resulted in 15 to 70 ft of land subsidence over the two mining panels and as much as 15 ft of subsidence over other parts of the mined area. Water levels in the lower aquifer had declined as much as 400 ft by January 1996, by which time the mine was completely flooded. Water-level declines of 50 ft and 135 ft were recorded at wells 7 mi north of the collapse area and 8 mi south of it, respectively, and over a dozen domestic wells screened in the middle aquifer in the collapse area went dry.

The decline in water levels decreased the pressure in the aquifers, causing (1) compression of fine-grained sediment in confining layers, which in turn resulted in land subsidence of 0.8 ft 2,900 ft south of the collapse and as much as 0.3 ft in Mt. Morris, 3 mi south of it, (2) exsolution of natural gas from ground water (three or four water wells began to produce gas), and (3) a reversal in the direction of ground-water flow. This southward reversal is probably the reason for the apparent intrusion of saline water into the lower aquifer at the north end of the Genesee Valley.

Collapse of Mine Ceiling

The salt bed that was mined at Retsof is 13 to 30 ft thick and slopes from an altitude of about 280 ft below sea level in the northern part of the mine to 560 ft below sea level in the southern part. Salt was mined at a depth of about 1,100 ft below land surface (altitude of 540 ft below sea level) in the collapse area. Most of the salt from the mine was extracted by a "room-and-pillar" technique that leaves large pillars (at least 75 ft by 75 ft) spaced about 65 ft apart (fig. 15). A different mining method—a "yielding pillar" technique—was used after 1990 in the southern part of the mine to control rockfalls and floor buckles (John T. Boyd Company, 1995; Denk and others, 1994). The yielding-pillar technique allowed excavation of large mining panels as much as 1,000 ft long, consisting of 40-ft-wide rooms separated by many small pillars (20 ft by 20 ft) that are designed to deform readily and transfer the roof load onto the walls and into the overlying rock. Both collapses occurred in yielding-pillar areas, but whether this technique contributed to the ceiling collapse, or whether the collapse resulted from structural weakness in the overlying bedrock, is uncertain. Reports that discuss possible causes for the collapses are summarized in NYSDEC(1997).

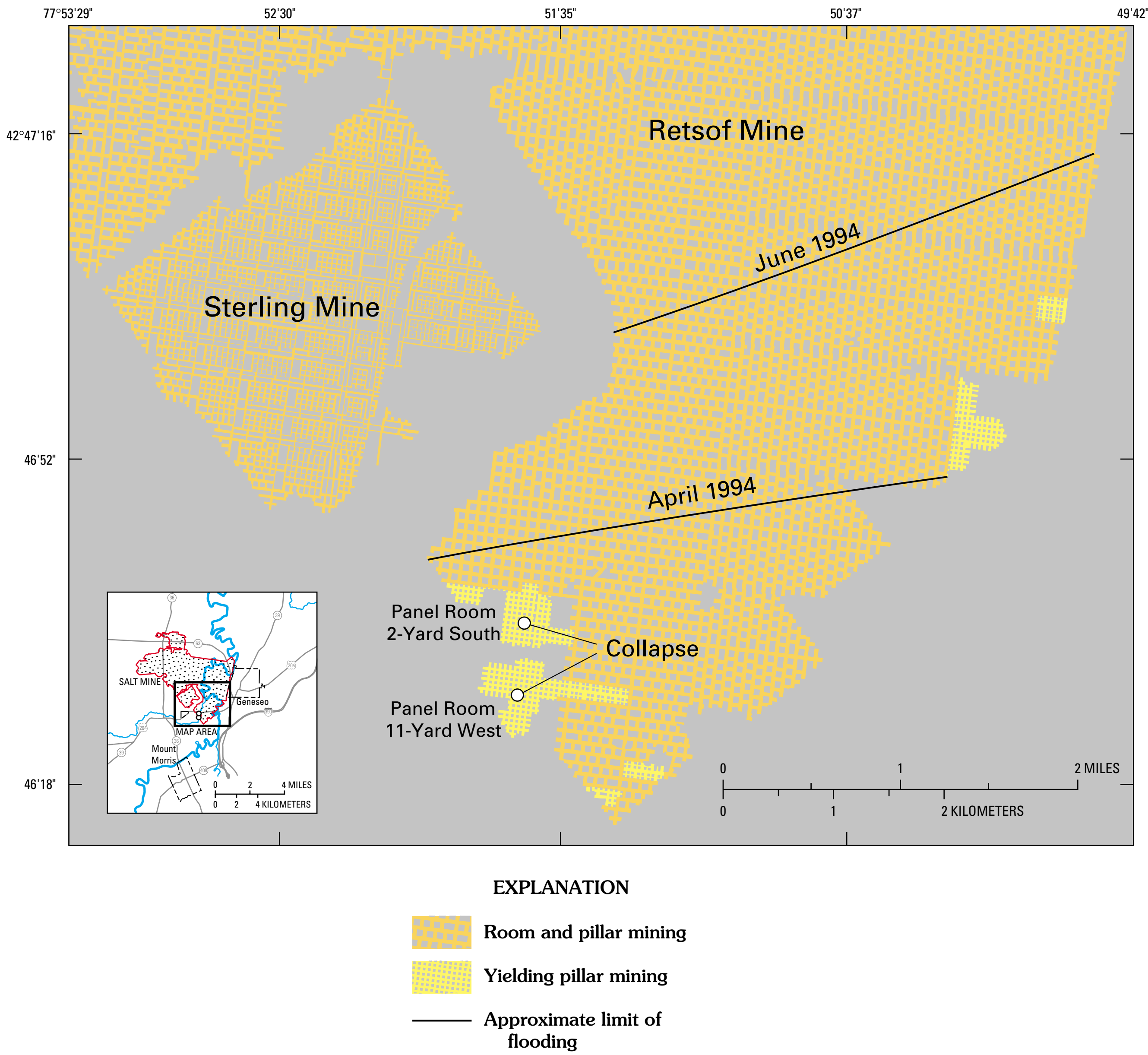


Figure 15. Locations of rooms and pillars in Retsof salt mine, Livingston County, N.Y., and extent of flooding in April and June 1994. (Location is shown in fig. 13.)

On March 12, 1994, the ceiling collapse in a panel containing yielding pillars (2-Yard South, fig. 15) caused a seismic event (magnitude 3.6) that was recorded by several siesmograph stations (NYSDEC, 1997). The resultant fracturing of shale overlying these panels formed a vertical rubble zone that propagated upward to the Bertie and Onondaga Limestones and allowed water from water-bearing zones to drain into the mine (fig. 16). The leaking ground water was undersaturated with respect to halite and caused rapid dissolution of salt, further weakening the pillars that supported the mine ceiling. The weight of the glacial sediments overlying the carbonate rocks caused the latter to collapse, allowing water from the lower aquifer to flow down through the rubble zone and into the mine. Subsequent seismic surveys indicated that the bedrock surface over part of the collapsed area subsided about 45 ft (Dobecki Earth Sciences, Inc., 1994). The upward propagation of subsidence through the glacial sediments caused a sinkhole to form at land surface on April 6, 1994 over panel 2-Yard South (fig. 3). The rate of ground-water flow into the mine increased abruptly after a second seismic event on April 8, 1994 (NYSDEC, 1997). The appearance of a second sinkhole over panel 11-Yard West on May 25, 1994, several weeks after the initial collapse, indicated that additional collapses had occurred either within the second panel or within cavities in the overlying shale bedrock.

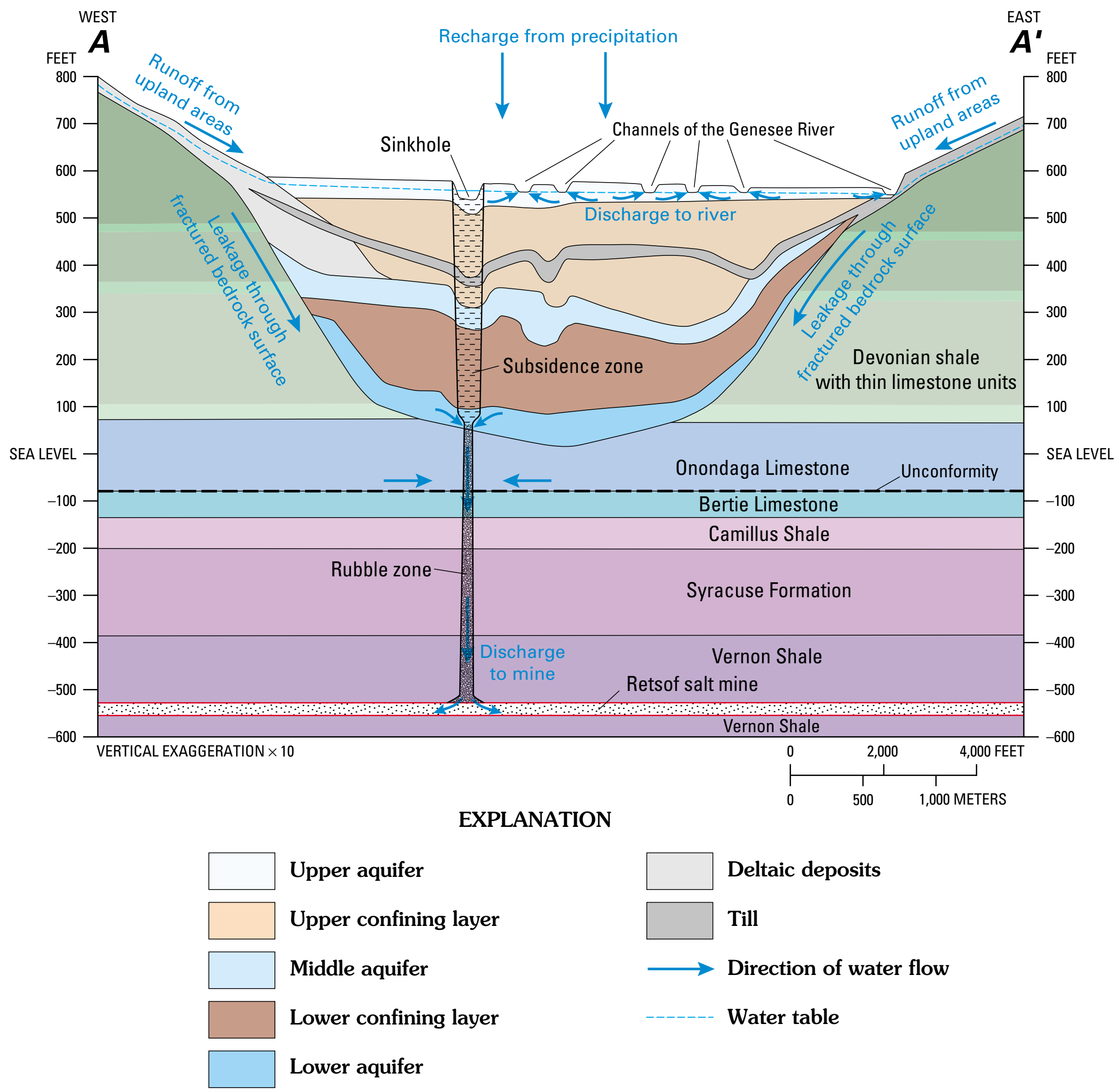


Figure 16. Stratigraphic section A–A′ depicting rubble zones over collapsed rooms in Retsof salt mine, Livingston County, N.Y. (Line of section is shown in fig. 8.)

In an attempt to save the mine, over thirty wells were drilled in April 1994 to either pump water from the mine or to inject grout into panel 11-Yard West to strengthen the weakened mine pillars. The pumping efforts were discontinued by the end of April 1994, however, after the rate of flooding exceeded the pumping capability, and attempts to prevent leakage of ground water to the mine by injecting grout into the rubble zone were unsuccessful and were abandoned after the second sinkhole appeared over panel 11-Yard West on May 25.

Rates of flow into the mine were estimated to be 5,500 gal/min on March 14, 1994 and 20,000 gal/min on April 8, 1994, as calculated from the rate of water-level rise within the mine and from estimates of the volume of flooded area (L.D. Milliken, ANSI, written commun., 1996). Estimated flow rates remained relatively constant at about 20,000 gal/min over the next 20 months but became less certain as older parts of the mine became flooded. Methane and hydrogen sulfide concentrations in the mine increased as the flooding progressed; in May 1995, gas was flared (burned) from three wells drilled in the collapse area. Mining ceased in September 1995, and the mine was completely flooded by January 1996. By this time all shafts leading to the mine had been sealed, and gas emissions from the three flare wells had ceased. Gas seeps in streambeds overlying the mine perimeter were observed during and after flooding of the mine.

Land Subsidence

Land subsidence within the Genesee Valley, whether caused directly or indirectly by the mine collapse, extended more than 3 mi north and south from the collapse area. Subsidence directly over the mine area ranged generally from 1 to 15 ft by February 1996 (fig. 17), except in the two sinkholes, where it was estimated to be as much as 70 ft (NYSDEC, 1997). Surface subsidence of 0.1 to 0.8 ft extended southward beyond the mine to Mt. Morris, 3 mi south of the collapse, where as much as 0.3 ft of subsidence was measured.

Most of the subsidence over the mine was caused by closure of the mine cavity; this process was accelerated by the dissolution of the salt pillars by water that flooded the mine. Mining engineers expect the land surface above most of the mine to subside 8 to 9 ft over the next 100 years, but the future rate of subsidence could decrease because the mine is now filled with brine that will support part of the weight of overlying sediment (van Sambeek, 1994; Shannon and Wilson, 1995). The subsidence is expected to cause fractures at land surface along the margins of the mined area (John T. Boyd Company, 1995). These fractures, and the tilting of the land surface in the direction of the mine, have damaged a highway bridge and four houses over the mined area; continued subsidence could damage other structures.

Additional subsidence above and beyond the mined area is expected to result from the compression of fine-grained sediments in the confining layers within the aquifer system (fig. 10). The compression results from the drainage of water from the fine-grained sediments as the intergranular stress (effective stress) increases in response to the declining water levels (pore pressure) in the aquifer system. Effective stress σ_e is defined as

$$\sigma_e = \sigma - p \qquad (1)$$

where σ = geostatic pressure [FL^{-3}] that results from the weight of the sediment and water above a given depth, and
p = pore pressure [FL^{-3}].

As the pore pressure decreases, the effective stress increases, and more of the weight is borne by the sediment matrix. Consequently, the pore volume n in the sediment decreases in proportion to the increase in effective stress $\Delta\sigma_e$ and the compressibility of the material α. The pore-volume decrease Δn is

$$\Delta n = -\alpha\Delta\sigma_e . \qquad (2)$$

The reduction in pore volume results in compression of the sediment and the expulsion of pore water, and the cumulative result of compression of sediments at various depths is subsidence of the land surface. The compressibility of fine-grained sediments is 2 or 3 orders of magnitude greater than that of coarse sediments and accounts for most of the subsidence (Freeze and Cherry, 1979). The simulation of this process in response to the collapse is described further on.

Surface Water

Land subsidence over the mined area lowered the channel of the Genesee River as much as 4.5 ft along a reach extending about 1.5 mi upstream from the mouth of Beards Creek (fig. 17) and widened the

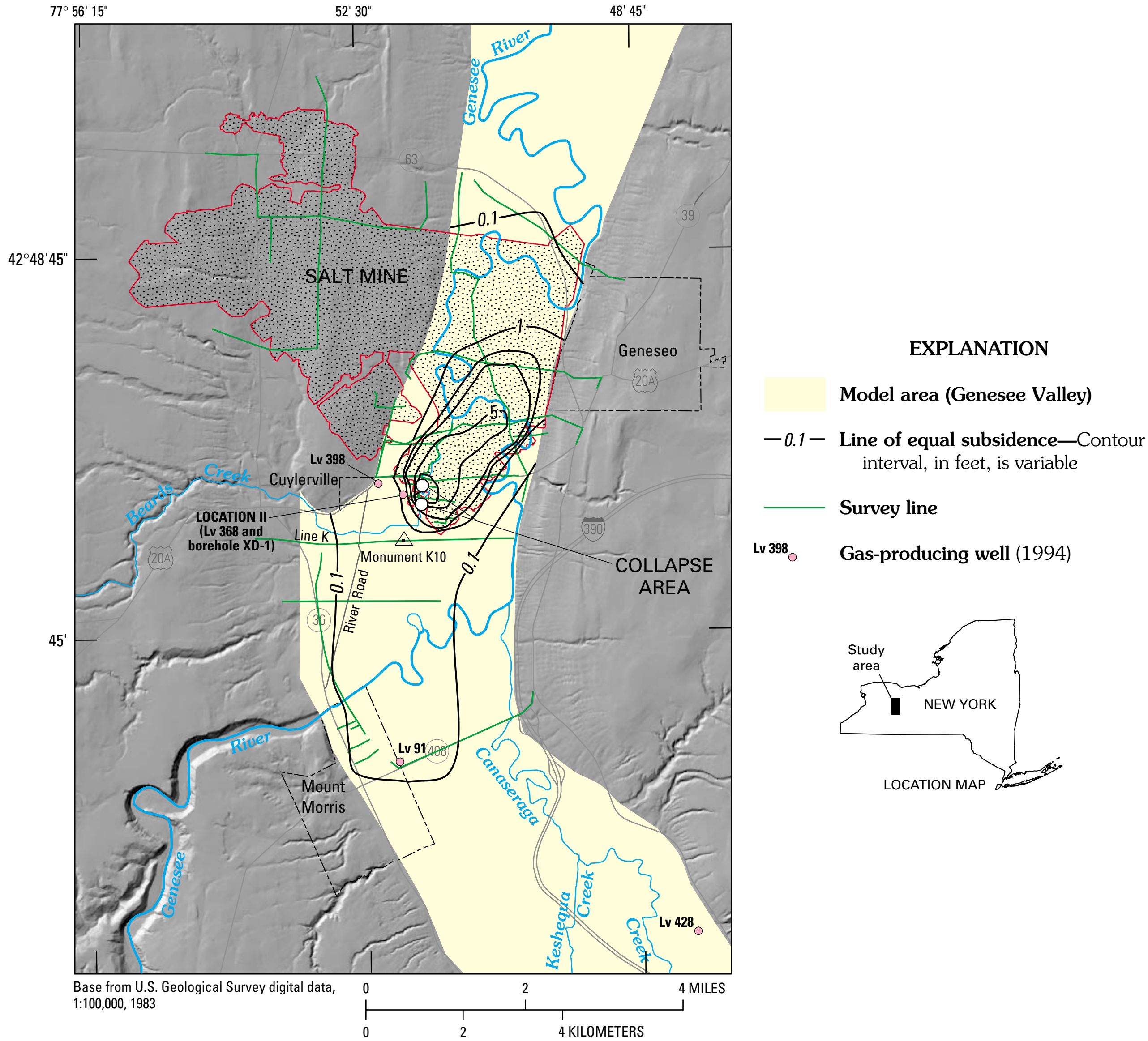


Figure 17. Land subsidence in vicinity of the Retsof salt mine, Livingston County, N.Y., February 1996. (Location is shown in fig. 13).

10-year flood plain by about 340 acres (Musetter Engineering, Inc., 1996). Increased sediment deposition in this area could temporarily decrease the sediment supply to downstream reaches and cause a slight increase in bank erosion and a widening of the channel. Increased bank erosion along the bend in the Genesee River just south (upstream) of the mined area along Route 20A (fig. 17) is expected and could necessitate repairs to a road embankment (Musetter Engineering, Inc., 1996).

Subsidence also altered the course of Beards Creek, causing it to flow through the collapse area and create a pond in each of the sinkholes that formed over the collapses. The subsidence of the creekbed initiated headward erosion of a knickpoint, which had progressed about 0.5 mi upstream by June 1995, lowering the creekbed by as much as 12 ft and widening it by about 40 ft (Musetter Engineering, Inc., 1996). Upstream progression of the knickpoint could undermine a highway bridge along River Road (fig. 17); therefore, grade-control structures were constructed to protect the bridge.

Streamflow measurements were made along Beards Creek and along the Genesee River from Mt. Morris to Avon to assess whether surface water was discharging into the mine. Measurements in Beards Creek on May 3, 1994 indicated that streamflow in the reach within the collapse area (Location II, fig. 17) increased about 15 percent—from 16.4 to 18.8 ft^3/s; this indicates that ground water was discharging to the creek and that no streamflow was infiltrating through the sinkholes to the mine. On July 14, 1994 all of the flow in Beards Creek (0.39 ft^3/s) was lost through infiltration to the underlying alluvial sediments upstream of the collapse area, and reappeared in the stream downstream from the collapse area; this indicates further that streamflow was not infiltrating to the mine.

The low vertical hydraulic conductivity (9×10^{-5} ft/d) and large thickness (250 ft) of the fine-grained material that underlies streams in the Genesee Valley probably prevents downward flow of water from the streams to the confined aquifers and the mine. Some reaches of the Genesee River lie adjacent to the valley wall, however; where water from the river could infiltrate downward through bedrock surface fractures to the confined aquifers. Streamflow measurements were made along the Genesee River from Mt. Morris to Avon on August 28, 1995 to assess this possibility during a period of relatively low flow (69 ft^3/s), when streamflow was not affected by regulation at upstream dams. The measurements indicated an increase in streamflow (by ground-water discharge) of 19 ft^3/s, which is within the range of 15 to 20 ft^3/s estimated in this study for this reach in similar low-flow periods during a drought in the early 1960's. The measurements indicate that the declining ground-water levels in the confined aquifers did not induce infiltration from the Genesee River and, again, indicate that water from the river did not enter the mine.

Ground-Water Hydrology

The collapses in the salt mine in March and April 1994 resulted in new ground-water flow paths from the lower aquifer to the mine and away from any downvalley outlet(s) farther north along the Genesee Valley. Before the collapses, ground water discharged through the downgradient boundary of the study area near Avon, where the hydraulic head was about 520 ft, as calculated from the estimated water-table altitudes (fig. 11). The formation of the rubble zone in the bedrock between the mine and the confined lower aquifer was analogous to "pulling the plug in a bathtub" and opened a drain in the bedrock surface at an altitude of about 80 ft. Water from the lower aquifer and bedrock fracture zones flowed through the permeable rubble zone and into the mine, which was at atmospheric pressure.

The rate of ground-water discharge to the mine exceeded the rate of recharge to the lower aquifer and caused water levels (hydraulic head) throughout the confined-aquifer system to decline. First the water levels declined abruptly, then continued to decline gradually for 2 years. Discharge from the lower aquifer to the rubble zone virtually ceased, however, once the hydraulic head in the mine had risen above the bedrock-surface altitude (80 ft above sea level) in January 1996. Once the mine was completely flooded and the rubble zone saturated, hydraulic head in the lower aquifer began to recover near the collapse area, although hydraulic heads in areas 8 mi north of the collapse had still not yet begun to recover by June 1996.

The discharge of ground water into the mine altered the hydraulic gradient in the lower aquifer and reversed the direction of ground-water flow for at least 9 mi north of the collapse area. After the gradient reversal, the former downgradient boundary of the study area near Avon became an upgradient boundary, and saline ground water from the north end of the valley flowed southward toward the collapse area. The decrease in hydraulic head in the lower aquifer also increased the vertical hydraulic gradient between the lower aquifer and the Bertie Limestone, and thereby enhanced upward migration of saline water from the bedrock to the aquifer. The intrusion of saline water caused increased salinity at five residential wells; this intrusion was first reported in March 1995 (Ralph VanHouten, Livingston County Department of Health, oral commun., 1996).

Ground-Water Levels

Water-level declines (drawdowns) resulted in both confined aquifers. Water levels in upland areas beyond the valley-fill aquifer system, and those at most of the wells screened in the upper aquifer, were unaffected by the mine collapse. Drawdowns at affected wells were calculated as the difference between measured water levels and those inferred for precollapse conditions. This calculation yielded conservative estimates of drawdown because hydraulic heads in the lower aquifer were probably higher than heads in the upper aquifer, as indicated by reports of artesian conditions in deep wells in the valley (Dr. Richard Young, Geology Department, State University of New York at Geneso, oral commun., 1996). The distribution of drawdown in January 1996, when water-level declines in the collapse area were greatest, is shown in figure 18,

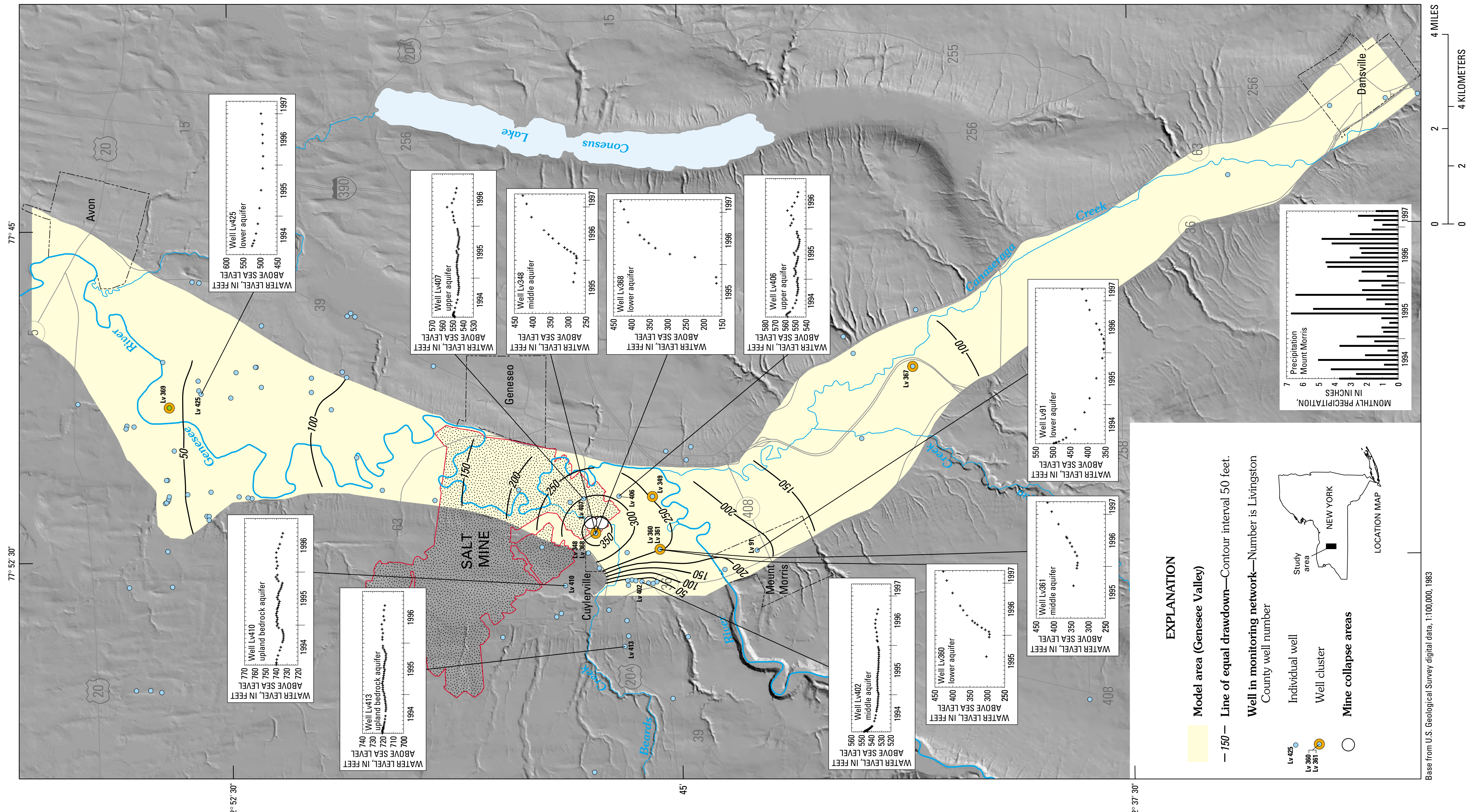


Figure 18. Drawdown in lower aquifer, Genesee Valley, Livingston County, N.Y., in January 1996, hydrographs showing hydraulic head in selected wells, and bar graph showing precipitation at Mt. Morris, 1994-97. (Location is shown in fig. 1.)

which includes hydrographs of water levels at selected wells.

The maximum drawdown in the lower aquifer was about 400 ft near the collapse area; drawdowns of 50 ft and 110 ft were recorded 7 mi north of the collapse area, and 8 mi south of it, respectively. The water level in well Lv91, 3 mi south of the collapse area, had declined about 60 ft by the end of March 1994 and by a total of 140 ft by March 1996 (fig. 18). The water level in well Lv425, 8 mi north of the collapse area, had declined more than 30 ft by the end of May 1994 and had declined an additional 30 ft by March 1996. Part of the hydrograph for well Lv368 (closest to the collapse area, location II) is missing because the well was capped and, therefore, no water levels could be measured after it began to produce gas after an attempt to obtain a water sample in October 1995. The hydrograph for well Lv346, screened in the Onondaga Limestone at the same location, indicates that recovery of the aquifer system began at the beginning of January 1996.

The maximum drawdown in the middle aquifer was about 250 ft in the collapse area; drawdowns of 50 ft and 135 ft were recorded 7 mi north of the collapse and 8 mi south of it, respectively. Despite the large drawdowns, the water level in the collapse area remained above the top of the middle aquifer, and confined conditions were maintained. Eight to ten wells screened in an aquifer in the Beards Creek valley east of state highway 36 (fig. 18) went dry by the end of March 1994, however, and about six wells 8 mi north of the collapse area in the Genesee Valley went dry between May and November 1994. Hydraulic heads in the middle aquifer near the collapse area (well Lv348) began to recover in February 1996, about 1 month after heads in the lower aquifer began to recover.

Water levels in water-bearing zones in the Onondaga and Bertie Limestone declined about 210 ft at location III (well Lv349), about 1.2 mi southeast of the collapse (fig. 18). Drawdown in the water-bearing zones was greater south of the collapse than to the north of it; drawdowns totaled about 70 ft at location V near Sonyea (well Lv365, about 7 mi south of the collapse) and only 7 ft at location I near Avon (well Lv369, 8 mi north of the collapse). The water level in each of these three wells was converted to freshwater hydraulic head, h_{fw}, by the following relation, to correct for the higher density of the saline water in these aquifers:

$$h_{fw} = \frac{p}{\gamma_{fw}} + z \qquad (3)$$

where p = pressure at the bottom of the well screen $[ML^{-1}T^{-2}]$,
γ_{fw} = specific weight of freshwater $[ML^{-2}T^{-2}]$, and
z = altitude of the bottom of the well screen [L].

The pressure p is given as

$$p = \gamma l \qquad (4)$$

where γ= specific weight of the saline water $[ML^{-2}T^{-2}]$, and
l = depth of water above the bottom of the well screen [L].

Substituting equation 4 into equation 3 indicates that the freshwater head is higher than the measured head by a factor of γ/γ_{fw}. This factor was computed to be 1.01 for the saline water in the Onondaga and Bertie Limestones, from the measured density of water samples from wells Lv369 and Lv371 at location I near Avon. The corrected freshwater head was 8.2 ft higher than the measured head at location V near Sonyea, 4.4 ft higher than that at location III 1.2 mi southeast of the collapse, and 1.6 ft higher than that at location I.

Water levels in wells located in the bedrock uplands (Lv413 and Lv410) and in the upper aquifer (Lv406 and Lv407) near the collapse area fluctuated seasonally from 1994-97, declining through the summer and fall, then rising in response to recharge through the following winter and spring (see fig. 18). Water levels in the upland wells showed no persistent decline from March 1994 through January 1996 (when those in the confined aquifers declined), and began their seasonal rise in October or November 1995, 2 months before ground-water discharge to the mine ceased; this suggests that upland areas were not affected by the mine collapse. Water levels in the upper aquifer reached a minimum in November 1995, however, and did not recover to a seasonal maximum until the end of January 1996, after ground-water discharge to the mine had ceased. Water levels in the upper aquifer also were lower in the winter of 1995-96 than in the following winter (1996-97), despite the greater amount of precipitation in 1995. These observations indicate that water levels in the upper aquifer near the collapse area were affected by the mine collapse but were lowered by only about 5 ft from March 1994 through January 1996.

Ground-Water Quality

Five domestic wells screened in the lower aquifer in the northern part of the Genesee Valley south of Fowlerville Road (fig. 19) began to produce water in 1995 whose salinity rendered the water unsuitable for consumption. Well Lv457 on Little Road (fig. 19) yielded freshwater when it was drilled in September 1994 (Ronald Hall, Dansville Water Wells, oral commun., 1996), but the elevated specific conductance (12,500 µS/cm) in March 1995 indicated that saline water had entered the well during the winter of 1994-95. The specific conductance of water samples from well Lv500, which is 3,700 ft southwest of Lv457 on Hogmire Road (fig. 19), increased from 4,700 to

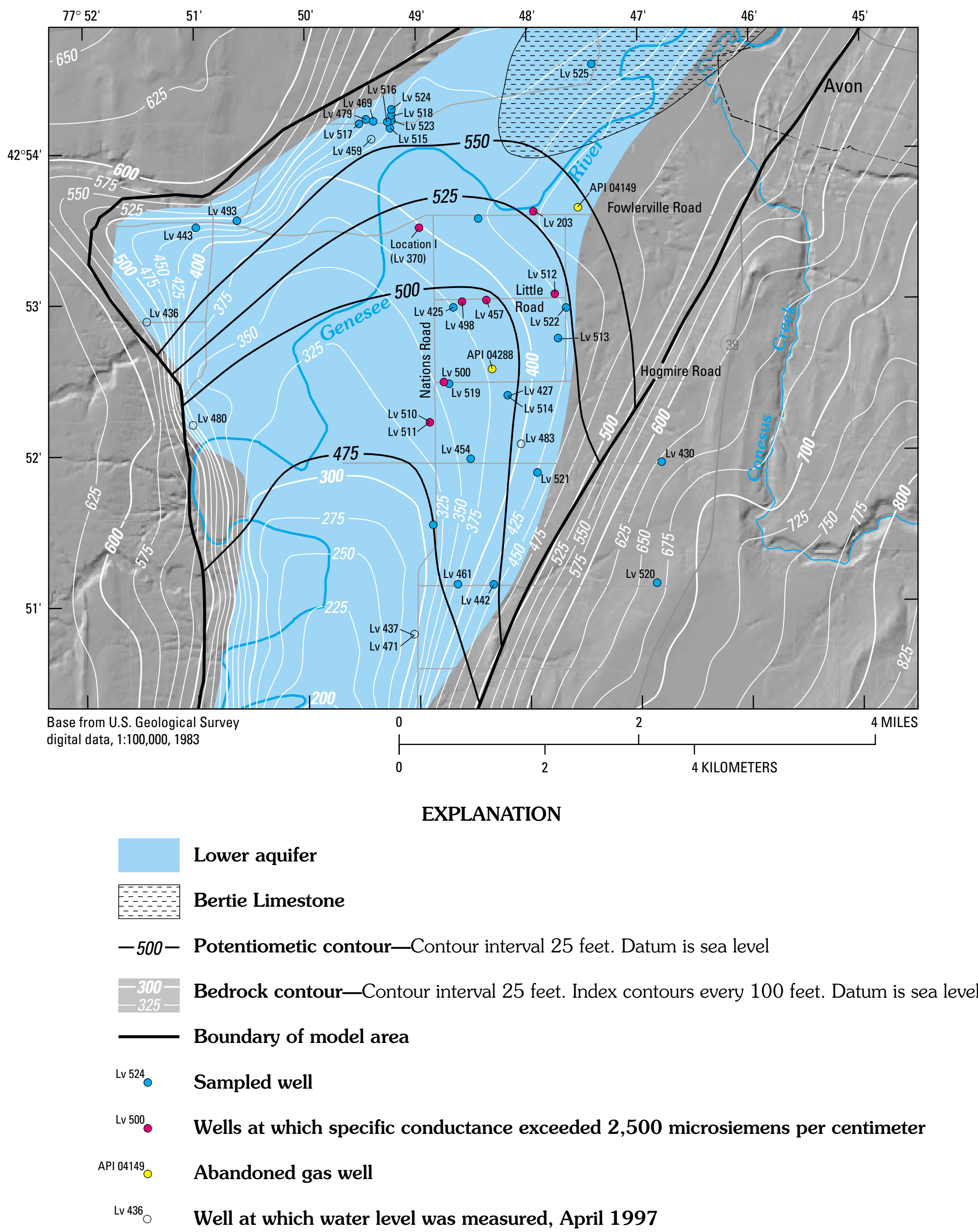


Figure 19. Locations of sampled wells in Genesee Valley, Livingston County, N.Y., and wells in which saline water was detected in April 1997. (Location is shown in fig. 13.)

12,400 µS/cm between June 1995 and January 1996. Six wells in the northern part of the Genesee Valley had elevated specific conductance values (greater than 2,500 µS/cm) in April 1997 (fig. 19); a seventh well along Fowlerville Rd. (Lv203) had a specific conductance of 12,100 µS/cm in 1966, as mentioned previously.

Water samples were obtained from 29 wells and 1 spring in the northern end of the Genesee Valley in April 1997 to investigate the source of salinity (fig. 19 and table 3). Nine of the wells were screened in the middle aquifer and 17 wells were screened in the lower aquifer. The remaining wells and the spring produced water from bedrock. Analyses of water from three other wells in this area are also available: Lv371 and Lv397 (middle and lower aquifer, respectively, at location I), sampled in 1995, and Lv203 (lower aquifer), sampled in 1966. Specific conductance values are generally lower in the middle aquifer than in the lower aquifer; the median specific conductance of samples unaffected by saline water was 600 µS/cm in the middle aquifer and 1300 µS/cm in the lower aquifer (groups 1 and 2 in figs. 20 and 21). Specific conductance values greater 2,500 µS/cm of water from six wells screened in the lower aquifer (group 3, figs. 20 and 21) are attributed to saline water migration.

Table 3. Concentrations of inorganic constituents in water from confined aquifers in Genesee Valley, Livingston County, N.Y., April 1997.

[Locations are shown on fig. 19. Concentrations are in milligrams per liter except as noted. µS/cm, microsiemens per centimeter at 25° C; Ca, calcium; Mg, magnesium; Na, sodium; K, potassium; SO_4, sulfate; Cl, chloride; Ba, barium; Sr, strontium; Br, bromide; $CaCO_3$, calcium carbonate; <, less than.; —, not analyzed]

Well number	Owner	Constituent or physical property											
		Ca	Mg	Na	K	Alkalinity, as $CaCO_3$	SO_4	Cl	Ba	Sr	Br	Specific conductance (µS/cm)	Dissolved solids
Middle aquifer:													
Lv442[a]	Millard	36	48	72	1.6	180	96	24	.033	9.7	.13	890	513
Lv443[b]	Cullinan	52	37	15	< 5	310	7	13	1.7	3.7	< 1	550	450
Lv458[b]	Crandall	64	41	37	< 5	250	60	43	< .2	2.6	< 1	620	378
Lv493[b]	Kelly	24	37	20	< 5	260	< 7	20	1.7	4.7	< 1	480	287
Lv513[a]	McKeown	17	16	84	2.1	150	8.1	44	3.0	34	.4	610	389
Lv518[b]	Orologio, B.	42	34	38	< 5	290	30	36	2	8.	< 1	620	319
Lv523	McMahon	64	56	151	8	246	42	296	2.2	14	2.7	1800	751
Lv524	Striker	41	34	43	< 5	254	8	36	.8	8.3	< 1	590	325
Lv525	House[d]	101	41	23	< 5	301	46	48	< .2	.17	< 1	844	400
Lower aquifer:													
Lv203[c]	Croston	—	—	—	—	—	1,430	3,540	—	—	—	12,100	—
Lv370[b]	Akzo Nobel	370	190	1400	190	40	33	3300	.9	35	39	15,100	7000
Lv425[a]	Hollinger	43	37	84	2.4	130	7.2	160	3.3	7.7	.66	950	511
Lv454[a]	Holt	24	16	83	20	140	29	23	.1	4.5	.15	550	347
Lv461[a]	Wadsworth	9.9	7.1	160	1	96	200	82	.014	3.5	.24	1010	627
Lv469[a]	Orologio, V.	74	52	160	5.9	160	25	290	.75	11	2.0	1620	867
Lv479[a]	Ward	68	52	150	5.4	160	33	280	.61	10	1.8	1800	839
Lv498[a]	Davis #3	120	110	280	4.4	120	12	840	9.4	22	2.5	3600	1630

Ratios of chloride to bromide concentrations in water samples from the lower aquifer indicate two sources of saline water. Eight of the 17 samples from the lower aquifer contained chloride-to-bromide ratios of 84 to 154, similar to ratios measured in Bass Island carbonate brine in western New York (110) and saline water from the Onondaga and Bertie Limestones (120 to 180). These data plot close to a line representing the average ratio in waters from the Onondaga and Bertie Limestones (150:1, fig. 22). Saline water at well Lv512 (dissolved solids concentration 6,710 mg/L) had a much higher chloride-to-bromide ratio (920), which is close to those characteristic of waters affected by halite dissolution (greater than 1,000:1) (Davis and others, 1998). Ratios for the remaining wells plot between the lines representing these two extreme ratios (fig. 22); this result indicates that the salinity of this water results from the mixing of saline water from the Onondaga and Bertie Limestones with water affected by halite from an unidentified source.

Saline bedrock water probably enters the lower aquifer where the Bertie Limestone subcrops north of Fowlerville Road (fig. 19). Before the mine collapse, saline water entering the lower aquifer probably flowed northward with the horizontal hydraulic gradient. The documented saline water at well Lv203 along Fowlerville Road (specific conductance 12,100 µS/cm in 1966, Kammerer and Hobba, 1967) marks the approx-

Table 3. Concentrations of inorganic constituents in water from confined aquifers in Genesee Valley, Livingston County, N.Y., April 1997—Continued

[Locations are shown on fig. 19. Concentrations are in milligrams per liter except as noted. µS/cm, microsiemens per centimeter at 25° C; Ca, calcium; Mg, magnesium; Na, sodium; K, potassium; SO_4, sulfate; Cl, chloride; Ba, barium; Sr, strontium; Br, bromide; $CaCO_3$, calcium carbonate; <, less than.; —, not analyzed]

		Constituent or physical property											
Well number	Owner	Ca	Mg	Na	K	Alkalinity, as $CaCO_3$	SO_4	Cl	Ba	Sr	Br	Specific conductance (µS/cm)	Dissolved solids
Lower aquifer:													
Lv500[b]	Baker/ Wadsworth	590	530	2400	39	130	240	5300	5.0	86	8.5	25,800	11,800
Lv510[a]	Jones (house)	80	86	260	3.8	110	7.8	700	7.6	23	1.5	3000	1430
Lv511[a]	Jones (barn)	42	42	160	2.9	110	11	300	1.5	14	.76	1300	753
Lv512[a]	Halpin	390	170	1500	20	120	490	3500	.23	47	3.8	18,400	6710
Lv514[a]	Banker	17	5.3	61	1.3	93	15	12	.14	2.5	.12	380	255
Lv515[a]	Klimasewski	56	43	110	3.9	150	18	240	1.8	11	1.7	1300	705
Lv516[a]	Zander	58	44	110	4.3	150	14	230	1.4	12	1.6	1280	844
Lv517[b]	Sullivan	71	53	120	9.4	320	48	220	.91	9.8	1.5	1420	676
Lv519[b]	Elliot	88	80	270	8.9	220	9	630	4.3	15.0	1.9	2600	1260
Lv521[b]	Harris	110	41	26	42	330	90	48	.32	.55	< 1	800	592
Lv522[b]	O'Connell[d]	128	41	60	41	330	79	89	< .2	2.6	< 1	1047	672
Bedrock:													
Lv430[b]	Lloyd	68	73	76	5.3	432	60	69	.58	4.8	< 1	1020	641
Lv450[b]	Morabito	58	35	20	< 5	210	40	40	< .2	.36	< 1	590	345
Lv520[b]	Morss	83	62	33	5	380	54	80	.2	.67	< 1	950	546

[a]Analysis by New York State Department of Health, Wadsworth Center, Albany, N.Y.
[b]Analysis by Lozier Laboritories, Rochester, N.Y.
[c]Analysis by Kammerer and Hoba, 1967.
[d]Aquifer uncertain.

imate southward limit of saline-water movement at that time because domestic wells further south produced potable water. After the mine collapse, the reversal of the horizontal hydraulic gradient in this area allowed saline water to flow southward from the Bertie subcrop toward the collapse area. The drawdown in the lower aquifer also increased the upward hydraulic gradient from below, thereby enhancing the potential for upward migration of saline waters through high-angle fractures in the bedrock south of the Bertie subcrop. These changes are probably the reason for the intrusion of saline water into areas south of Fowlerville Road.

The halite detected in water samples from the lower aquifer probably originated from the Salina salt beds. Two abandoned gas wells that penetrated the Salina salt beds are near the five domestic wells that yielded water with specific conductance values greater than 2,500 µS/cm (fig. 19). Halite-bearing water could have flowed upward through the gas wells or outside the well casing and leaked into the lower aquifer if the wells were not properly sealed when abandoned. Past solution mining of halite could explain the salinity in the sample from well Lv397 (fig. 13), but the other affected wells are north of the only known locations of solution mining, and the recent changes in salinity appear to have originated from the north; therefore, the salinity probably was not derived from these solution-mining sites. Halite-bearing water from road-salting

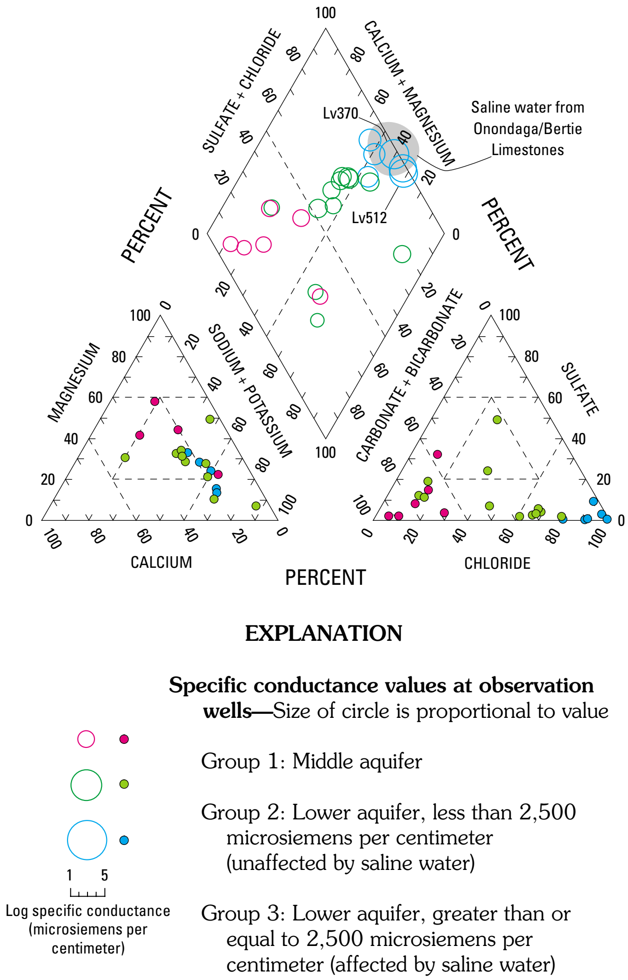


Figure 20. Percentage of major cations and anions in water from northern end of Genesee Valley, Livingston County, N.Y., April 1997.

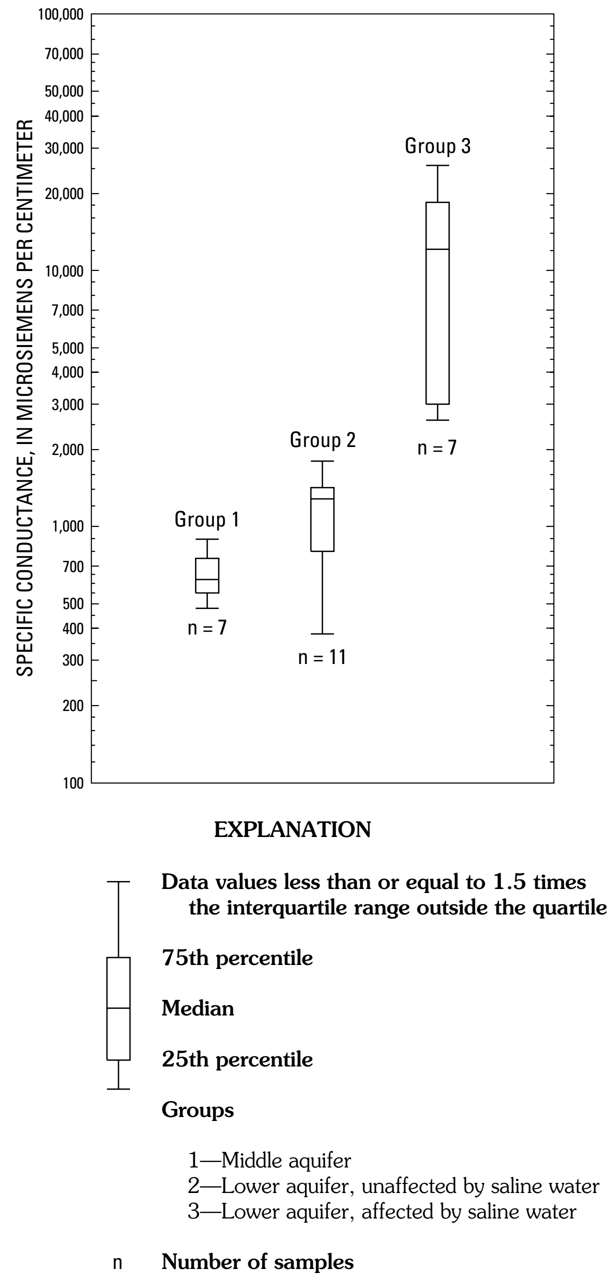
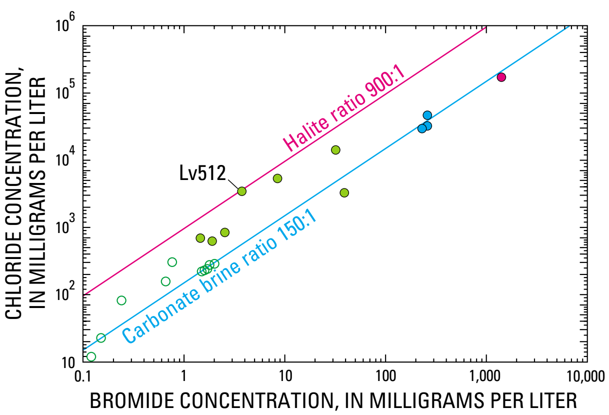


Figure 21. Specific conductance of water from northern end of Genesee Valley, Livingston County, N.Y., April 1997.

operations could migrate through bedrock fractures to wells near the valley walls, but the highest salinity values were at the center of the bedrock valley and, therefore, probably did not result from road salting. Although the source of halite-bearing water in the lower aquifer remains uncertain, the inflow of this water probably resulted from change in hydraulic gradient that followed the mine collapse.

The potential for intrusion of saline water also is a concern near the collapse area, where the rubble zone forms a permeable pathway between the flooded mine and the lower aquifer. The flooded mine is filled with brine, which is heavier than freshwater and has a specific gravity ($\gamma_{brine}/\gamma_{fw}$) of about 1.2. The brine will be forced upward through the rubble zone as the mine cavity closes under increasing geostatic pressure and could eventually enter the lower aquifer in the collapse area; it also might flow laterally through fracture zones in the Onondaga and Bertie Limestones. The level (altitude) of brine is being monitored through a water-level comparison of a well that penetrates the mine (Lv396) with one that is screened in the Onondaga and Bertie Limestones in the collapse area (Lv309; Akzo Nobel Salt, Inc., 1996). Records indicate that the brine level in the collapse area was over 120 ft below the bottom of the Bertie Limestone in 1998 (fig. 23).



EXPLANATION

Lower aquifer with specific conductance

○ Less than 2,500 microsiemens per centimeter—Group 2

● Greater than 2,500 microsiemens per centimeter—Group 3

● **Onondaga and Bertie Limestones**

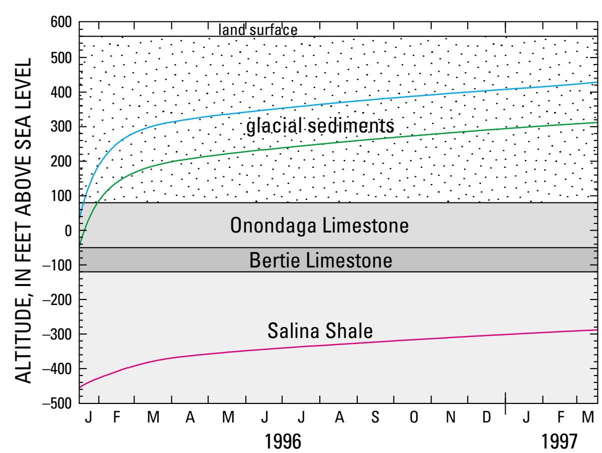
● **Bass Island carbonate brine**

Figure 22. Relation between chloride and bromide concentrations in water from the lower aquifer and the Onondaga and Bertie Limestones in Livingston County, N.Y., and the Bass Islands carbonate brine in western New York.

Natural Gas Exsolution

Methane and hydrogen sulfide were detected in the mine after the collapses and accumulated to unsafe levels periodically throughout the summer of 1994. Five or six wells, originally drilled as part of a grouting program to save the mine, were used to vent the gas from the mine, and three of the wells were flared in May 1995. The gas probably originated from a fracture zone in the Bertie Limestone, where gas had previously been detected in other boreholes (L. D. Milliken, ANSI, oral commun., 1996). The methane-to-ethane ratio in gas samples from three wells screened in the rubble zone in the collapse area ranged from 5 to 21, which indicates that the gas was produced primarily by thermogenic decomposition of organic matter (John C. Fountain, Dept. of Geology, State University of New York at Buffalo, written commun., 1995).

Some water wells screened in the middle aquifer as far as 6 mi from the collapse area began to produce gas after the collapse. Gas was flared from two wells screened near bedrock—Lv398 on the west side of the



EXPLANATION

Estimated brine level in collapse area (well Lv 309)

Brine level in mine (well Lv 396)

Water level in collapse area (well Lv 309)

Figure 23. Measured brine level (altitude) in the mine, estimated brine levels in the collapse area (location II), and water levels in the collapse area, Livingston County, N.Y., January 1996 through March 1997.

valley in Cuylerville, and Lv428 on the east side of the valley in Groveland (fig. 17), after the wells went dry in April and July 1994, respectively. Gas also was produced in well Lv91 screened in the lower aquifer in Mt. Morris beginning in December 1994. The methane-to-ethane ratio in gas samples from these three wells ranged from 54 to 135, indicating that the gas was produced primarily through biogenic decomposition of organic matter (J. C. Fountain, State University of New York, Buffalo, written commun., 1995).

Methane also was detected near a plugged access shaft at the northern end of the mine in January 1996, shortly after the mine had become completely flooded and ground-water levels had begun to recover. Methane emissions continued through the summer of 1996 but gradually diminished. They were apparently vented through bedrock outcrops above the edge of the mine and probably resulted from the increased pressure that developed within the mine in response to the rising ground-water levels (J. C. Fountain, State University of New York, Buffalo, written commun., 1996).

SIMULATION OF GROUND-WATER FLOW

Ground-water flow through the aquifer system in the Genesee Valley was simulated with three-dimensional models with MODFLOWP (Hill,1992) to represent flow conditions before and after the ceiling collapses in the Retsof salt mine. The models were based on the generalized geologic sections presented earlier and used homogeneous distributions of hydraulic properties and, therefore, gave a simplified representation of actual aquifer conditions.

The first model (3D-A, table 4) was calibrated to measured water levels and flow rates in steady- and transient-state simulations and was used to calculate ground-water budgets and determine the time required for water levels to return to precollapse conditions. Results of transient-state simulations were used to assess the effects of the reversal of ground-water flow directions on the location and rate of saline-water intrusion into the lower aquifer in the northern part of the valley. Model 3D-A was later modified to create model 3D-B, which was used to assess the effects of gas exsolution on the volume of water released from storage within the confined aquifers.

One-dimensional models (1D-A and 1D-B, table 4) with MODFLOW (McDonald and Harbaugh, 1988) with the interbed-storage package (Leake and Prudic, 1991) were then constructed to simulate compression of fine-grained sediments in the upper and lower confining layers, as described in the section "Simulation of land subsidence." Estimated values for hydraulic properties of the confining layers obtained through calibration of the one-dimensional models then were incorporated in three-dimensional model 3D-A to create model 3D-C, which was used to depict the spatial distribution of land subsidence. Four different assumptions were made concerning the storage properties of confined aquifers and confining-layer sediments during calibration of the three-dimensional and one-dimensional models. A summary of the five models used, and the assumptions made concerning storage properties and applications of each model, is given in table 4.

Model Design

Three-dimensional model 3D-A was prepared with MODFLOWP , a computer program based on MODFLOW, which uses a nonlinear-regression method to estimate model parameter values. The modeled area extends 30 mi from Avon southward to the Valley Heads Moraine near Dansville and represents an area of 61.3 mi^2 (fig. 2). The modeled area generally coincides with the boundaries of the upper aquifer but also includes the Fowlerville Moraine, where the upper aquifer is extremely narrow.

Model Layers and Grid

The three aquifers and two confining layers that form the aquifer system (fig. 10) were represented by five model layers. Water-bearing zones in the bedrock were not represented in the model because little information on their hydraulic properties is available, and to include them would require a substantial increase in the size of the modeled area.

Model layer 1 represents the upper aquifer, which includes three components—the alluvial sediments that cover most of the valley floor, the deltaic and kame deposits along parts of the valley walls, and the Fowlerville Moraine and adjacent fine-grained lacustrine sediments (fig. 24). Model layer 2 represents the upper confining layer (lacustrine sediments and till), which separates the upper aquifer from the middle aquifer; layer 2 also includes the lower part of the Fowlerville Moraine. Model layers 3 and 5 represent the two confined aquifers, and model layer 4 represents the lower confining layer, which separates them. Model layers 3 and 4 include delta and kame deposits that are assumed to extend from the confined aquifers to land surface. Model layer 5 implicitly represents fractures at the bedrock surface that are in hydraulic connection with the lower aquifer; the transmissivity of model layer 5 estimated in model calibration includes the contribution from bedrock fractures.

Each of the five model layers was divided into a uniformly spaced grid of 300 ft with 470 rows and 194 columns. Model layers 1 and 2 (upper aquifer and upper confining layer) contain 19,005 active cells and represent a wider area (61.3 mi^2) than the other model layers because the bedrock valley narrows with depth (fig. 10). Layer 5 contains 17,024 active cells and represents an area of 55 mi^2 and its horizontal extent (fig. 25*B*) coincides with the assumed horizontal extent of the lower confining layer.

The thickness of sediments represented by each cell in model layers 1 and 2 was calculated from isopach (thickness) maps of the upper aquifer and upper confining layer, respectively. The thickness of morainal and lacustrine sediments in the vicinity of the Fowlerville moraine was divided equally between model layers 1 and 2. The middle and lower aquifers were assumed to be of uniform thickness (10 ft throughout model layer 3 and 25 ft throughout model layer 5, respectively), except where these aquifers pinch out near the bedrock valley walls. The thickness of the lower confining layer (model layer 4) was computed as

Table 4. Summary of models of ground-water flow and land subsidence in Genesee Valley, Livingston County, N.Y., assumptions used in defining storage coefficients, and the applications of each

[Locations of seismic-survey monument K10 and location II are shown in fig.17]

Model	Assumptions used in defining storage coefficients [a]	Model applications
Three-dimensional model		
3D-A	Constant specific storage S_s in confined aquifers: $S_s = \rho g \alpha$	Water-level drawdown and recovery Saline-water migration
3D-B	Variable specific storage S_s' in confined aquifers: $S_s' = S_s + S_{sg}$	Gas exsolution from confined aquifers
3D-C	Elastic specific storage[b] in confining layers: $\overline{S}_s = b^{-1}\Sigma S_{se(i)} b_i$, $i = 1, b$	Spatial distribution of land subsidence
One-dimensional models		
1D-A: upper confining layer at location II	Elastic specific storage S_{se} and inelastic specific storage S_{sin} in confining layers: $S_{se(i)} = \rho g m / d_i$ $S_{sin(i)} = S_{se(i)} * 30$	Pore pressure at location II
1D-B: upper and lower confining layers at monument K10	Same as in model 1D-A	Land subsidence at monument K10

[a]Variables:

ρg = specific weight of water $[ML^{-2}T^{-2}]$,
α = compressibility of aquifer material $[ML^{-1}T^{-2}]^{-1}$,
S_{sg} = given in equation 18 $[L^{-1}]$,
m = a constant estimated by PEST regression $[MT^{-2}]$,
d_i = depth of cell i in models C and D [L],
b = thickness of confining layer [L], and
b_i = unit thickness of cell i in models C and D [L].

[b]Elastic storage $S_{se(i)}$ estimated from PEST regression (Doherty and others, 1994) neglecting inelastic compression.

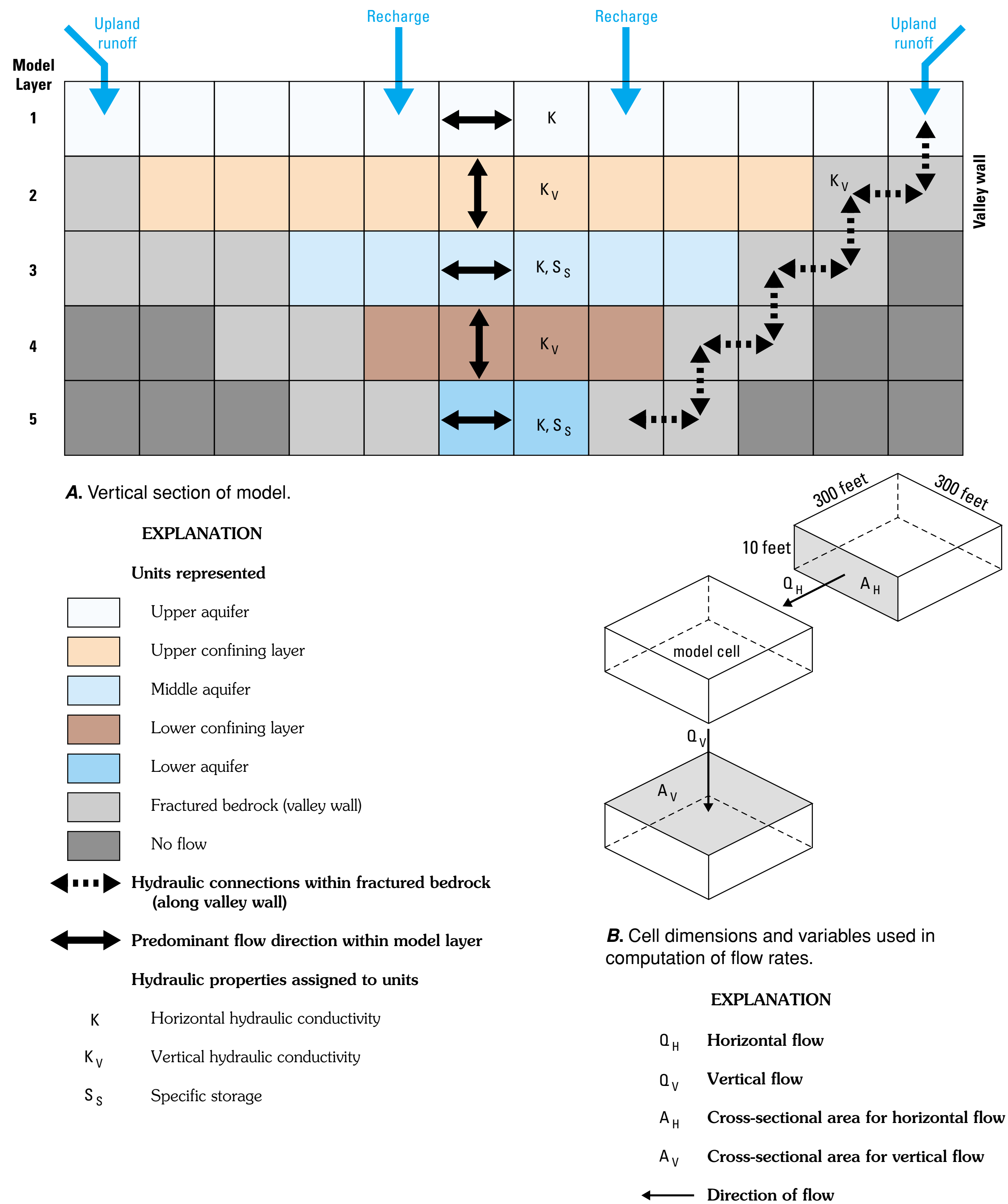
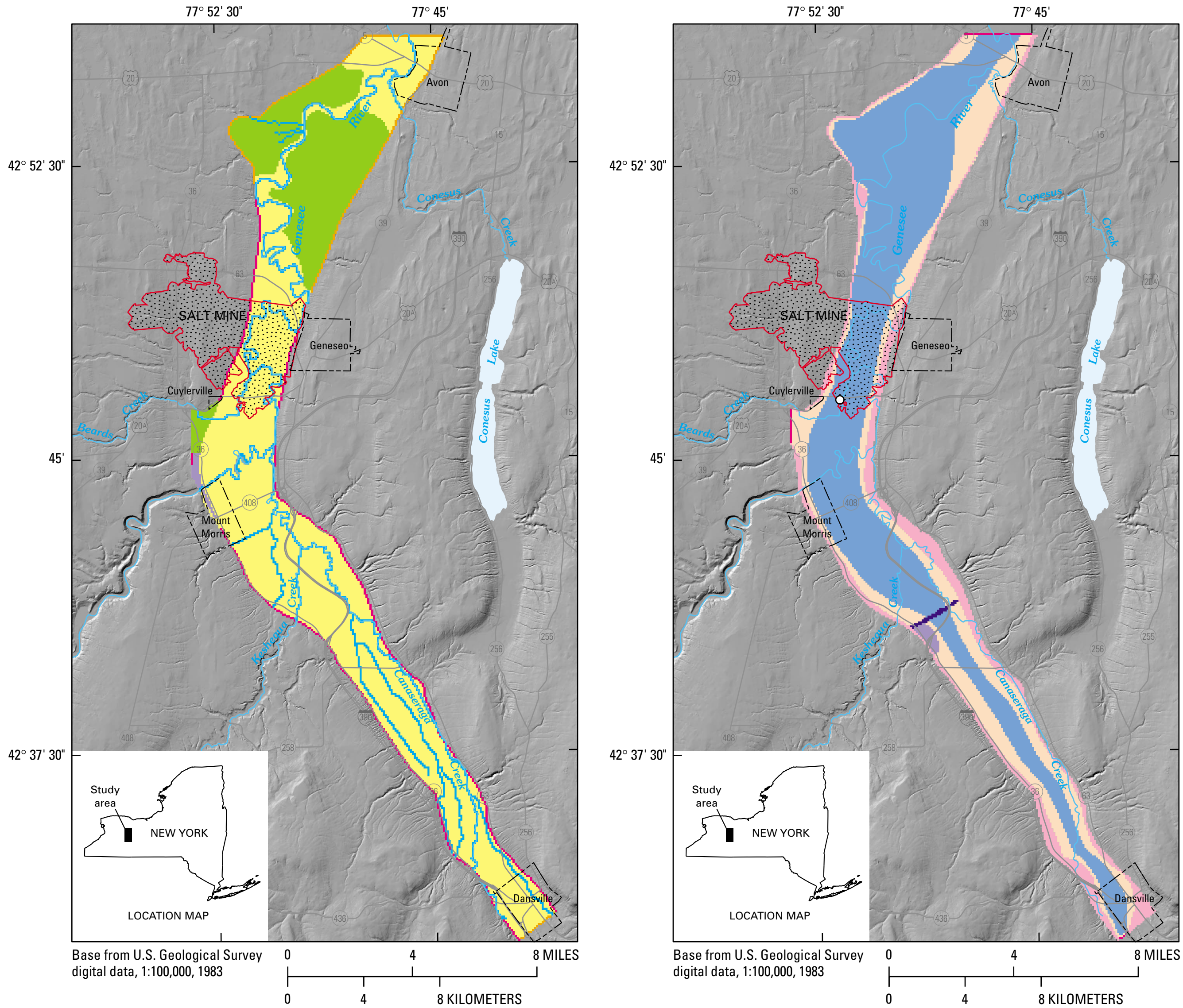


Figure 24. ***A***. Vertical section of model with hydraulic properties assigned to each layer and representation of vertical hydraulic connections between model layers along valley walls. ***B***. Cell dimensions and variables used in computation of flow rates between cells.



EXPLANATION

Model layer 1

- Alluvial deposits
- Fine-grained deposits
- Delta
- River and stream cells
- Head-dependendent boundary cells
- Upland runoff cells

Model layers 3, 4, and 5

- Lower aquifer
- Weathered bedrock in layer 3
- Weathered bedrock in layer 4
- Kame delta (layers 3 and 4)
- Ice-contact deposits (layers 3 and 4)
- Head-dependent boundary (layers 3 and 5)
- Collapse area (layer 5)

Figure 25. Boundary conditions in ground-water-flow model of aquifer system in Genesee Valley, Livingston County, N.Y.: ***A***. Upper aquifer (model layer 1). ***B***. Confined aquifer system (model layers 3, 4 and 5).

the difference between the altitude of the top of model layer 5 (altitude of bedrock surface plus thickness of model layer 5) and the bottom altitude of model layer 3 (altitude of land surface less combined thickness of model layers 1 through 3).

Boundary Conditions

Boundary conditions that represent inflows and outflows of water to and from the aquifer system were specified in the three model layers that correspond to the upper aquifer and the two confined aquifers (model layers 1, 3, 5).

Model Layer 1

A constant-flow boundary was used to represent direct recharge to the upper aquifer from precipitation; recharge rates for cells along the valley walls were increased to account for the additional recharge that enters the upper aquifer along the valley edges from channeled and unchanneled runoff from upland areas (fig. 25*A*). Head-dependent boundaries were used to represent other inflows from upland areas, which include: (1) the thick glacial sediments in the Valley Heads Moraine at the south end of the valley, (2) the thick lacustrine sediments near the Fowlerville Moraine south of Avon, and (3) the subcrop of the Onondaga Limestone near the Fowlerville Moraine. Head-dependent boundaries also were used to represent outflow through the northern (downgradient) boundary, and flow to and from perennial stream channels, including the Genesee River, Canaseraga Creek, and their tributaries.

Head values along head-dependent boundaries were specified from assumed water-table altitudes or from estimates of stream stage. Stream-stage estimates were computed from land-surface altitudes derived from digital elevation model (DEM, U.S. Geological Survey, 1996) data with 34-ft (10-m) resolution. The conductance (C) for model cells representing head-dependent boundaries was computed from the relation:

$$C = \frac{KA}{l}, \tag{5}$$

where K= hydraulic conductivity (LT^{-1});

A = cross-sectional flow area (L^2); and

l = length of flowpath between cell center and model boundary (L).

Hydraulic properties of cells representing lateral flows through head-dependent boundaries were assigned as follows:

K (hydraulic conductivity) was specified according to the type of material at the boundary;

A (cross-sectional area) was computed as the product of the cell width (w) and thickness (b); and

l (flowpath length) equaled w.

The K value for cells representing flows to and from perennial streams was assumed to be 1 ft/d, and the ratio A/l was based on the length, width, and thickness of the stream channel as follows:

cells representing the Genesee River: 3×10^4 ft;

cells representing named creeks: 1×10^4 ft; and

cell representing unnamed tributaries: 3×10^3 ft.

Model layers 3 and 5

Head-dependent boundaries were used in model layer 3 (middle aquifer) to represent inflow from upgradient areas at the Valley Heads Moraine and in the Beards Creek valley (fig. 25*B*); head-dependent boundaries were specified for model layers 3 and 5 (middle and lower aquifers) to represent underflow through the northern (downgradient) boundary. Drain boundaries were used in layer 5 to represent discharge from the lower aquifer to the rubble zone in the collapse area, as discussed in the section “Transient-state simulations.”

Model layers 2 through 5

No-flow lateral boundaries were used in model layers 2 through 5 to represent the contact between the aquifer system and the shale bedrock at the valley wall. Hydraulic connectivity in active cells near the perimeters of model layers 2 through 5 was increased to represent vertical leakage through permeable deposits, bedrock fractures, or both, along the valley wall (fig. 24*A*). The ratio of horizontal to vertical hydraulic conductivity specified for these cells was scaled by a factor to account for differences in the cross-sectional flow area that result from the cell geometry. The flow rate (Q) between adjacent model cells is given by

$$Q = KiA, \tag{6}$$

where i = hydraulic gradient between the cells.

The thickness of permeable deposits and(or) fractured bedrock along the valley walls was assumed to be 10 ft; as a result, the cross-sectional area for horizontal flow (A_H) was 3×10^3 ft^2, and the cross-sectional area for vertical flow (A_V) was 9×10^4 ft^2 (fig. 24*B*). The vertical hydraulic conductivity (K_V) was, therefore, reduced by a factor of 30 in cells representing leakage along the valley walls, such that vertical anisotropy in these cells equaled 1.0.

Model Calibration

Parameter values representing aquifer properties were adjusted through steady-state and transient-state simulations to produce a model that would approximate the water levels and flow rates measured (1) prior to the collapses, (2) during the flooding of the mine, and (3) during the period of water-level recovery after the mine was filled. The steady-state simulation represented precollapse conditions, when no large stresses were imposed on the aquifer system; the resulting hydraulic heads provided initial conditions for a 29-month (876-day) transient-state simulation (model 3D-A in table 4) that began with the first collapse in March 1994 and ended in August 1996, 8 months after the mine was flooded. The transient-state simulation represented the two postcollapse periods—drainage from the aquifer system to the mine (March 1994 to January 1996), and recovery of water levels after the mine had completely filled (January 1996 through August 1996). Model parameters were estimated through a nonlinear regression method (Cooley and Naff, 1990) that minimized differences between measured and computed heads and flows.

Calibration Procedure

Initial estimates were specified for 25 parameters in model A (table 5), 12 of which were estimated through nonlinear regression.

Table 5. Numbers of parameters representing Genesee Valley aquifer properties in model 3D-A, as estimated and fixed in nonlinear regression for Genesee Valley aquifer system, Livingston County, N.Y.

[—, no parameter specified]

Aquifer property	Model layer				
	1 Upper aquifer	2 Upper confining layer	3 Middle aquifer	4 Lower confining layer	5 Lower aquifer
Horizontal hydraulic conductivity:					
Estimated	3	—	1	—	1
Fixed	—	1	—	1	—
Vertical hydraulic conductivity:					
Estimated	—	—	—	2	—
Fixed	—[a]	1	—[a]	2	—
Specific storage:					
Estimated	—	—	1	—	1
Fixed	1	1	—	1	—
Recharge:					
Estimated	3	—	—	—	—
Other fixed properties:					
Streambed conductance	3	—	—	—	—
Conductance along valley walls[b]	—	1	—	—	—
Specific storage along valley walls[b]	—	1	—	—	—
Total					
Estimated: 12	6	0	2	2	2
Fixed: 13	4	5	0	4	0

[a]Vertical hydraulic conductivity value taken from underlying layer.

[b]Same value specified in model layers 2 through 5.

Steady-state simulations

Steady-state simulations were used to estimate values for six parameters that represent horizontal hydraulic conductivity (herein referred to as hydraulic conductivity) and recharge in the alluvial, deltaic, and fine-grained deposits in model layer 1 (upper aquifer) (fig. 25*A*). Hydraulic heads computed from the steady-state simulation were compared with the assumed water-table altitude at 29 locations chosen to represent the spatial distribution of head shown in figure 11, and ground-water discharges computed from the steady-state simulation were compared with base flows estimated from stream discharge measured during the spring or fall from 1973 through 1993 at USGS gaging stations on Canaseraga Creek and on the Genesee River at Jones Bridge Rd. and Avon (fig. 4). Base flow along two stream reaches was estimated from discharge measured during 15 five-day periods with steadily declining flow, when storm runoff and stream regulation were absent.

Improvements in model results were identified through comparison of the sum of squared errors (*SSE*) given by

$$SSE = \Sigma[w_i^{1/2} e_i]^2 \text{ , } i = 1, n \qquad (7)$$

where e_i = difference between the measured and calculated values of measurement i;

$w_i^{1/2}$ = square root of the weight assigned to the error in the measured value of measurement i,

$w_i^{1/2} e_i$ = weighted residual corresponding to measurement i; and

n = number of measurements;

and the standard error of estimate (*SE*) given by

$$SE = \left[\frac{SSE}{n-p}\right]^{1/2} , \qquad (8)$$

where p = the number of model parameters estimated by the regression.

The weights, w_i, were chosen according to procedures in Hill (1992) to account for the different units associated with head measurements (130 to 680 ft) and flow measurements (1.0×10^6 to 3.5×10^6 ft^3/d). The w_i values were adjusted such that all the head and flow measurements were weighted equally.

Transient-state simulations

The six parameter values obtained from the steady-state simulations were used in transient-state simulations to estimate values for six additional parameters representing hydraulic conductivity and specific storage of model layers 3 and 5 (middle and lower aquifers, respectively), and vertical hydraulic conductivity of model layer 4 (lower confining layer). These parameters were applied to all cells within these model layers except those representing the weathered bedrock (fig. 25*A*). A value also was estimated for a separate parameter representing vertical hydraulic conductivity of the lower confining layer at two model cells in the collapse area to represent the increase in permeability that probably resulted from subsidence of sediments within the confining layer above the rubble zone in the bedrock. Results of transient-state simulations were compared with two estimates of ground-water discharge to the mine (March and September 1994), and with 354 water-level measurements made at 51 wells from March 1994 through August 1996. Of these 51 wells, 24 were screened in the lower aquifer, 26 were screened in the middle aquifer, and 1 was screened in the upper aquifer.

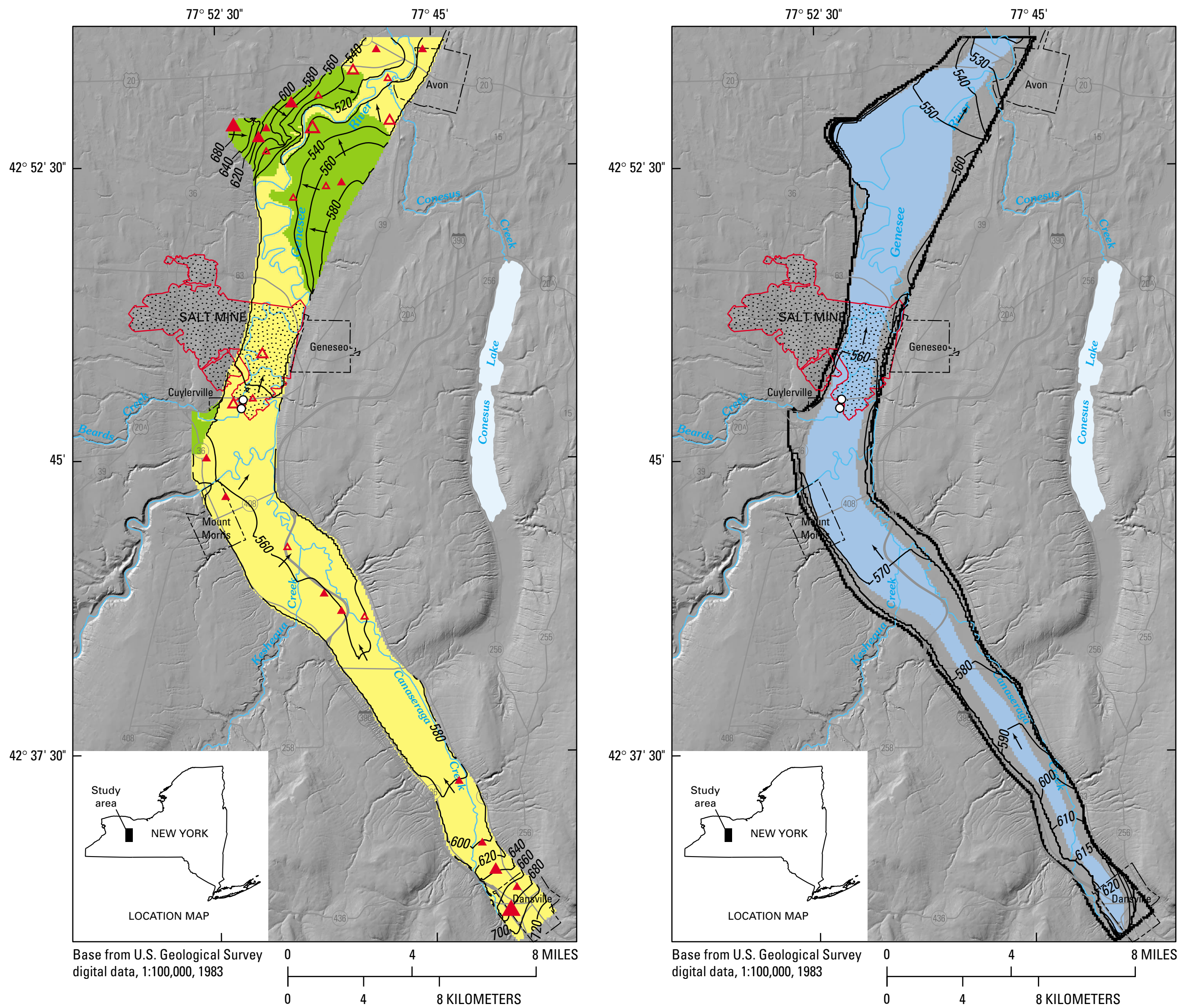
Additional steady-state simulations were then conducted with the six parameter values from the transient-state simulations to obtain new estimates for the six parameters used in model layer 1. These new estimates then were used in additional transient-state simulations to obtain a final set of six parameters values for model layers 3, 4 and 5.

Precollapse Conditions

The base flows and hydraulic-head distribution in the water-table aquifer computed with the steady-state simulation matched the measured heads and flows reasonably well, and the predicted directions of ground-water flow were consistent with the directions assumed for precollapse conditions.

Head distribution

The hydraulic-head distribution in the upper aquifer, as computed by the steady-state simulation with model 3D-A (fig. 26*A*) was similar to assumed water-table elevations prior to the collapses (fig. 11). The model overpredicted heads by 10 to 20 ft in areas where fine-grained sediments are present near the Fowlerville Moraine, but the simulated directions of ground-water flow were consistent with flow directions



A. Upper aquifer (model layer 1).

B. Lower aquifer (model layer 5).

EXPLANATION

Model layer 1

Alluvial deposits

Fine-grained deposits

560 **Water-table contour—** Contour interval 20 feet. Datum is sea level. Arrow indicates direction of ground-water flow

Collapse area

Difference between predicted and observed head, in feet

Less than –20

–10 to –20

0 to –10

0 to 10

10 to 20

Greater than 20

EXPLANATION

Model layer 5

Lower aquifer

Boundary of model area

560 **Potentiometric contour—** Contour interval 10 feet. Datum is sea level. Arrow indicates direction of ground-water flow

Collapse area

Figure 26. Distribution of hydraulic head and weighted residuals computed by steady-state simulation (model 3D–A) for Genesee Valley, Livingston County, N.Y. ***A.*** Upper aquifer (model layer 1). ***B.*** Lower aquifer (model layer 5).

assumed for precollapse conditions. The standard error in heads was 8.7 ft, and the range in computed heads was 520 to 720 ft. The hydraulic-head distribution computed for the lower aquifer indicated northward flow from the Valley Heads Moraine to the northern model boundary (fig. 26*B*). The horizontal hydraulic gradient computed in the steady-state simulation was relatively flat (3×10^{-4} ft/ft) throughout most of the lower aquifer, but was steeper (2×10^{-3} ft/ft) near the model boundaries and was generally oriented outward, which would indicate upward discharge through the assumed permeable zone along the perimeter of the valley.

Recharge and discharge

More than half of the computed inflow to the aquifer system consisted of recharge from precipitation on the valley floor (51 percent); another 5 percent was derived from upland runoff (table 6). More than one-third of the inflow to the aquifer system consisted of infiltration from the Genesee River and Canaseraga Creek; the remainder (9 percent) was contributed as underflow from upgradient boundaries (Valley Heads and Fowlerville Moraines, Onondaga Limestone subcrop, and Beards Creek Valley). Almost all of the computed outflow (98 percent) was discharged to the Genesee River and Canaseraga Creek. The computed base flow (outflow minus inflow) of Canaseraga Creek (45 ft^3/s or 3.9×10^6 ft^3/d) was 12 percent greater than the estimated baseflow (40 ft^3/s), and the computed base flow to the Genesee River (31 ft^3/s or 2.7×10^6 ft^3/d) was 25 percent less than the estimated base flow in this reach (40 ft^3/s).

Recharge through alluvial and deltaic sediments was estimated to be 20 in/yr (table 7), 50 percent greater than the estimate of 14 in/yr obtained from average annual runoff (Lyford and Cohen, 1988, fig. 3). The estimated rate of recharge through the fine-grained sediments was 3.9 in/yr. The recharge rate specified in model cells representing upland runoff was twice the recharge rate specified for alluvial and deltaic sediments and was equivalent to 0.2 ft^3/s per mile of valley wall—slightly smaller than the rate calculated by MacNish and Randall (1982) for recharge from unchanneled upland runoff in similar terrane immediately south of the Genesee watershed. The estimated rate of recharge from upland runoff (0.8 in/yr derived from 90 mi^2 of upland area) is a relatively small fraction of the average precipitation (30 in/yr), and rates exceeding 1.3 ft^3/s per mile of valley wall have been estimated for other areas in the glaciated northeast (Morrissey and others, 1988). The actual rate of direct recharge from precipitation probably is less than the estimated rate, and the actual rate of recharge from upland runoff probably is greater than the estimated rate, primarily because not all of the channeled runoff is explicitly represented in the model.

Aquifer properties

Estimated hydraulic conductivity values of the alluvial sediments was 280 ft/d, that for the deltaic sediments was 500 ft/d, and that for the fine-grained sediments was 0.93 ft/d . The estimated hydraulic

Table 6. Simulated water budget for Genesee Valley aquifer system, Livingston County, N.Y., under average steady-state conditions before collapses of Retsof mine ceiling in March and April 1995

[Flow rates are in millions of cubic feet per day.]

Inflow			Outflow		
Source	Rate	Percentage of total	Location	Rate	Percentage of total
Recharge from precipitation	5.30	51			
Upland runoff	.52	5			
Genesee River	2.60	25	Genesee River	5.30	51
Canaseraga Creek	1.04	10	Canaseraga Creek	4.89	47
Underflow	.94	9	Underflow	.21	2
TOTAL	10.4	100		10.4	100

Table 7. Parameter values estimated for model layer 1 (upper aquifer in Genesee Valley aquifer system, Livingston County, N.Y.), through nonlinear regression; also model error in steady-state simulation 3D-A and measured mean stream discharges

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

Variable	Value
Aquifer property	
Recharge, inches per year	
alluvial and deltaic sediments	20
fine-grained sediments	3.9
upland runoff	0.8[a]
Hydraulic conductivity, feet per day	
alluvial sediments	280
fine-grained sediments	0.93
deltaic sediments	500
Model error	
Sum of squared errors, feet squared	2,351
Standard error in heads, feet	8.7
Mean stream discharge, ft^3/s	
Genesee River (40 ft^3/s measured)	31
Canaseraga Creek (40 ft^3/s measured)	45

[a]Recharge derived from 90 square miles of upland area.

conductivity value for alluvial sediments is reasonable for coarse-grained sediments but seems too large for the alluvial sediments that contain fine sand and silt. A lower value would decrease the simulated base flow, however, and would increase model error. Optimum values were obtained by the nonlinear regression for the recharge rate and hydraulic conductivity of the alluvial sediments. The approximate individual confidence intervals (at the 95-percent confidence level) for these values were 3 to 36 in/yr for recharge and 8 to 380 ft/d for hydraulic conductivity; these wide ranges indicate that the parameter values were not estimated precisely. The confidence intervals are approximate because the required assumptions on which they are based (an unbiased model that is effectively linear with normally distributed weighted residuals) are not strictly met by the steady-state simulation. The large uncertainty in parameter estimates probably results from the relatively small number of measured heads and flows for comparison with simulated values, and from the low sensitivity of the steady-state simulation to these measurements as a result of the extensive boundaries specified in model layer 1.

Despite the uncertainty in estimated values, the hydraulic-head distribution in the upper aquifer, and the base flows computed with the steady-state simulation, are reasonable; therefore, the head distributions computed for the lower model layers probably represent valid initial conditions for the transient-state simulation. Sensitivity analyses indicate that none of the model parameters obtained through steady-state simulations significantly affected the results of transient-state simulations.

Postcollapse Conditions (Flooding and Water-Level Recovery)

Transient-state simulations of postcollapse conditions included additional boundary conditions in model layer 5 (lower aquifer) for the first simulation period, which represented ground-water discharge into the mine from the overlying aquifer system. Head-dependent boundaries were specified in two model cells—one above each of the two rubble zones associated with the two collapses. The first of these additional boundaries, which represented the collapse over panel 2-Yard South (fig. 15), was active at the start of the first simulation period (March 1994); the second boundary became active 1 month later to represent the collapse over panel 11-Yard West (April 1994). These boundaries were assigned heads equal to the aquifer-bottom (bedrock surface) altitude and arbitrary conductance (C) values (eq. 5) large enough that flow through the boundaries would be limited by the hydraulic conductivity of the lower aquifer, rather than that of the rubble. These boundary conditions were removed at the start of the second simulation period, which represented recovery of ground-water levels once the mine was filled.

The boundaries are considered reasonable representations of actual conditions during flooding because the mine was at atmospheric pressure, and the hydraulic gradient controlling discharge from the lower aquifer depended solely on the head in the aquifer. The hydraulic conductivity of the rubble zone connecting the aquifer to the mine is unknown but is assumed to be larger than that of the lower aquifer. Partial dewatering of the lower aquifer could have occurred above the rubble zone and formed a seepage face that, in effect, decreased the saturated thickness of the aquifer. Butler (1957) investigated the effects of seepage faces on ground-water flow and found that the saturated thickness could be decreased by as much as 50 percent at such boundaries. Decreasing the boundary conductance by 50 percent increased *SSE* more than 100 percent, however, and decreased the simulated discharges into the salt mine to less than 40 percent of those measured; therefore, the saturated thickness over the collapse area was not appreciably decreased in transient-state simulations.

The middle aquifer was assumed to remain under confined conditions throughout the transient-state simulations, although unconfined conditions were measured in parts of this aquifer in the Beards Creek Valley and near the Fowlerville Moraine. The transition from confined to unconfined conditions was not represented in the model because the altitude of the top of the middle aquifer could not be estimated precisely. As a result, the storage coefficent specified for the middle aquifer is too low in these areas because more water is released from storage under unconfined conditions than under confined conditions. The additional volume of water released is probably insignificant, however, because these areas represent only a small part of the total aquifer area.

Drawdowns

The distribution of drawdowns in the lower aquifer in January 1996 as computed by transient-state simulation with model 3D-A (fig. 27) is similar to the measured distribution (fig. 18). Computed drawdowns in wells near the collapse area exceeded the actual 400-ft drawdowns by less than 10 ft and are about 60 ft less than the actual 110-ft drawdowns in wells 7 mi to the south. Computed drawdowns also are generally about 15 ft less than the actual 50-ft drawdowns in wells 8 mi north of collapse area, although they are underpredicted in some wells in this area and overpredicted in others. The standard error in heads computed for the confined aquifers was 33 ft, and the range in heads was 100 to 735 ft.

Locations II and IV. The simulated change in drawdown with time is in close agreement with drawdowns measured at wells near the collapse area (location II). Computed and measured drawdowns in both the middle and lower aquifers match particularly well at locations II and IV (figs. 28 *A, C*). The maximum recorded drawdown at each of these locations is accurately simulated, but no data are available for comparison with drawdowns that occurred soon after the collapse because the wells were not completed until 1 to 2 years thereafter. Water levels in well Lv368 (fig. 28*A*) recovered more slowly than simulated early in the recovery period, perhaps because natural gas in the lower aquifer temporarily dewatered the aquifer at this location.

Location III. The model overpredicted drawdowns at location III (fig. 28*B*) and provided a particularly poor match in the middle aquifer (where measured drawdowns were relatively small) given its proximity to the collapse area. This disparity, and the differences in the isotopic content between waters near location III and from the rest of the confined-aquifer system, indicate that the confined aquifers at this location are hydraulically isolated. Stratigraphic information to document this lack of continuity is unavailable, however; therefore, no changes were made in the model design to decrease the error in computed drawdowns at this location. For this reason, measured drawdowns at well Lv351 (location III, middle aquifer) were not included in the nonlinear regression.

Location V. Computed drawdowns at location V, 7 mi south of the collapse area (fig. 28*D*) were underpredicted by 80 ft in the middle aquifer and 60 ft in the lower aquifer. The drawdown distribution at this location is unusual in that measured drawdowns in the middle aquifer (135 ft) were 25 ft greater than in the lower aquifer (110 ft). This indicates that (1) the two confined aquifers are hydraulically connected near, or somewhere north of, this location, and (2) the hydraulic conductivity and (or) thickness and width of the lower aquifer decreases south or east of the area of hydraulic connection and thereby causes steeper gradients than in the middle aquifer. The kame deposit mapped near Sonyea (fig. 8) indicates that the ice margin temporarily halted there during a recession; if a large thickness of coarse-grained ice-contact material accumulated nearby during the deposition of the lower confining layer, this material would now form a hydraulic connection between the confined aquifers. The vertical hydraulic conductivity of the lower confining layer, which separates the middle aquifer from the lower aquifer, was increased locally by a factor of 10 in the model to represent the assumed hydraulic connection (fig. 25*B*). The computed drawdowns in the two confined aquifers were nearly equal as a result of this change in model design but were still less than the measured drawdowns.

Location I. The model variably underpredicted and overpredicted drawdowns at the site 8 mi north of the collapse area, near the Fowlerville moraine. The match for wells Lv483 and Lv496 (screened in the middle and lower aquifers near this location, respectively) was reasonable (fig. 28*E*). Drawdowns in the

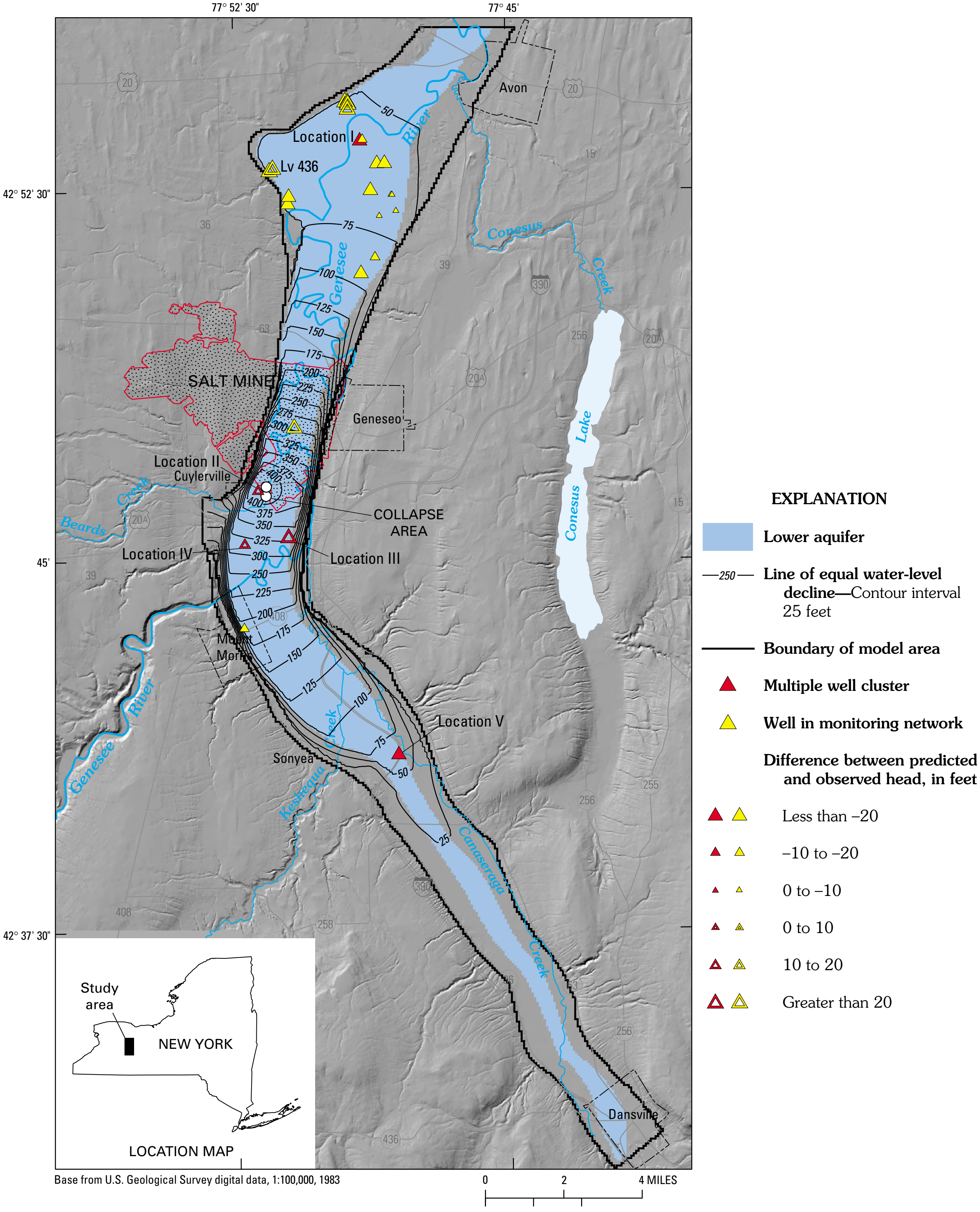
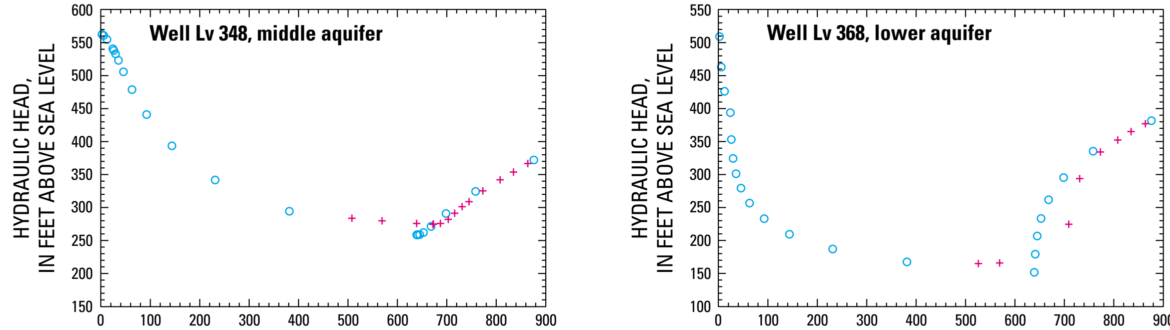
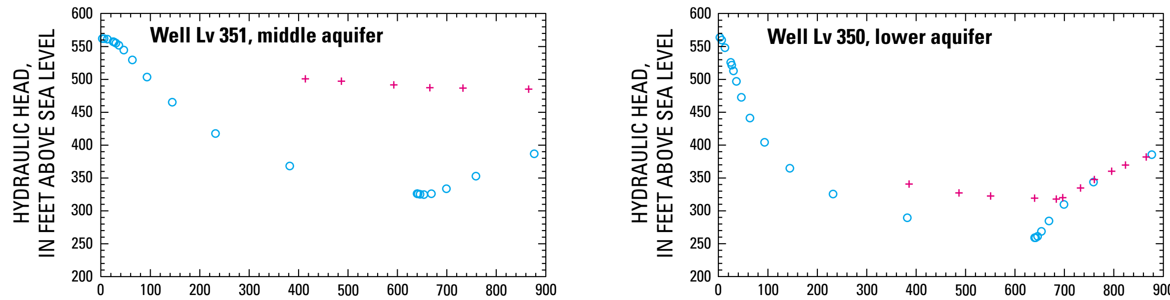


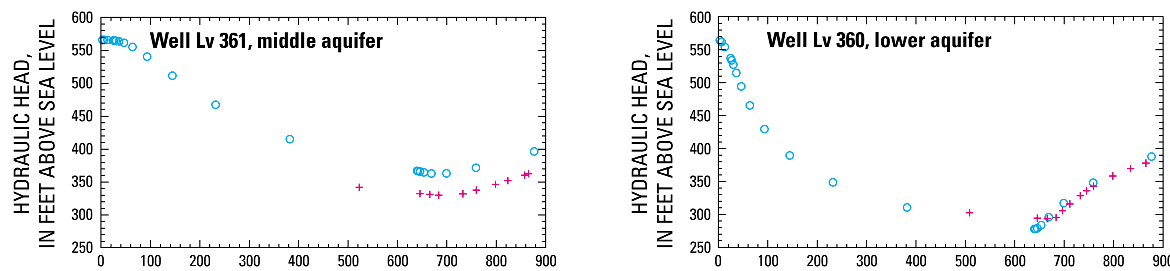
Figure 27. Distribution of drawdown in lower aquifer in Genesee Valley, Livingston County, N.Y. in January 1996, and range of weighted residuals computed with values from transient-state simulation (model 3D-A).



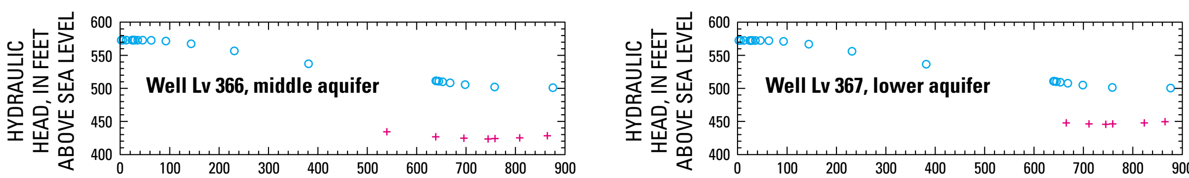
A. Location II, wells Lv348 and Lv368.



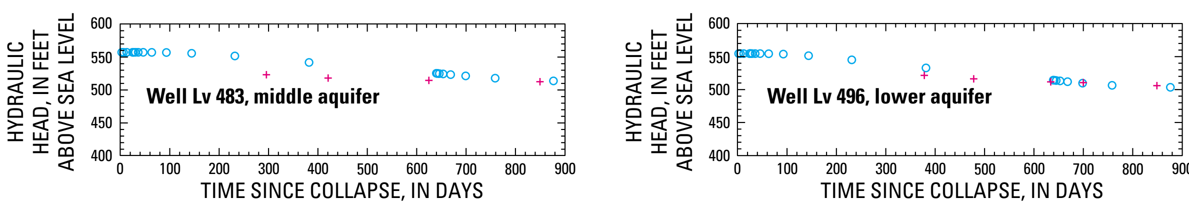
B. Location III, wells Lv351 and Lv350.



C. Location IV, wells Lv361 and Lv 360.



D. Location V, wells Lv366 and Lv367.



E. Site near Fowlerville moraine, wells Lv483 and Lv496.

EXPLANATION

Hydraulic head, in feet above sea level

+ Measured

○ Simulated

Figure 28. Water levels measured in middle and lower aquifers in Genesee Valley, Livingston County, N.Y. during 29 months after collapses at Retsof salt mine, and values simulated by transient-state simulation (model 3D-A): ***A.*** Location II, wells Lv348 and Lv368. ***B.*** Location III, wells Lv351 and Lv350. ***C.*** Location IV, wells Lv361 and Lv360. ***D.*** Location V, wells Lv366 and Lv367. ***E.*** Site near Fowlerville moraine, wells Lv483 and Lv496. (Well locations are shown in figs. 4 and 18.)

lower aquifer near the center of the valley and south of the Fowlerville moraine were generally underpredicted by 20 to 30 ft, whereas those on the west side of the valley near the hamlet of Fowlerville were overpredicted by 30 to 50 ft. The glacial stratigraphy in this area is poorly defined, and the simplified representation of the aquifer system in the model clearly does not accurately represent the distribution of drawdown in this part of the Genesee Valley.

Water budget

The computed water budget for the entire aquifer system during the first simulation period (March 1994 to January 1996), which represented the period of drainage from the aquifer to the mine (table 8), was similar to that computed by the steady-state (precollapse) simulations (table 6). Ground water released from storage provided about 11 percent of the total inflow in the water budget for the entire aquifer system and provided most of the water that discharged to the mine. This result is consistent with measured tritium concentrations at location II (near the collapse area), which indicated the ground water is more than 40 years old, because ground-water age at location II would be younger (and tritium concentrations larger) if most of the water that discharged to the mine were derived from recent recharge. The computed water budget for the confined part of the aquifer system (model layers 2 through 5) indicates that most of the inflow (58 percent) was derived from storage in the lower aquifer, and that 15 percent was derived from storage from the rest of the confined system; another 24 percent of the total inflow was derived from vertical leakage to the confined-aquifer system from the upper aquifer along the valley walls and through deltaic deposits.

Table 8. Simulated water budget for aquifer system in Genesee Valley, Livingston County, N.Y., during period of mine flooding, March 1994 through January 1996. A. Entire aquifer system. B. Confined-aquifer system

[Flow volumes are in millions of cubic feet; <, less than]

Inflow			Discharge		
Source	Volume	Percentage of total	Location	Volume	Percentage of total
A. Entire aquifer system (model layers 1 through 5)					
Storage	840	11	River and streams	6,390	84
Recharge and upland runoff	3,720	49	Underflow	190	3
River and streams	2,430	32	Salt mine	1,020	13
Underflow	610	8			
TOTAL	7,600	100	TOTAL	7,600	100
B. Confined part of aquifer system (model layers 2 through 5)					
Storage					
Upper confining layer	7	< 1			
Middle aquifer	66	6			
Lower confining layer	90	9			
Lower aquifer	614	58			
Underflow			Underflow		
Middle aquifer	20	2	Middle aquifer	1	< 1
Lower aquifer	< 1	< 1	Lower aquifer	23	3
Vertical leakage:			Vertical leakage:		
Valley walls	203	19	Valley walls	6	< 1
Deltaic deposits	50	5	Salt mine	1,020	97
TOTAL	1,050	100	TOTAL	1,050	100

Nearly all (97 percent) of the water discharged from the confined-aquifer system flowed into the salt mine. Computed discharge to the mine in March 1994 (20 ft^3/s) was nearly twice the estimated discharge (11 ft^3/s), and the computed discharge for September 1994 (28 ft^3/s) was little more than half the estimated discharge (45 ft^3/s). The estimated volume of water discharged to the mine was 1.1×10^9 ft^3, which represents 55 to 60 percent of the mine volume, as calculated from the tonnage of salt removed (1.34×10^8 tons) and an extraction ratio of 0.66 (Henry Klugh, oral commun., Akzo Nobel Salt, Inc., 1998). An alternative estimate of the extraction ratio (0.55) based on maps of the mined area indicates that the volume of ground-water discharge is equal to about 70 percent of the mine volume (Richard Young, oral commun., Geological Sciences, State University of New York-Geneseo, 1997). The underprediction of discharge to the mine in September 1994 indicates that drawdowns in the aquifer system were generally greater than simulated, particulary near location V at Sonyea, where they were underpredicted by 60 to 80 ft.

Estimates of aquifer properties

The parameter values estimated by nonlinear regression for the six aquifer properties are listed in table 9, together with their approximate individual confidence intervals and coefficients of variation. The parameter values were log-transformed to ensure that the values remained positive in the regression. The confidence intervals were computed based on the method described in Hill (1992) under the assumption that the model is correct and linear in the vicinity of the optimum set of values, and that the parameter values are normally distributed. The reported statistics are only

Table 9. Optimum parameter values estimated for confined aquifer system (model layers 2 through 5) in Genessee Valley, Livingston County, N.Y., through nonlinear regression in transient-state simulation (model 3D-A), and their approximate confidence intervals at 95-percent level, and discharge to mine estimated from volume of flooded area in March and September 1994.

[ft/d, feet per day, ft^2, feet squared; ft^{-1}, per foot; ft^3/s, cubic feet per second.].

Variable	Value	Approximate individual confidence interval	Coefficient of variation (percent)[a]
Aquifer property			
Hydraulic conductivity, ft/d			
Middle aquifer	3.7	1.3 - 11	30
Lower aquifer	300	180 - 480	26
Vertical hydraulic conductivity of lower confining layer, ft/d			
Collapse area	2.7×10^{-2}	2.3×10^{-4} - 3.1	37
Remainder of layer	1.2×10^{-3}	4.1×10^{-4} - 3.2×10^{-3}	4
Specific storage, ft^{-1}			
Middle aquifer	6.9×10^{-5}	1.3×10^{-5} - 3.6×10^{-4}	5
Lower aquifer	2.9×10^{-4}	1.4×10^{-4} - 6.2×10^{-4}	3
Model error			
Sum of squared errors, ft^2	4.04×10^5		
Standard error in heads, ft	32.9		
Discharge to mine, ft^3/s	Simulated	Estimated from volume of flooded area	
March 1994	20	11	
September 1994	28	45	

[a]Coefficient of variation on log-transformed parameter.

approximate, however, because the model does not meet these assumptions, but they indicate qualitatively the relative reliability of the optimum parameter values.

Hydraulic conductivity. Hydraulic conductivity (K) of the lower aquifer was estimated to be 300 ft/d, which is about 60 percent of the value estimated from specific-capacity data from well Lv91 in Mt Morris, and within the range of values reported for sand and gravel (Freeze and Cherry, 1979). The estimated hydraulic conductivity of the middle aquifer was about 4 ft/d, which is equivalent to values commonly reported for silty sand. The coefficients of variation for the estimated log K-values (26 percent and 30 percent, respectively) indicate that the model was sensitive to these parameters and that changing the optimum values would increase model error.

The estimated transmissivity of the lower aquifer is 7,500 ft^2/d, of which 15 percent (1,100 ft^2/d) could result from fractures at the bedrock surface where the aquifer is underlain by the Onondaga Limestone. A separate regression was done in which the hydraulic conductivity of the lower aquifer was assumed to be 15 percent greater in areas overlying the subcrop area of the Onondaga Limestone than in the remainder of the valley. The addition of bedrock fracture transmissivity to the lower aquifer did not improve the model fit, however, and resulted in no significant change in model error or estimated parameter values.

The estimated vertical hydraulic conductivity (K_v) values for the lower confining layer were 2.7×10^{-2} ft/d in the collapse area and 1.2×10^{-3} ft/d for the remainder of the layer. The larger K_v value estimated for the collapse area is consistent with the hypothesis that the permeability of the confining layer was increased by subsidence, but the large confidence interval and coefficient of variation associated with the estimate indicates that a reliable value cannot be determined from the available data.

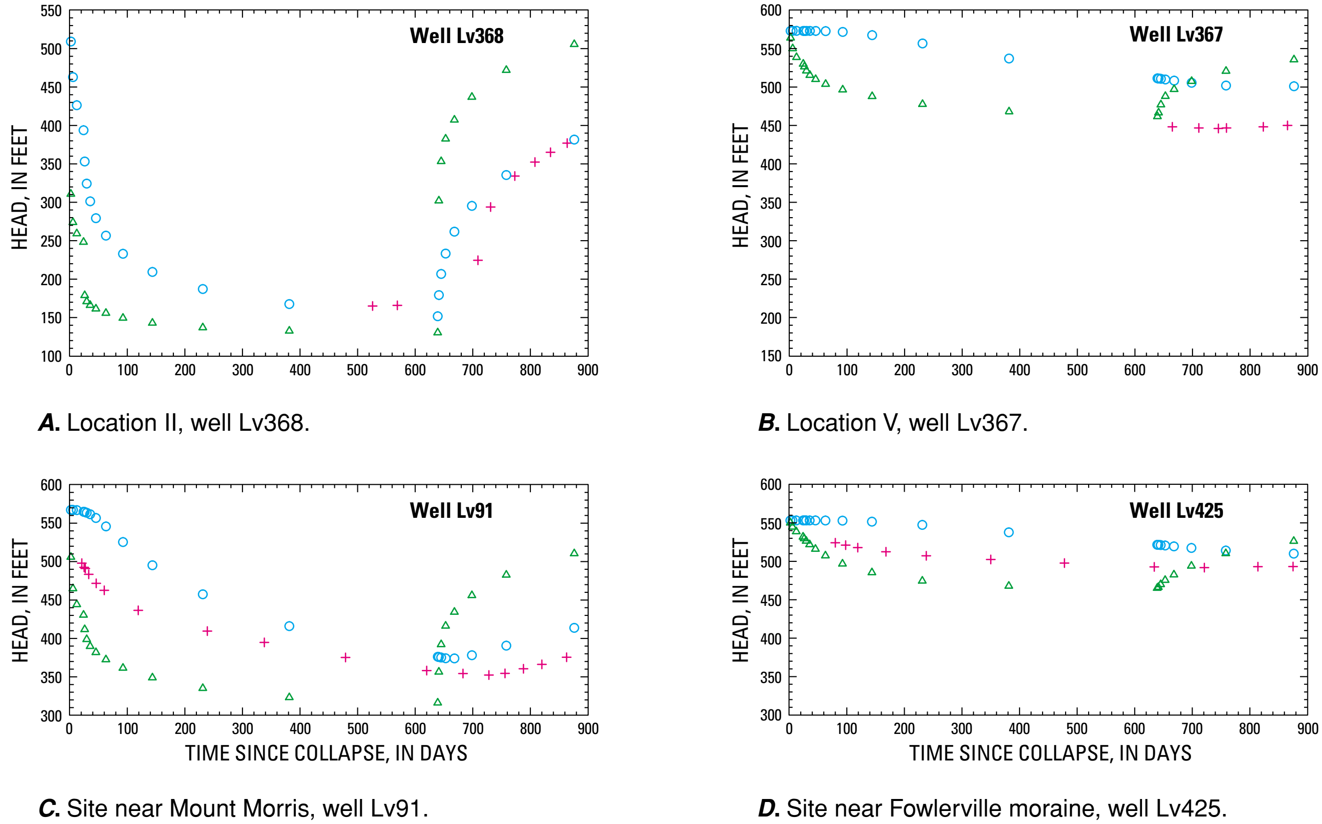
Specific storage. Specific storage S_s of a confined aquifer equals the volume of water released from storage per unit volume of aquifer under a unit decline in hydraulic head. Specific storage (S_s) values estimated for the lower and middle aquifers were 2.9×10^{-4} ft^{-1} and 6.9×10^{-5} ft^{-1}, respectively. Typical values of S_s corresponding to elastic compression of similar aquifers range from 1×10^{-6} ft^{-1} to 3×10^{-6} ft^{-1}, as indicated by extensometer measurements of land subsidence caused by pumping (Ireland and others, 1984; Riley, 1998). Most of these values were derived from alluvial basins in the southwestern United States; a similar value (2.3×10^{-6} ft^{-1}) was also calculated for a glacial sand and gravel aquifer in Alaska (Nelson, 1982). The coefficients of variation for the estimated log S_s values were extremely small (3 to 5 percent), and assigning the lower aquifer a lower specific storage value of 2.3×10^{-6} ft^{-1} greatly increased model error. No combination of parameter values that include this low S_s value was found through nonlinear regression that provided an acceptable match to the measured water levels, especially during the recovery (see, for example, fig. 29). The larger-than-expected values of specific storage obtained through nonlinear regression can be explained by the exsolution of natural gas, discussed in the next section.

Remaining aquifer properties. The values for seven remaining aquifer properties that were specified as fixed in the nonlinear regression are listed in table 10. The vertical hydraulic conductivity value for the upper confining layer was based on permeameter tests that used core samples of the lacustrine sediments and till (Alpha Geoscience, 1996c). The horizontal hydraulic conductivity value specified for the lower confining layer, and the vertical hydraulic conductivity value specified for the valley walls, were obtained through preliminary regressions that decreased model error but did not converge to optimum values. Storage values specified in the model are within the range of those reported in Ireland and others (1984).

Table 10. Fixed parameter values used in nonlinear regression for transient-state simulation (model 3D-A) of confined aquifer system in Genessee Valley, Livingston County, N.Y.

[ft/d, feet per day; ft^{-1}, per foot.]

Hydraulic property	Value
Horizontal hydraulic conductivity, ft/d	
Lower confining layer	.1
Vertical hydraulic conductivity, ft/d	
Upper confining layer	1.0×10^{-5}
Valley walls	2.5
Specific storage, ft^{-1}	
Upper confining layer	1.0×10^{-5}
Lower confining layer	5.0×10^{-6}
Valley walls	1.0×10^{-6}
Specific yield of unconfined aquifer	.25



EXPLANATION

Hydraulic head, in feet above sea level

+ Measured

○ Computed with specific storage values estimated by regression

△ Computed with published value of specific storage $2.3 \times 10^{-6} \text{ft}^{-1}$

Figure 29. Water levels measured in the lower aquifer in Genesee Valley, Livingston County, N.Y. during 29 months after collapses at Retsof salt mine, and values simulated by transient-state simulation (model 3D-A) from two alternative values of specific storage: ***A.*** Location II, well Lv368. ***B.*** Location V, well Lv367. ***C.*** Site near Mt. Morris, well Lv91. ***D.*** Site near Fowlerville moraine, well Lv425. (Well locations are shown in figs. 4 and 18.)

Model bias

Model 3D-A, where optimum parameter values obtained through nonlinear regression were used, represents the drawdown and recovery of water levels reasonably well, and the computed maximum drawdown in the collapse area and the extent of water-level declines in the middle and lower aquifers match the measured values closely. The weighted residuals ($w_i^{1/2}e_i$) indicate, however, that the model is biased and thus may not correctly represent certain unidentified processes or properties within the aquifer system. Although the mean of the weighted residuals (0.13 ft) is near zero, which indicates that the regression succeeded in minimizing parameter error, the distribution of residuals about the mean is not random, as would be expected if the model accurately represented all aspects of the aquifer system.

A plot of weighted residuals and simulated heads (fig. 30) shows two clusters of residuals. Residuals in cluster A indicate significant model bias—heads in wells near the collapse area are generally underpredicted (drawdowns were greater than measured), and those in wells far from the collapse area are overpredicted (drawdowns were less than measured). This discrepancy is partly the result of exsolution of natural gas from ground water—a process that was not represented in the model. Model bias also could result from the uniform hydraulic-conductivity distribution assumed in the model, particularly near location V, where residuals are largest and local variation in hydraulic conductivity could account for the overprediction of head. Stratigraphic data are not available to support this hypothesis, however. Cluster B, in contrast, displays no apparent bias and generally corresponds to heads greater than 490 ft measured in wells near the Fowlerville Moraine and in the Beards Creek Valley. Autocorrelation in the residuals resulting from multiple measurements made at different times at the same location explains some trends in the cluster, but the effect of autocorrelation on regression estimates of parameter values is unknown.

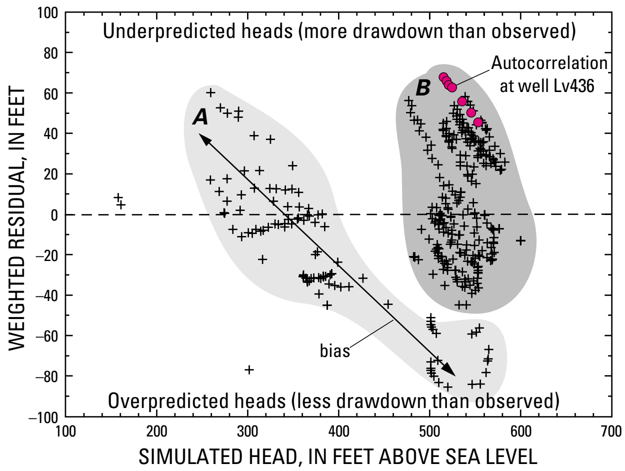


Figure 30. Simulated heads and weighted residuals computed from transient-state simulation (model 3D-A) of aquifer system in Genesee Valley, Livingston County, N.Y.

Model Results

Results of model simulations indicate that exsolution of natural gas from ground water lessened the water-level declines near the collapse area and delayed the propagation of drawdowns into areas distant from the collapse. Hydraulic heads simulated by 12.4-yr transient-state simulations indicate that water levels will return to precollapse conditions by about the year 2006. Ground-water flow paths computed from steady-state and transient-state simulations indicate that water-level declines in the confined aquifers after the mine collapse reversed the natural hydraulic gradient in the northern part of the Genesee Valley, inducing saline water from water-bearing zones in the bedrock to flow southward into the lower aquifer.

Effect of Natural Gas on Specific Storage

Specific storage values reported from studies of land subsidence are much smaller than those required to simulate the relatively slow decline and recovery of water levels measured in the Genesee Valley after the mine collapse. Specific storage S_s of a confined aquifer is usually a constant, defined as

$$S_s = \rho g(\alpha + n\beta) \quad , \tag{9}$$

where ρg = specific weight of water γ [$ML^{-2}T^{-2}$],
α = compressibility of aquifer material [$ML^{-1}T^{-2}$]$^{-1}$,
β = fluid compressibility [$ML^{-1}T^{-2}$]$^{-1}$and
n = porosity [dimensionless].

A decrease in hydraulic head results in a decrease in fluid pressure p and an increase in effective stress σ_e (see eq. 1). The release of water from storage in response to (1) aquifer compression through the increase in σ_e ($\rho g\alpha$), and (2) expansion of water through the decrease in p ($\rho g n\beta$) are accounted for in equation 9.

The larger S_s values obtained by the regression indicate that more water was released from storage than can be accounted for by the compressibility of water and aquifer material alone; therefore, the release of dissolved gas was examined as a factor in the discrepancy. The releases of biogenic gas from several wells affected by water-level declines indicate that gas had exsolved from ground water and was present as a free phase over a wide area in the lower aquifer. Exsolution of gas could significantly increase specific storage because (1) the exsolved gas would displace water in the pore volume, forcing additional releases from storage, and (2) gas compressibility is about 100 times greater than aquifer compressibility under the pressures measured in the confined-aquifer system. A gas-partitioning equation was developed to compute the contribution to specific storage provided by the exsolution of a free gas phase.

The gas-partitioning equation relates the volume of gas in the vapor phase V_v to the ambient pressure p and can be derived from the mass equation for gas in an arbitrary volume of aquifer material V_a

$$M_T = C_w V_w + C_v V_v, \tag{10}$$

where M_T = total mass of gas [M],
C_w = concentration of gas in water [ML^{-3}],
V_w = volume of water [L^3],
C_v = concentration of gas in vapor [ML^{-3}], and
V_v = volume of vapor [L^3].

The sum of the water- and vapor-filled volumes equals the volume of pore space and is expressed as

$$V_w + V_v = nV_a. \tag{11}$$

Combining equations 10 and 11 and solving for V_v gives the gas-partitioning equation

$$V_v = \frac{M_T - C_w nV_a}{C_v - C_w} \tag{12}$$

Methane solubility in water follows Henry's law for the pressure range and temperature (150 to 240 psi and 60° F) that occurred during mine flooding (Lekvam and Bishnoi, 1997). From Henry's law, the concentration of methane in water C_w at pressure p can be computed from the methane solubility C_0 measured at a reference pressure p_0 by

$$C_w = C_0 \frac{p}{p_0}, \tag{13}$$

and is related to the methane concentration in the vapor-phase pressure C_v by the dimensionless Henry's law constant K_h as

$$C_v = K_h C_w. \tag{14}$$

Assuming the pore space was initially saturated with water ($V_w = nV_a$; $V_v = 0$), and the water was saturated with methane ($M_T = C_w V_w$) at a pressure p_{init} that prevailed before the mine collapse, then from eqn. 13

$$M_T = C_0 \frac{p_{init}}{p_0} nV_a. \tag{15}$$

Combining equations 13, 14 and 15 with equation 12 and simplifying gives

$$V_v = \frac{(p_{init} - p)nV_a}{p(K_h - 1)}. \tag{16}$$

or, in units of head h:

$$V_v = \frac{(h_{init} - h)nV_a}{h(K_h - 1)}. \tag{17}$$

The specific storage resulting from gas exsolution S_{sg} is the volume of gas produced (water released) from the volume of aquifer material V_a under a unit change in pressure (h_{init} - h= 1) or

$$S_{sg} = \frac{n}{h(K_h - 1)}. \tag{18}$$

where h = hydraulic head calculated above the midpoint of the aquifer.

Equations 17 and 18 indicate that the specific storage of a confined aquifer containing dissolved gas and undergoing a water-level decline is not constant, but is a function of head (pressure), which varies temporally and spatially. Increasing fluid pressure during water-level recovery would result in dissolution of the released gas and would lower S_{sg} values as the volume of free gas in the pore volume declined.

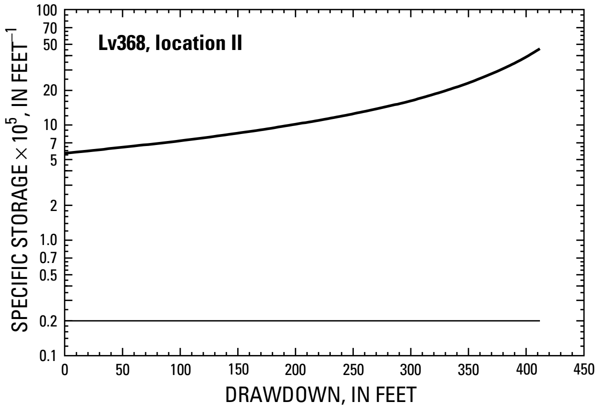
The potential magnitude of gas exsolution in the lower aquifer was calculated assuming that the methane concentration in water (C_w) was at saturation when the mine collapse occurred. Specific storage at a model cell near the collapse area (location II) was calculated for each time step from equation 18 with computed heads from the 29-month transient-state simulation (model 3D-A), and values of $n = 0.3$ and $K_h = 27.02$ (Schwarzenbach and others, 1993). The combined specific storage S_s' was computed as the sum of S_{sg} (eq. 18) and S_s (eq. 9) and is expressed as

$$S_s' = S_s + S_{sg}, \tag{19}$$

where $S_s = 2 \times 10^{-6}$ ft^{-1}, a value representative of aquifer compressibility.

Computed specific storage values S_{sg} in the 29-month transient-state simulation decreased when water levels recovered to represent dissolution of the gas. This computation assumed that the gas was trapped within the confined aquifers and did not escape from the system, and that the time required for gas dissolution was shorter than the time steps used in the transient-state simulation (2 to 120 days).

The resulting S_s' values near location II, the point of maximum water-level drawdown at the collapse area (fig. 31), ranged from 2.7×10^{-5} ft^{-1} to 1.8×10^{-4} ft^{-1}.



EXPLANATION

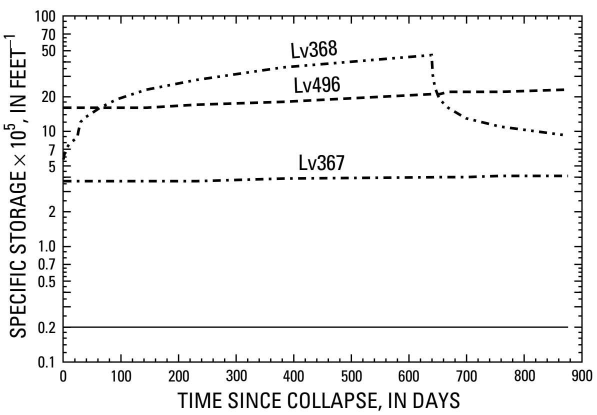
Specific storage

$S_{s'}$, water and gas

S_s, without gas

Figure 31. Effect of gas exsolution and expansion on specific storage values in lower aquifer during drawdown at location II near collapse area in Genesee Valley, Livingston County, N.Y., March 1994 through January 1996. (Location is shown in fig. 4.)

The S_s' value was initially greater (8.0×10^{-5} ft^{-1}) near location I in the north (well Lv496 in fig. 32), where the lower aquifer (depth 210 ft) is shallower than at location II (depth 480 ft), because the lower pressure at location I allowed greater expansion of exsolved gas. The maximum S_s' value at location I (1.2×10^{-4} ft^{-1})



EXPLANATION

Specific storage

$S_{s'}$, water and gas, Lv368, location II

$S_{s'}$, water and gas, Lv496, near Fowlerville moraine

$S_{s'}$, water and gas, Lv367, location V

S_s, without gas

Figure 32. Change in specific storage during drawdown and recovery at selected wells screened in lower aquifer in Genesee Valley, Livingston County, N.Y., during 29 months after collapses at Retsof salt mine.

was less than the value at location II, however, because water-level declines were less than those near the collapse area. The value of S_s' at location V in the south, where the aquifer depth was more than 700 ft, ranged from 1.8×10^{-5} ft^{-1} to 1.9×10^{-5} ft^{-1}. Unlike specific storage values at location II, the values at locations I and V continued to increase after the mine was filled in January 1996 (fig. 32) because water levels continued to decline at these locations and did not begin to recover until about 6 months later. The maximum percentage of pore volume occupied by gas was calculated to be 7 percent at the collapse area (location II) and less than 1 percent at locations I and V.

The effects of gas exsolution were incorporated into the computer program MODFLOWP (model 3D-B in table 4) in which the combined specific storage S_s' (eq. 19) was computed in each model cell for every time step. Two simulations were run with model 3D-B. In the first simulation, C_w was assumed at saturation throughout the confined aquifers, so that gas exsolution began with the onset of water-level declines. In the second simulation, the methane concentration was below saturation in parts of the middle and lower aquifers, such that

$$S_{sg} = 0, for(h < h_0), \tag{20}$$

where h_0 = head at which C_w is at saturation. In the second simulation gas exsolution did not occur until declining water levels reduced the solubility to the dissolved gas concentration at the threshold h_0.

Optimum parameter values were estimated in these two simulations by nonlinear regression with UCODE (Poeter and Hill, 1998). The regression method in UCODE is similar to that used in the model 3D-A, except that sensitivities of model parameters required by the UCODE method are estimated through a perturbation technique, rather than computed directly from an analytical expression, as in MODFLOWP. The set of parameters estimated in the first regression was the same set estimated in model 3D-A, while two additional parameters representing h_0 values for the middle and lower aquifers were estimated in the second regression. Specific storage values estimated with model 3D-B for both regressions (table 11) were similar and near the upper end of the range reported for sand and gravel aquifers (2.3×10^{-6} ft^{-1}) in studies of land subsidence. Hydraulic conductivity of the lower aquifer was estimated to be 230 ft/d assuming methane concentration at saturation or 180 ft/d assuming methane concentra-

Table 11. Optimum parameter values estimated for confined aquifer system (model layers 2 through 5) in Genessee Valley, Livingston County, N.Y., through nonlinear regression in transient-state simulation with models 3D-A (constant specific storage) and 3D-B (variable specific storage values based on gas exsolution).

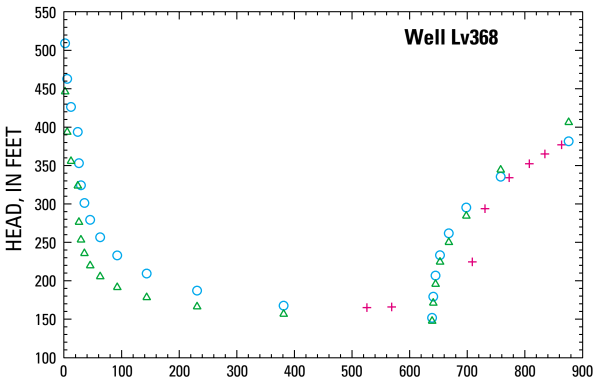
[ft/d, feet per day, ft^2, feet squared; ft^{-1}, per foot; —, not estimated].

Variable	Model 3D-A	Model 3D-B with methane concentration: At saturation	Model 3D-B with methane concentration: Less than saturation
Aquifer property			
Hydraulic conductivity, ft/d			
Middle aquifer	3.7	3.7	3.1
Lower aquifer	300	230	180
Vertical hydraulic conductivity of lower confining layer, ft/d			
Collapse area	2.7×10^{-2}	2.7×10^{-2}	3.0×10^{-2}
Remainder of layer	1.2×10^{-3}	1.2×10^{-3}	1.4×10^{-3}
Specific storage, ft^{-1}			
Middle aquifer	6.9×10^{-5}	2.6×10^{-6}	2.3×10^{-6}
Lower aquifer	2.9×10^{-4}	2.5×10^{-6}	2.3×10^{-6}
Threshold head h_o, ft			
Middle aquifer	—	—	86
Lower aquifer	—	—	520
Model error			
Sum of squared errors, ft^2	4.04×10^5	4.17×10^5	3.93×10^5
Standard error in heads, ft	32.9	33.3	32.4

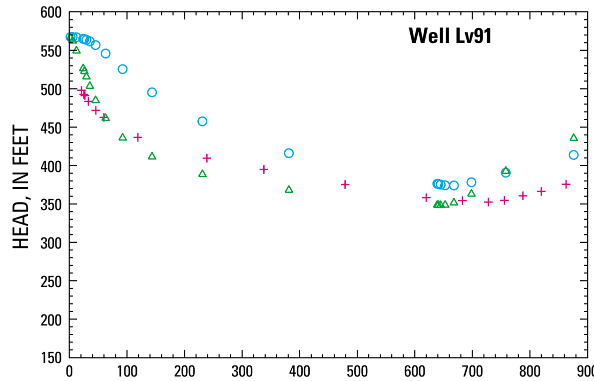
tion below saturation. These values are less than the 300 ft/d estimated with model 3D-A, in which the initial specific storage value was larger and required a larger hydraulic conductivity to simulate the rapid propagation of drawdown through the lower aquifer. The methane concentrations corresponding to the estimated h_o values in the second regression were 65 mg/L in the middle aquifer and 390 mg/L in the lower aquifer.

Including the effects of gas exsolution in model simulations caused little change in model error (table 11). Drawdowns simulated by model 3D-B were about the same as those computed with model 3D-A for areas near the collapse (fig. 33*A*) but matched more closely the measured drawdowns in areas farther away (fig. 33*B*, *C*). Drawdowns far from the collapse area were underpredicted (fig. 33*D*), indicating that the computed S_s'-values were still too large in these areas. Computed water levels also recovered much sooner than those measured. Some of the free gas phase that accumulated in the lower aquifer near the collapse area was thermogenic and originated from deeper reservoirs, but the volume of thermogenic gas within the lower aquifer is unknown and cannot be accounted for in model simulations. As a result, model 3D-B underestimates the volume of water released from storage near the collapse area and overestimates the volume released far from the collapse, producing a trend of underpredicting drawdowns in wells far from the collapse area. Displacement of water from the pore volume by thermogenic gas and variability in the initial methane concentration within the lower aquifer probably account for much of the remaining model error.

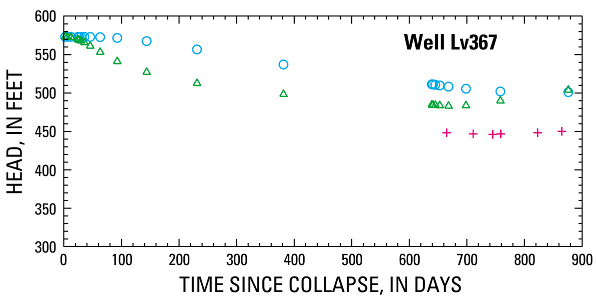
The exsolution of natural gas from ground water adequately explains the response of water levels in the confined part of the aquifer system to the collapse within the mine. The release of additional water from storage in response to gas exsolution lessened water-level declines near the collapse area and slowed the propagation of drawdowns into areas away from the



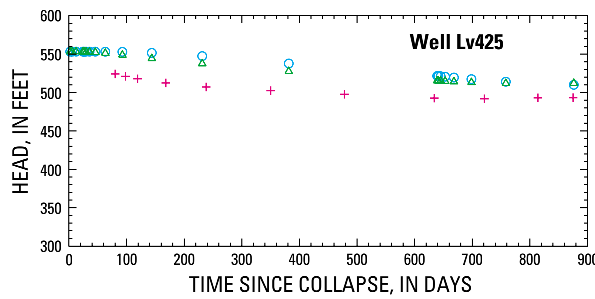
A. Location II, well Lv368.



C. Site near Mount Morris, well Lv91.



B. Location V, well Lv367.



D. Site near Fowlerville moraine, well Lv425.

EXPLANATION

Hydraulic head, in feet above sea level

+ Measured

○ Computed with specific storage values estimated by regression

△ Computed with gas exsolution

Figure 33. Water levels measured in lower aquifer in Genesee Valley, Livingston County, N.Y. during 29 months after collapses at Retsof salt mine, and values simulated by transient-state simulations with models 3D-A (constant specific storage) and 3D-B (variable specific storage values based on gas exsolution with methane concentrations below saturation): ***A.*** Location II, well Lv368. ***B.*** Location V, well Lv367. ***C.*** Site near Mt. Morris, well Lv91. ***D.*** Site near Fowlerville moraine, well Lv425. (Well locations are shown in fig. 4.)

collapse. Uncertainty remains, however, as to the initial distribution of gas concentrations within the confined aquifers, the fate of the exsolved gas and its effect on the relative permeability of the confined-aquifer system.

The slower-than-simulated recovery of water levels suggests that one or more unknown but important processes are not represented in model 3D-B. For example, although the rate of gas exsolution was probably rapid because gas was present throughout the pore volume, the rate of gas dissolution could be limited by the size of the interface separating gas-filled pores from water-filled pores. A relatively high rate of gas exsolution followed by a lower rate of gas dissolution could explain the measured rapid drawdown of water levels in the confined aquifers followed by a relatively slow water-level recovery. Some of the exsolved gas could have remained trapped within the pore volume, or migrated upward toward land surface, rather than redissolving as water levels recovered. Laboratory studies have shown that displacement of water by gas can cause gas to become trapped in pore spaces and thereby reduce the effective porosity of aquifer material (Czolbe, 1997).

If gas occupied 7 percent of the pore volume in the lower aquifer at location II, as computed with model 3D-B, the relative permeability of aquifer material to water (K_w) could decrease by as much as 50 percent if the gas were distributed uniformly throughout the pore volume (Coats and Richardson, 1967). This would not prevail for long, however, because the gas would soon migrate upward and form bubbles that would be trapped by the lower confining layer, and this would reduce the saturated thickness (and transmissivity) by only 7 percent. If the gas migration were impeded, however, the decreased relative permeability could delay water-level recovery in some parts of the lower aquifer. Additionally, the upward migration of gas along the valley walls could reduce the K_w of permeable sediments that provide hydraulic connection between the confined aquifers and land surface and thereby decrease the potential for recharge and slow the recovery of water levels.

A two-phase flow model that represents water and gas movement could be used to simulate gas migration and its potential effects on relative permeability, but the use of such a model for this setting would be impeded by uncertainty regarding the initial gas concentration, the migration pathways, and the altitude and topography of the bottom of the lower confining layer, which probably forms a barrier to upward gas migration.

Water-Level Recovery

Water-level recovery in the confined-aquifer system after the collapse within the salt mine was simulated through 12.4-year transient-state simulations beginning with the first collapse in March 1994. Water levels simulated by model 3D-A (constant storage, table 4) had recovered nearly 90 percent by the year 2000 and returned to precollapse conditions at most locations by the year 2006 (fig. 34). Decreasing the specific storage value (to 2.3×10^{-6} ft^{-1}) for the lower aquifer shortened the recovery to precollapse conditions by 7 years (1999) because the contribution of storage to the inflow decreased sharply (18 percent, as opposed to the initial 73 percent), whereas the contribution from vertical leakage along the valley walls more than doubled (55 percent of inflow, as opposed to the initial 24 percent).

The effect of the vertical leakage rate on water-level recovery was investigated through two additional simulations in which the conductance of cells (see eq. 5) representing vertical leakage through permeable deposits and bedrock fractures along the valley wall was increased or decreased by 1 order of magnitude. Water levels simulated with the increased conductance returned to precollapse conditions by the year 2002, but those with the decreased conductance had not completely recovered by the year 2006 (fig. 35). Simulated water budgets indicated that a 1-order-of-magnitude increase in conductance caused the vertical leakage rate to increase by about a factor of 3.

Effect of Mine Collapse on Saline-Water Intrusion

Hydraulic heads and flow rates computed by model simulations were used with a particle-tracking routine developed by Pollock (1989) to generate ground-water flow paths in the northern part of the Genesee Valley, which was affected by intrusion of saline water within the lower aquifer. Flow paths were generated to delineate the potential migration of saline water from suspected sources at the Bertie subcrop and

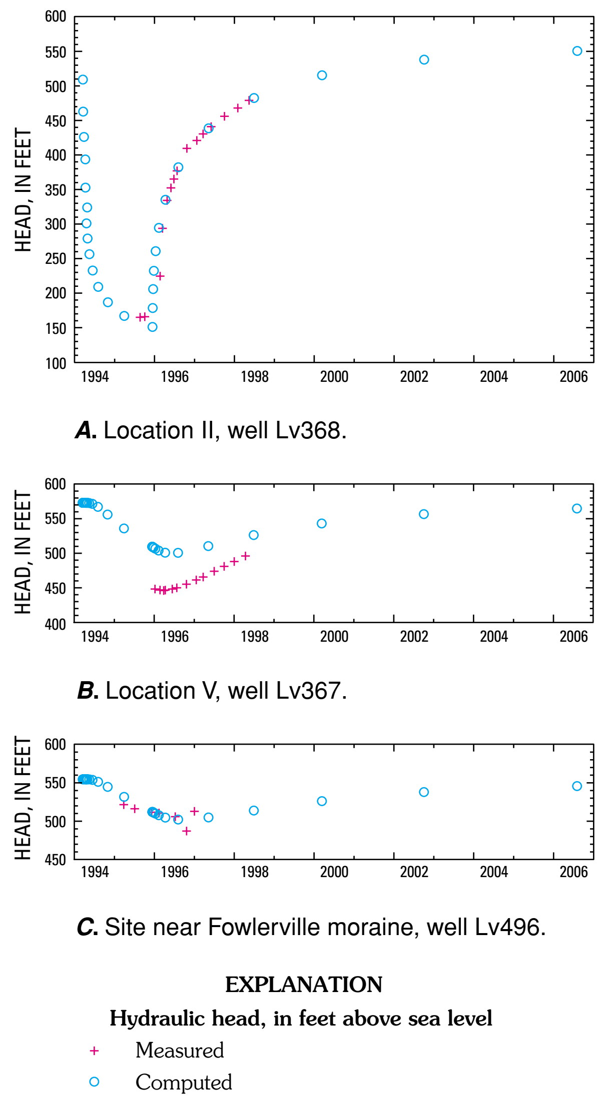


Figure 34. Water levels measured in lower aquifer in Genesee Valley, Livingston County, N.Y. during first 900 days after collapses at Retsof salt mine, and water levels simulated for 1994 through 2006 by model 3D-A: ***A.*** Location II, well Lv368. ***B.*** Location V, well Lv367. ***C.*** Site near Fowlerville moraine, well Lv496. (Well locations are shown in fig. 4.)

along Fowlerville Road (fig. 36) through the lower aquifer under both pre- and postcollapse conditions. The difference in density between fresh and saline waters was not considered in the flow-path delineation because its effect is small, as discussed below.

Whether saline water moved to the affected wells as horizontal flow in the lower aquifer or as upward leakage through fractures in the bedrock, or both, is uncertain. Horizontal migration would result in a continuous plume of saline water within the aquifer, whereas vertical migration would result in locally discontinuous areas with greater-than-average concentrations of dissolved solids. Flow paths were delineated through model simulations to assess the potential for horizontal migration through the lower aquifer.

Ground-water flow velocity. Ground water in the northern part of the Genesee Valley under precollapse conditions flowed northeastward through the lower aquifer at an estimated mean velocity of about 2 ft/d, if an effective porosity of 0.3 is assumed for the aquifer material (fig. 36*A*). The length of the vectors in figure 36A indicates the distance traveled from selected model cells in 29 months (the duration of the 876-day transient-state simulation). The mine collapses caused the hydraulic gradient to reverse, and ground water began moving southward toward the collapse area. Estimates of the onset of the flow reversal and the velocity of ground-water flow are dependent on the assumed values for specific storage and effective porosity of the aquifer material.

Flow paths derived from the heads and flows computed by model 3D-A indicate that the reversal had begun by January 1996 and that the mean velocity was about 1 ft/d (fig. 36*B*). These flow paths indicate that saline water originating north of Fowlerville Road would not have reached the affected wells by the winter of 1994-95, when salinity first appeared in well Lv457. The specific storage value (2.9×10^{-4} ft^{-1}) specified for the lower aquifer in this area in model 3D-A does not account for exsolution of natural gas, however, and is probably too large; therefore, an additional simulation was run in which the specific storage value was decreased more than 100-fold to 2.3×10^{-6} ft^{-1}. Results indicated that ground water could have begun to flow southward at a mean velocity of 2 ft/d about 2 months after the first collapse with (fig. 36*C*); nevertheless, saline water would not have reached the affected wells by the winter of 1994-95, even at this faster flow rate.

The apparent velocity of saline water can be estimated from the time of the appearance of salinity in wells Lv457 and Lv500 along Little and Hogmire Roads, respectively (fig. 36*C*), if saline water is assumed to have moved southwestward from well Lv457. Water from well Lv457 was potable when this well was drilled in September 1994 but had a specific conductance of more than 12,000 µS/cm in March 1995, and water from well Lv500 had a specific con-

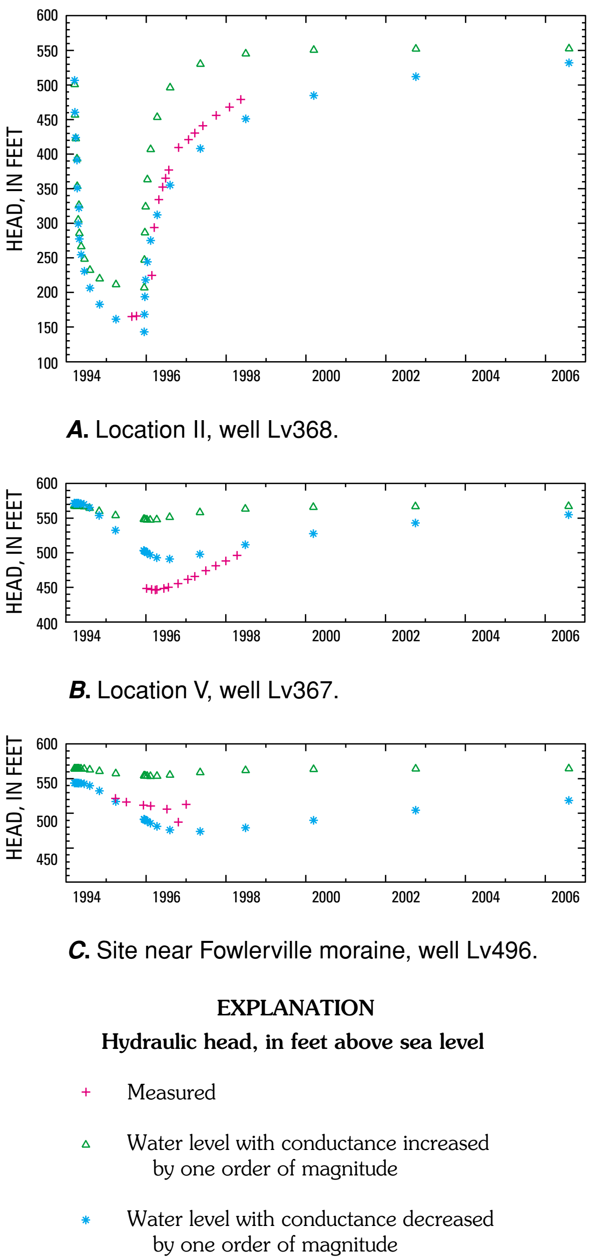
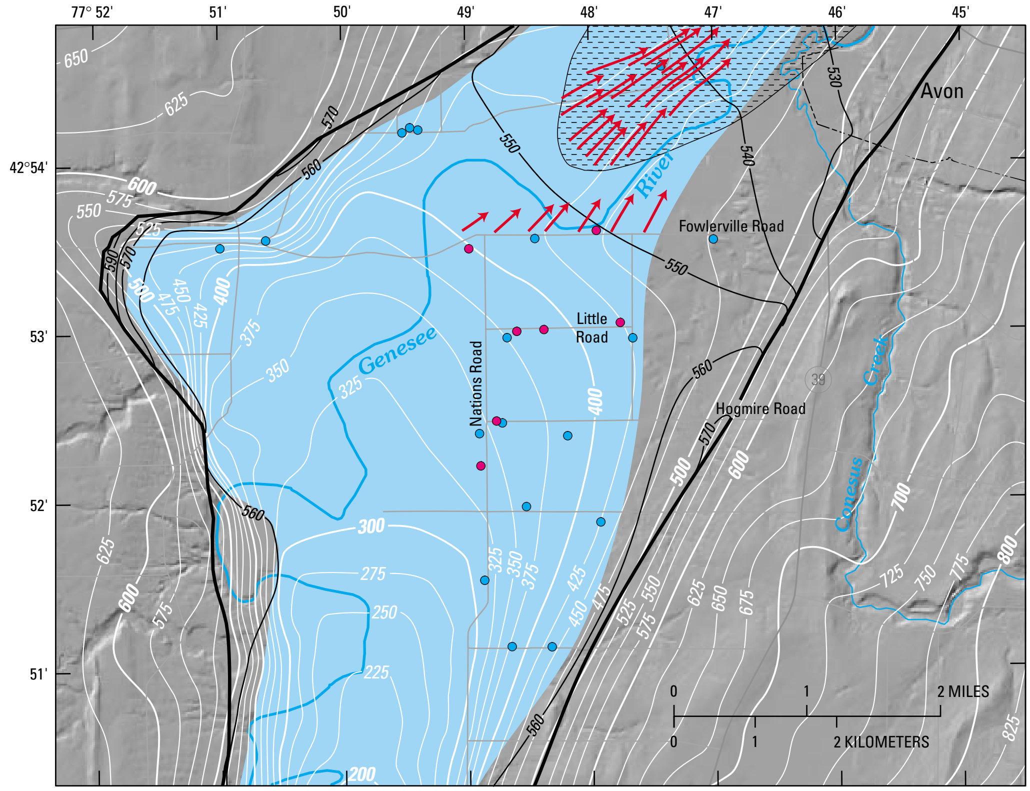
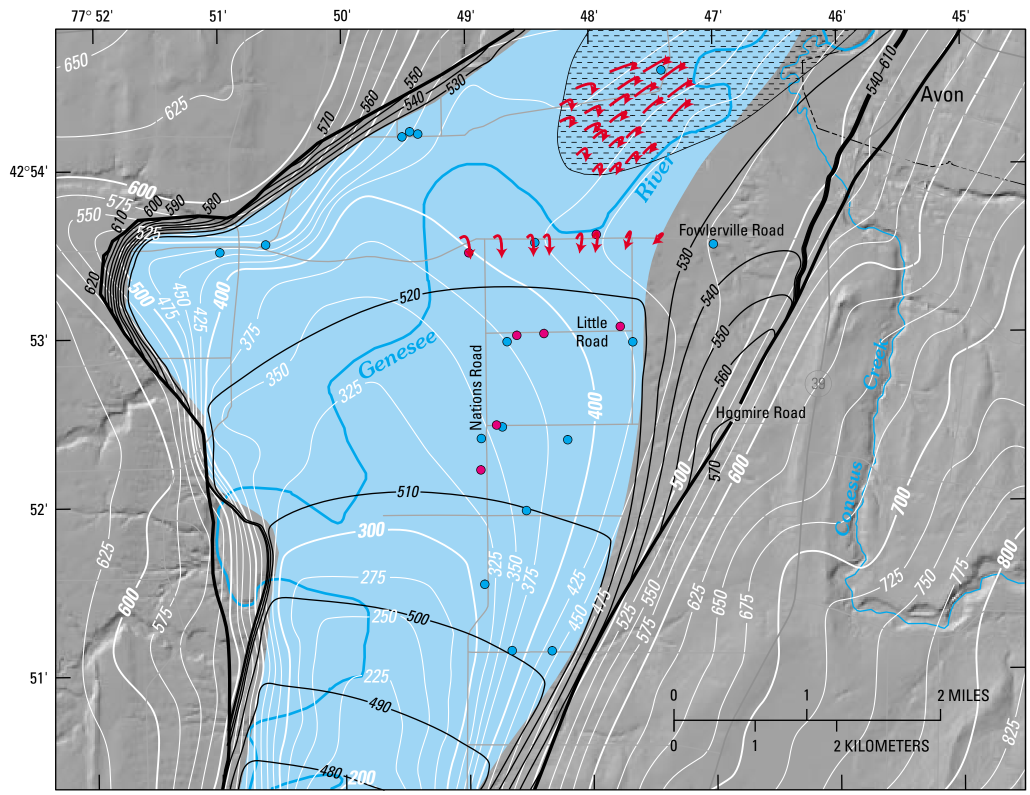


Figure 35. Water levels measured in lower aquifer in Genesee Valley, Livingston County, N.Y. during 29 months after collapses at Retsof salt mine, and water levels simulated for 1994 through 2006 by model 3D-A with vertical conductance increased and decreased by one order of magnitude: ***A.*** Location II, well Lv368. ***B.*** Location V, well Lv367. ***C.*** Site near Fowlerville moraine, well Lv496. (Well locations are shown in fig. 4.)



Base from U.S. Geological Survey digital data, 1:100,000, 1983

A. Under precollapse (March 1994) conditions; flow paths computed with steady-state simulation (model 3D-A).

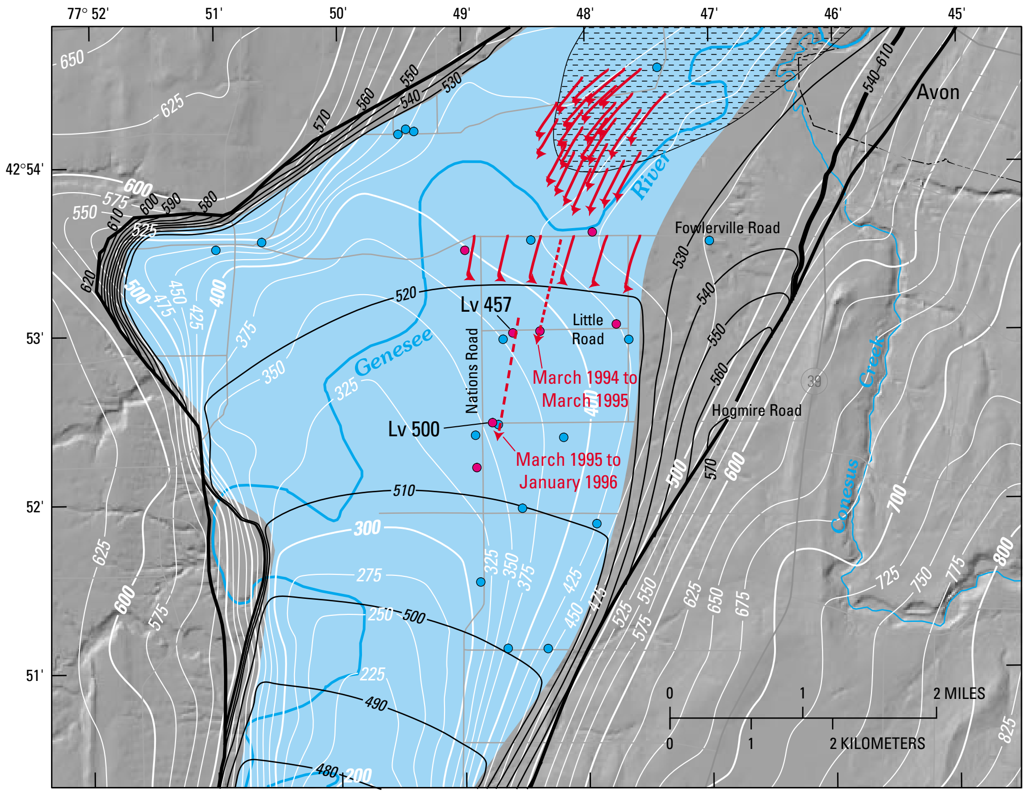


Base from U.S. Geological Survey digital data, 1:100,000, 1983

B. In August 1996; flow paths computed with transient-state simulation (model 3D-A).

Figure 36. Hydraulic-head distribution in the lower aquifer in northern part of the Genesee Valley, Livingston County, N.Y. and ground-water flow paths from time of first Retsof mine collapse (March 1994) through August 1996 (29 months): ***A.*** Under precollapse (March 1994) conditions; flow paths computed with steady-state simulation (model 3D-A). ***B.*** In August 1996; flow paths computed with transient-state simulation (model 3D-A).

ductance of 4,700 µS/cm in June 1995 which had increased to 12,400 µS/cm in January 1996. The maximum traveltime between these two wells is 450 days (September 1994 through January 1996); therefore, the ground-water velocity is at least 8 ft/d (3,700 ft / 450 d); this is 4 to 8 times the range (1 to 2 ft/d) indicated by model 3D-A. The effective porosity *n* can be computed from the relation



Base from U.S. Geological Survey digital data, 1:100,000, 1983

C. In August 1996; flow paths computed with transient-state simulation (model 3D-A) assuming high-permeability paths and with specific storage value (2.3×10^{-6} ft^{-1}) calculated from published values obtained in land-subsidence studies.

EXPLANATION

Lower aquifer

Bertie limestone

— *550* — **Simulated potentiometic contour**—Contour interval 10 feet. Datum is sea level

300 *350* **Bedrock contour**—Contour interval 25 feet. Index contours every 100 feet. Datum is sea level

Ground-water flow path—Indicates direction and distance travelled:
29 months after 1994 mine collapse;
March 1994 to March 1995
March 1995 to January 1996

Boundary of model area

Sampled well

Affected well—Specific conductance exceeded 2,500 microsiemens per centimeter

Figure 36— Continued. Hydraulic-head distribution in the lower aquifer in northern part of the Genesee Valley, Livingston County, N.Y. and ground-water flow paths from time of first Retsof mine collapse (March 1994) through August 1996 (29 months): ***C.*** In August 1996; flow paths computed with transient-state simulation (model 3D-A) assuming high-permeability paths and with specific storage value (2.3×10^{-6} ft^{-1}) calculated from published values obtained in land-subsidence studies. (Locations are shown in fig. 13.)

$$v = \frac{Ki}{n} \qquad (21)$$

where v = ground-water velocity [LT^{-1}],
K = hydraulic conductivity [LT^{-1}], and
i = hydraulic gradient [LL^{-1}].

The effective porosity of the lower aquifer is about 5 percent, as indicated by equation 21, and the hydraulic conductivity value estimated from the regression (300 ft/d), and the hydraulic gradient measured in January 1996 (1.2×10^{-3} ft/ft). When an effective porosity of 5 percent and a specific storage of 2.3×10^{-6} ft^{-1} were specified, the resulting ground-water flow paths (fig. 36*C*) were consistent with the assumed traveltimes between Fowlerville Road and well Lv457 (March 1994 through March 1995), and between wells Lv457 and Lv500 (March 1995 through January 1996). An effective porosity value of 5 percent is unreasonably low for the sand and gravel that forms the aquifer material, however; the commonly reported range is 25 to 40 percent (Dennis LeBlanc, USGS, written commun.,1991).

Alternatively from that described above, the calculated velocity of 8 ft/d could be explained if the effective porosity along the flow path travelled by saline water were increased to a realistic value of 30 percent and the hydraulic conductivity were increased to 2,000 ft/d . This hydraulic conductivity value, although much larger than the regression estimate of 300 ft/d, is within the range reported for sand and gravel. Coarse deposits with hydraulic conductivity values of 2,000 ft/d could be present in the lower aquifer. Little stratigraphic information is available to confirm this result, however, and additional information would be needed for a detailed representation of hydraulic conductivity in the ground-water flow model.

Direction of flow. The effect of water density on the direction of ground-water flow can be assessed through a comparison of local hydraulic and elevation gradients. The direction and rate of saline-water flow is given by the Darcy velocity vector **v**, defined as (Davies, 1989)

$$\mathbf{v} = -K\left[\nabla h_{fw} + \frac{\Delta\rho}{\rho_{fw}}\nabla E\right], \qquad (22)$$

where ∇h_{fw} = gradient of freshwater hydraulic head[$L \bullet L^{-1}$],

$\Delta\rho$ = density difference between saline water and freshwater (ρ_{fw})[ML^{-3}], and (23)

∇E = elevation gradient [$L \bullet L^{-1}$]. (24)

The Darcy velocity vector is the sum of two vectors—one defined by the hydraulic gradient ∇h_{fw}, and the other by elevation gradient ∇E, scaled by the factor $\Delta\rho/\rho_{fw}$. This factor was computed as 4.5×10^{-3}, from the measured density of water samples from wells Lv370 (dissolved solids concentration 7,000 mg/L) and Lv371 (dissolved solids concentration 410 mg/L) at location I, near the Fowlerville Moraine. The elevation gradient near well Lv500 (1.2×10^{-2} ft/ft) is almost 100 times greater than the hydraulic gradients under both precollapse conditions (3.9×10^{-4} ft/ft) and postcollapse conditions of August 1996 (1.2×10^{-3} ft/ft). The density difference between the saline water and freshwater is small enough, however, that the elevation gradient is equivalent to about 10 precent of the hydraulic gradient after it is scaled by the factor $\Delta\rho/\rho_{fw}$; this confirms that the density difference has little effect on the direction of horizontal ground-water flow.

Sources of saline water. Several questions remain concerning the source of salinity, as well as the velocity, and direction of saline water migration in the lower aquifer. The source could be the Onondaga and Bertie Limestones, the Salina saltbeds, or a combination of both, as discussed previously. Saline water could have moved to the affected wells as horizontal flow in the lower aquifer or as upward leakage through fractures in the bedrock, or both. Estimates of the velocity of saline water ranged from 1 to 8 ft/d, depending on the values of hydraulic conductivity, specific storage, and effective porosity specified for the lower aquifer. Saline water also could migrate through paths of high permeability that are not represented in model 3D-A.

Saline water will probably continue to move southward toward the collapse area until water levels in the confined aquifers recover to the point where gradients in the lower aquifer are once again northward—or until about the year 2006, as indicated by the simulations of model 3D-A (fig. 34). The saline water then will move northward with the hydraulic gradient because the small density difference between saline and fresh water is not large enough to alter the direction of horizontal flow. The northward gradient (3.9×10^{-4} ft/ft) estimated for precollapse conditions is about one-third the southward gradient in August 1966 (1.2×10^{-3} ft/ft); therefore, (1) the northward velocity of saline water will be one-third the southward velocity during postcollapse conditions, and (2) if saline water migrates southward for 10 years after the collapses, another 30 years will be required for freshwater to flush the area into which the saline water has intruded. This calculation is conservative because the southward gradient was not constant, but gradually increased to a maximum in January 1996, when drawdowns throughout the aquifer system were largest. Dispersive mixing of freshwater and saline water will probably dilute the saline water, however, and decrease the time required for ground water to become potable.

Additional study based on stratigraphic and geochemical information from new test holes would be needed to determine the source of salinity, velocity of saline-water migration, and the extent of saline-water intrusion; this study would provide the data needed to refine the ground-water flow model 3D-A to incorporate local variations in the hydraulic conductivity in the lower aquifer.

SIMULATION OF LAND SUBSIDENCE

As much as 0.8 ft of land subsidence developed south of the Retsof salt mine in response to water-level declines and the resulting compression of fine-grained sediments in the upper and lower confining layers after the mine collapse. Two 1-dimensional models (models 1D-A and 1D-B, table 4) were developed with MODFLOW with the interbed-storage package (Leake and Prudic, 1991) to estimate hydraulic properties of the confining layers from observations of (1) pore pressure near the collapse area, and (2) land subsidence 2,800 ft south of the collapse area. The interbed-storage package (IBS1) allows the specification of two storage coefficients—one to represent elastic compression, and one to represent inelastic compression, of fine-grained sediments. Estimates of hydraulic values derived from calibration of the two 1-dimensional models were then substituted in the three-dimensional flow model 3D-A described earlier to depict the spatial distribution of land subsidence (model 3D-C, table 4).

One-Dimensional Simulations

One-dimensional, transient-state simulations were run to investigate vertical flow and compression in the confining layers at location II, where pressure transducers were installed in borehole XD1, and at monument K10 along survey line K, 2,800 ft south of XD-1 (fig. 17). Two pressure transducers were installed in borehole XD1 within the upper confining layer at a depth of 216 ft in September 1995. The transducers were placed within a saturated sand pack, and the borehole was sealed to land surface with an expansive grout to isolate the monitored interval from atmospheric pressure (Alpha Geoscience, 1996c). A datalogger monitored the transducers at 15-min intervals to provide a continuous record of pore pressure in the upper confining layer from September 1995 to the present (1998). Land-surface altitudes were surveyed monthly along survey line K from September 1994 through July 1996, and then quarterly to the present (Akzo Nobel Salt, 1994-97; Kenneth Cox, ANSI, written commun., January 1997).

Model Design

Generalized stratigraphic sections (fig. 37) show the thickness and depth of the confining layers and the confined aquifers at both locations (location II and monument K10). The model for location II (model 1D-A in table 4) represented only the upper confining layer and used a column of 257 cells of unit volume (1 ft^3). The model for survey monument K10 (model 1D-B in table 4) represented both the upper and lower confining layers with a column of 446 unit-volume cells. Aquifer boundaries in both models were assigned hydraulic heads generated from transient-state simulations in model 3D-A. Subsidence resulting from compression of the confining layers was computed as the sum of the volume of water released from storage.

The confining layers in each model were assigned values of vertical hydraulic conductivity K_v and specific storage S_s. The K_v value for the upper confining layer in both models was obtained through the nonlinear regression method described below, and the K_v value for the lower confining layer (model 1D-B) was fixed at the value estimated from model 3D-A (1.2×10^{-3} ft/d). The S_s values assigned to both confining layers were chosen to represent elastic and inelastic compression and were estimated by regression from the relation

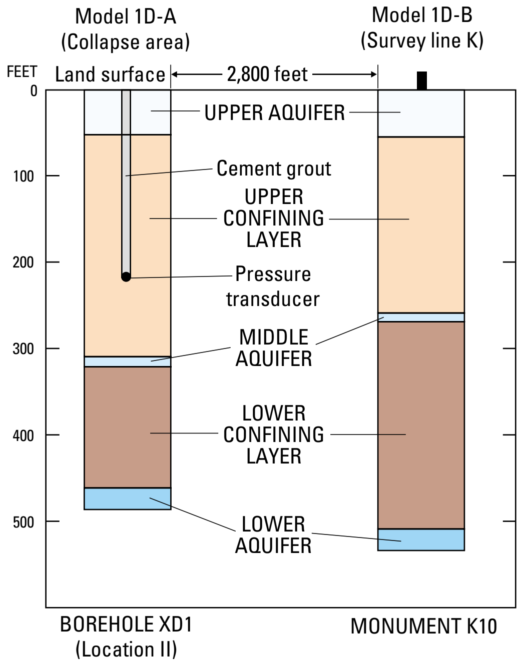


Figure 37. Generalized stratigraphic sections showing relative thickness of units and the depths of confined aquifers at borehole XD1 (location II) and monument survey K10 (2,800 ft south of the collapse area), in Genesee Valley, Livingston County, N.Y. (Locations are shown in fig. 17.)

$$S_s = \rho g(\alpha + n\beta) , \qquad (25)$$

where β = compressibility of water (4.3×10^{-7} ft^{-1}),
n = mean porosity measured in consolidation tests (0.3),
α = compressibility of the confining-layer sediments, defined as the ratio of the volumetric deformation undergone by sediments to the magnitude of the stress that causes deformation.

The value α can be estimated from consolidation tests from the relation

$$\alpha = -\frac{(\Delta e)/(1 + e_0)}{\Delta\sigma_e} , \qquad (26)$$

where Δe = change in void ratio caused by a change in effective stress $\Delta\sigma_e$, and
e_0 = initial void ratio.

Consolidation-test results for compressible materials (such as clays) typically display a two-part curve corresponding to an elastic (recoverable) response at relatively low stress, and an inelastic (nonrecoverable) response at higher stresses that are greater than the previous maximum stress applied to the material (fig. 38*A*). The increase in effective stress in the lower confining layer near location II after the mine collapse (12 ton/ft^2 or 167 psi) also is indicated in fig. 38*A*. Compressibility in the inelastic range is generally 20 to 100 times greater than in the elastic range (Riley, 1998).

Consolidation curves for lower confining-layer sediments indicate that the stress resulting from water-level declines after the collapse corresponded to the transition from the elastic to the inelastic stress range (fig. 38*B*). The maximum stress applied in the consolidation tests was insufficient to cause an inelastic response of these sediments; therefore, the previous maximum stress undergone by the lower confining-layer sediments cannot be estimated from the consolidation curves. The increased stress in the upper confining layer was not much greater than the previous maximum stress because relatively little drawdown occurred in the upper confining layer.

Compressibility under ambient stress before the mine collapse was computed from consolidation curves for 15 samples of confining-layer sediments (Alpha Geoscience, 1995a,b,c; 1996a,b) through equation 26. The results (fig. 38*B*) indicate a decreasing void ratio during the compression phase of the test (increased stress $\Delta\sigma_c$) and an increasing void ratio during the rebound phase (decreased stress $\Delta\sigma_r$). Compressibility at the ambient stress σ_e was computed for the compression and rebound phases of the tests.

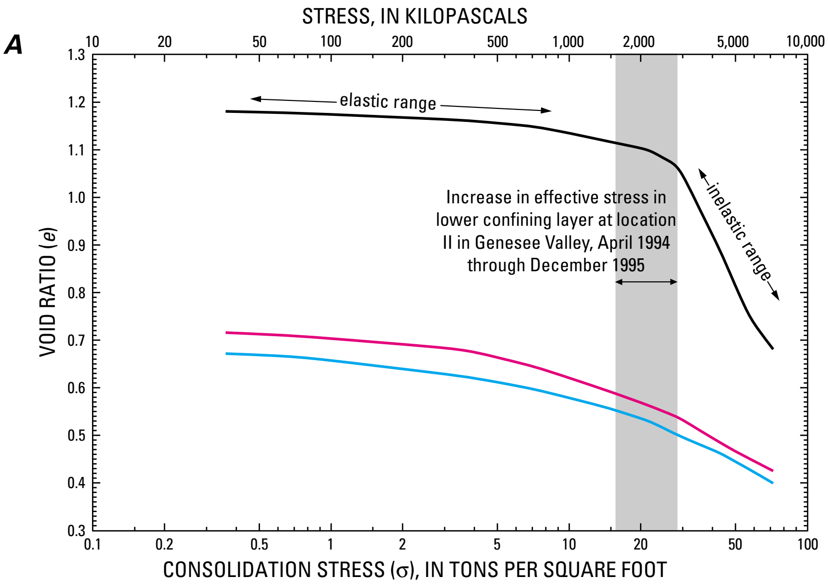
Compressibility of the sediments generally declined with increasing σ_e at increasing burial depths (d) , as expected from empirical relations given in Jorgenson (1980) and Neuzil (1986). The relation of compressibility to depth is

$$\alpha = \frac{m}{d} \qquad (27)$$

where m = constant (MT^{-2}).

For example, eq. 27 with an m-value of 5.5×10^{-2} ft/psi gives the compressibility values computed from consolidation curves during compression with a correlation coefficient (r^2) of 0.94 (fig. 39). Substituting eq. 27 into eq. 25 yields

$$S_s = \rho g\left(\frac{m}{d} + n\beta\right) . \qquad (28)$$



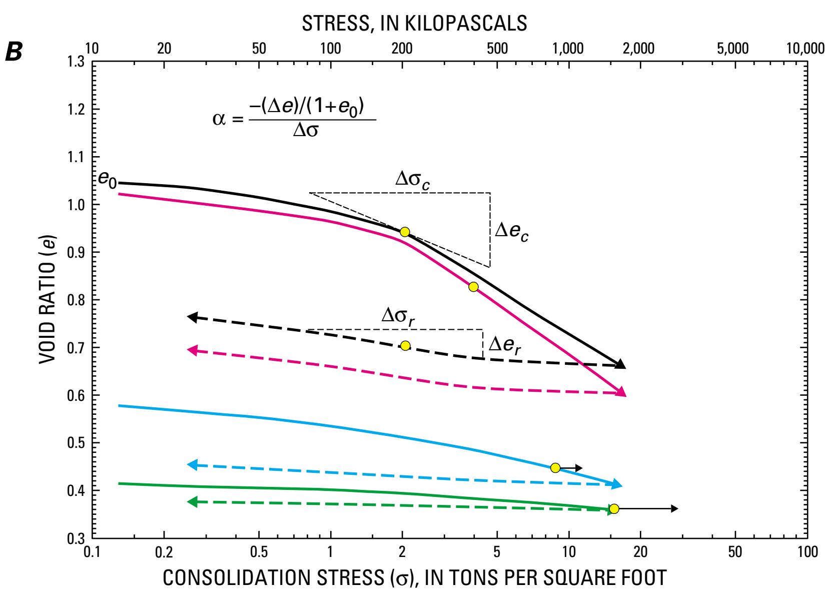
EXPLANATION

Core samples from borehole in California—Sample number and depth below land surface (from Johnson, 1984)

23L-83, 699 feet

23L-84, 746 feet

23L-85, 984 feet



EXPLANATION

Sample and depth—Solid line is compression phase (*c*), dashed line is rebound phase (*r*), arrow indicates change in stress during test

Upper confining layer

ST-1, 53 feet, above till

ST-2, 98 feet, above till

ST-3, 218 feet, below till

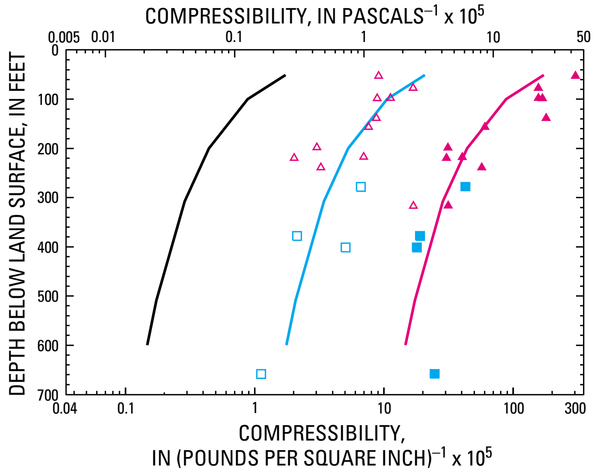
Lower confining layer

ST-4, 401 feet

Increase in effective stress after mine collapse, April 1994 though December 1995

Ambient effective stress before mine collapse, April 1994

Figure 38. Examples of consolidation curves for: ***A.*** Compressible clays in California. (Modified from A. I. Johnson, 1984, fig. 4.16, corehole 14/13-11D1). ***B.*** Confining-layer sediments from well Lv346 at location II near the collapse area, Livingston County, N.Y. (Location is shown in fig. 4.)



EXPLANATION

Estimated *m*-values, in foot per psi relating compressibility α **to depth d—**$\alpha = m / d$ (equation 27)

ID-B (table 4) with elastic and inelastic storage, 2.4×10^{-4}

ID-B (table 4) with elastic specific storage, 6.5×10^{-3}

Estimated from compression data, 5.5×10^{-2}

Consolidation tests

Compression	Rebound
▲ Upper confining layer	△ Upper confining layer
■ Lower confining layer	□ Lower confining layer

Figure 39. Relation of compressibility to depth as derived from consolidation curves of confining-layer sediments in Genesee Valley aquifer system, Livingston County, N.Y.

Compressibility values computed from the rebound phase of the consolidation tests are about 1 order of magnitude smaller than those computed from the compression phase. Possibly, the sediment samples had expanded under atmospheric conditions prior to the compression phase at the beginning of the consolidation test and, therefore, were not representative of ambient conditions. By the end of the compression phase, however, the samples had been recompressed to a state of stress that resembled ambient conditions more closely than at the begininning of the tests.

Values of elastic and inelastic storage were specified for cells in models 1D-A and 1D-B through the following procedure: an elastic storage value $S_{se(i)}$ was computed for each cell i from equation 28 with the depth below land surface d_i and an m-value estimated by nonlinear regression (The product ρg was specified as 62.4 lb/ft^3 or 0.434 psi/ft). The magnitude of stress (preconsolidation head H_{pc}) at which the transition from elastic to inelastic storage occurred was also estimated in the regression. Inelastic storage values $S_{sin(i)}$ were specified to be either 30 or 50 times greater than $S_{se(i)}$ values. The sensitivity of model results to specific storage values that were estimated from the compression and rebound phases of the consolidation tests and that represented elastic and inelastic compression is discussed below.

Model Calibration

A nonlinear regression method (PEST; Doherty and others, 1994) was used to estimate three parameters (K_v, m, and H_{pc}) in models 1D-A (location II) and 1D-B (monument K10). The perturbation method in PEST is similar to the UCODE method used in the three-dimensional model 3D-B. One feature of the perturbation method is that two separate models can be linked in the same regression; thus, eight observations of pressure at location II (model 1D-A) were combined with four observations of subsidence at monument K10 (model 1D-B) to obtain values of K_v , m, and H_{pc}. Weights w_i (see eq. 7) were assigned to the pressure and subsidence observations (1 ft and 10 ft, respectively) to account for differences in the measurement ranges (404 to 435 ft in pressure and 0.18 to 0.78 in subsidence). Parameters were estimated through a 12.4-year transient-state simulation representing precollapse and postcollapse conditions—the same period used in the three-dimensional simulations of water-level recovery discussed earlier.

The parameter values were estimated by the PEST regression were considered reliable, and the K_v-value was close to that specified in the three-dimensional flow model (table 12). The weighted residuals from models 1D-A and 1D-B do not indicate any unusual model bias and appear to be normally distributed. The individual confidence intervals given in table 12 are approximate, however, because the assumption of model linearity was not verified. Mean elastic specific-storage $\overline{S_{se}}$ values for the upper and lower confining layers were computed from the following relation

$$\overline{S_{se}} = \frac{1}{b}\sum S_{se(i)} b_i \text{ , } i = 1, n \qquad (29)$$

where b_i = 1-foot thickness of each cell,
b = confining-layer thickness,

$S_{se(i)}$ = elastic storage at depth i, and

n = number of cells representing the confining layer in the subsidence model.

The mean S_{se} value (2.3×10^{-6} ft^{-1}) compares well with that estimated for a confining layer composed of glacial drift in Anchorage, Alaska (2.3×10^{-6} ft^{-1}) from extensometer measurements during an aquifer test (Nelson, 1982). The preconsolidation head of 410 ft corresponds to a prior loading of about 150 ft of water at land surface and indicates that the confining-layer sediments are overconsolidated; this result is consistent with the presence of till in the upper confining layer and suggests that underlying sediments had previously undergone higher stresses during a temporary glacial advance than under the ambient conditions before the mine collapse. Further evidence of overconsolidation is apparent in the consolidation curves (fig. 38*B*), which indicate that samples from beneath the till had a lower void ratio e_0, and were much less compressible, than those obtained from shallower depths above the till. Specifying values of inelastic storage that were either 30 or 50 times the S_{se}-values resulted in little difference in model error.

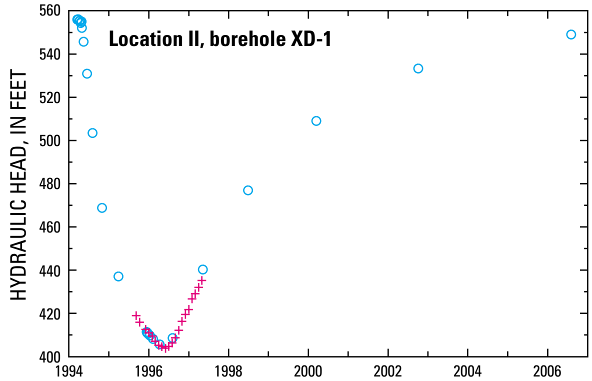
Computed pressure changes in the upper confining layer (model 1D-A) were in close agreement with pressures measured in borehole XD1 at location II during drainage of the confined aquifer system and the subsequent recovery once the mine was flooded in January 1996 (fig. 40). The maximum residual was about 5 ft, and the mean error was 2.1 ft, less than 2 percent of the computed 145-ft drawdown. The first pressure measurement was not obtained until September 1995, however; thus, the abrupt pressure declines that fol-

Table 12. Optimum parameter values estimated for confining layers In Genesee Valley aquifer system, Livingston County, N.Y., through nonlinear regression in one-dimensional models 1D-A (location II) and 1D-B (Monument K10), their approximate confidence intervals at 95-percent level, and the mean values specified in three-dimensional model 3D-C (entire aquifer system)

[ft, feet; —, not specified or not computed. Locations are shown in fig. 17.]

Variable	Value estimated in one-dimensional models 1D-A and 1D-B	Approximate individual confidence interval	Mean values specified in three-dimensional model 3D-C
Aquifer property			
Vertical hydraulic conductivity (feet per day)			
Upper confining layer	6.8×10^{-6}	4.2×10^{-6} - 1.1×10^{-5}	3.6×10^{-5}
Lower confining layer	1.2×10^{-3} [a]	—	1.2×10^{-3}
Mean elastic specific storage (ft^{-1})			
Upper confining layer	2.3×10^{-6}	1.4×10^{-6} - 4.1×10^{-6}	2.2×10^{-5}
Lower confining layer	1.1×10^{-6}	7.8×10^{-7} - 1.7×10^{-6}	7.9×10^{-6}
Preconsolidation head (feet)	410	370 - 458	—
Error in heads in model 1D-A at location II			
Sum of squared errors (feet squared)	35.4		
Standard error (feet)	2.1		
Subsidence at monument K10, as computed in model 1D-B (feet)	**Simulated subsidence**	**Measured subsidence**	
Upper confining layer	.06	—	.05
Lower confining layer	.67	—	.50
Total at monument K10	.73	.78	.55

[a]Fixed value in regression.



EXPLANATION

+ Measured

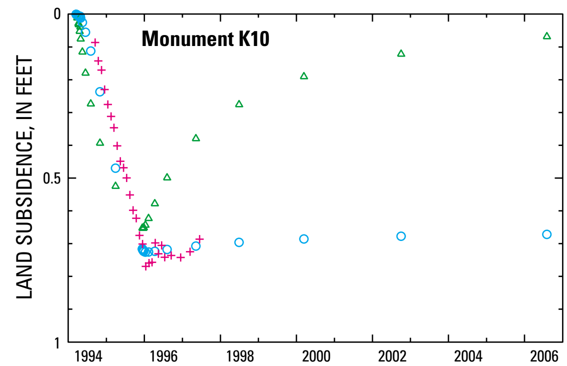
○ Computed

Figure 40. Pressures in upper confining layer at location II in Genesee Valley, Livingston County, N.Y. as computed by one-dimensional model 1D-A and measured at borehole XD-1. (Location is shown in fig. 17.)

lowed the collapses were not represented in the measured data.

The predicted subsidence at monument K10 (model 1D-B) closely matched the observed subsidence, and the maximum simulated subsidence (0.73 ft) was about 6 percent less than that observed (0.78 ft) (fig. 41*A*). About 90 percent of the simulated compression was inelastic, nonrecoverable compaction. The consistent downward trend in subsidence measurements along the K survey line suggests that the measurements from February 1995 through January 1996 were relatively accurate, but whether the land surface rebounded slightly once water levels began to recover in January 1996, as simulated in the model, is difficult to discern because the measured values have wide scatter.

The response of the confining layers to drainage and recovery of the aquifer system is illustrated by profiles of hydraulic head at K10 as a function of depth, as computed with model 1D-B (fig. 41*B*). Pressure declines propagated more rapidly through the lower confining layer than through the upper confining layer because the vertical hydraulic diffusivity ($\kappa_v = K_v/S_s$) of the lower unit is much greater (1,100 ft^2/d and 3.0 ft^2/d, respectively). Vertical hydraulic diffusivity governs one-dimensional propagation of pressure changes in confining layers and determines the attenuation and delay of the pressure response (for example, Keller and others, 1989; Yager and Kappel, 1998). About 92 percent of the computed subsidence (0.67 ft, table 12) was



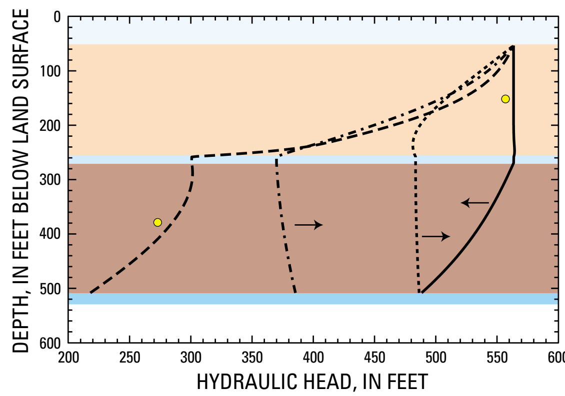
EXPLANATION

+ Measured

○ Elastic and inelastic specific storage

△ Elastic specific storage

A. Computed and measured subsidence.



EXPLANATION

Upper aquifer

Upper confining layer

Middle aquifer

Lower confining layer

Lower aquifer

Date and time since collapse

April 1994, 24 days

December 1995, 639 days

August 1996, 876 days

June 1998, 1,163 days

→ Direction of head change

Average head in three-dimensional model 3D-A, December 1995

B. Simulated hydraulic head as a function of depth in upper and lower confining layers, April 1994 through August 1996.

Figure 41. Results of one-dimensional model 1D-B simulations of land subsidence at monument K10 in Genesee Valley, Livingston County, N.Y.: ***A.*** Computed and measured subsidence. ***B.*** Simulated hydraulic head as a function of depth in upper and lower confining layers, April 1994 through August 1996. (Location is shown in fig. 17.)

the result of drainage and compression in the lower confining layer. Flow through the lower confining layer was nearly at steady state before the recovery of water levels in January 1996, as indicated by the approximately linear hydraulic gradient. The rate of flow through the upper confining layer was still increasing at this time, however, and drainage from the top of the upper confining layer continued through June 1998, long after water levels had recovered in the lower confining layer. A result of the relatively slow response of the upper confining layer to pressure changes was that sediments near the top of the layer were being compressed, whereas the underlying sediments near the bottom were expanding.

Elastic compressibility values computed from the m value of 2.4×10^{-4} ft/psi estimated by the regression were about 1 order of magnitude less than those calculated for confining layer sediments during the rebound phase of the consolidation tests (fig. 39). Model sensitivity to specific storage was investigated in two alternative regressions in which the observed subsidence was assumed to result solely from elastic compression. In both regressions, the K_v value was estimated, and m values were either estimated or specified (fig. 39). The S_{se} values obtained in these regressions (eq. 29) were larger than those computed in the initial regression, which represented both elastic and inelastic compression. Results of the two alternative regressions matched the pressure response observed at borehole XD-1 (model 1D-A) equally well, however, because estimated K_v-values were larger also than in the initial regression, so that the vertical hydraulic diffusivity was unchanged. This result indicates that a range of K_v and S_{se} values can be used to match the observed pressures in model 1D-A as long as the ratio of the values ($\kappa_v = K_v/S_s$) remains constant.

The alternative regression with the estimated m value of 6.5×10^{-3} ft/psi yielded a maximum subsidence of 0.66 ft at monument K10—about 85 percent of the measured value, and the elastic compressibilities matched reasonably well the values calculated for the rebound phase of consolidation tests (fig. 39). This simulation indicates that the land surface would rebound after water levels had begun to recover and the compressed sediments had expanded elastically—a result that is clearly unrealistic (fig. 41*A*). The other alternative regression—that with the specified m value of 5.5×10^{-2} ft/psi—resulted in a maximum subsidence of 5 ft, about 6 times greater than the measured value (fig. 42). These results suggest that compressibility values derived from consolidation tests do not accurately represent the actual compressibilities under field conditions.

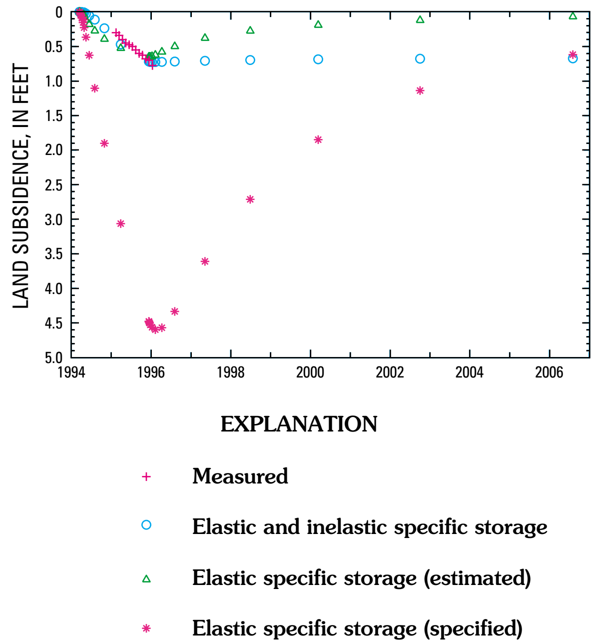


Figure 42. Land subsidence measured 1994-96 at monument K10 in Genesee Valley, Livingston County, N.Y., values computed by one-dimensional model 1D-B for 1994–2006, and values obtained from alternative regressions in which land subsidence was assumed to result only from elastic compression. (Location is shown in fig. 17.)

Three-Dimensional Simulation

The one-dimensional models 1D-A and 1D-B indicate that land subsidence south of the Retsof mine was the result of both elastic and inelastic compression of confining-layer sediments, and that two values are required for accurate representation of water released from storage during the observed water-level decline. The program MODFLOWP, which was used to construct three-dimensional model 3D-A, allows only one storage value for each model cell, however. The S_s-values specified in model 3D-A for the upper and lower confining layers (1.0×10^{-5} ft^{-1} and 5.0×10^{-6} ft^{-1}, respectively) were close to the mean $\overline{S}_{se}$-values computed by the alternative regression with models 1D-A and 1D-B, in which elastic compression was assumed (2.2×10^{-5} ft^{-1} and 7.9×10^{-6} ft^{-1}, respectively). In that regression, the close match between computed and maximum subsidence at monument K10 indicates that an approximate value for the volume of water released from storage in the confining layers can be obtained from a single specific-storage value. In addition, the water released from storage in the two confining layers was estimated to be less than 10 percent of the total water budget for the confined-aquifer system (table 8). Representing the effects of elastic and inelastic storage by a single value, therefore, probably does not greatly affect the hydraulic head distribution computed by three-dimensional model 3D-A.

Values of K_v and $\overline{S}_s$ that were obtained through regression with one-dimensional models 1D-A and 1D-B (assuming elastic compression) were incorporated into three-dimensional model 3D-C (tables 4 and 12) to compute the spatial distribution of subsidence when water-level drawdown was a maximum. Model 3D-C calculated the subsidence at each cell as the sum of the volume of water released from storage in each confining layer (model layers 2 and 4), divided by the horizontal cell area. Model 3D-C depicts only qualitatively the distribution of maximum subsidence, however, because (1) it used a single value to represent elastic and inelastic storage, and (2) it represents each confining layer by a single model layer, rather than by many 1-ft-thick layers as in models 1D-A and 1D-B. Vertical profiles of hydraulic head as a function of depth, as computed in model 1D-B (fig. 41*B*), indicate that the average head computed by the model 3D-C for the lower confining layer represents the vertical hydraulic gradient reasonably well, but the average head computed for the upper confining layer does not—the heads near the bottom of this unit, where most of the pressure decline occurred, were significantly overpredicted. As a result, model 3D-C underpredicted drainage from the upper confining layer and the subsequent compression of sediments (table 12). The maximum subsidence at monument K10 as computed by the three-dimensional model 3D-C (0.55 ft) is about 70 percent of that observed (0.8 ft). The cumulative subsidence in February 1996 along survey line K as computed by model 3D-C was reasonably consistent with the measured subsidence (fig. 43), although offset 3,000 ft to the east. The underprediction of maximum subsidence in model 3D-C is partly the result of representing the confining layers as single model layers, as discussed above. The 3,000-ft discrepancy in the location of maximum subsidence along survey line K suggests that the distribution of confining-layer thickness specified in model 3D-C does not accurately represent the actual thickness of fine-grained confining-layer sediments.

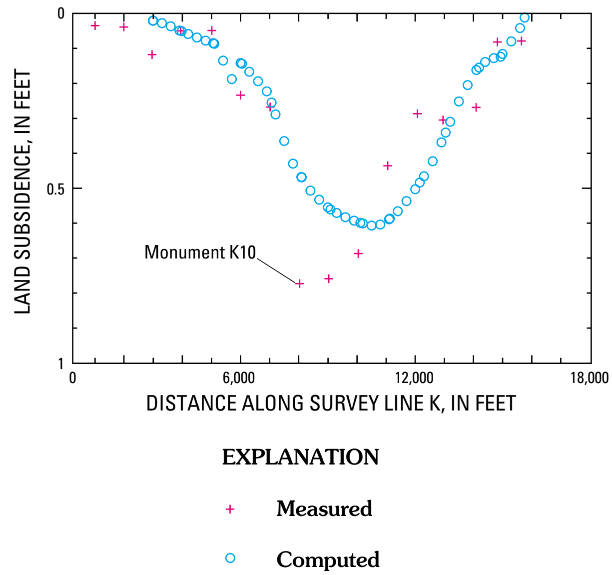


Figure 43. Land subsidence along survey line K in Genesee Valley, Livingston County, N.Y., as measured in February 1996 and as computed by three-dimensional model 3D-C. (Location is shown in fig. 17.)

Results of model 3D-C indicate that as much as 0.1 ft of subsidence had occurred by February 1996 over an area covering about 16 mi^2 that extended 5 mi north, and 7 mi south, of the collapse area (fig. 44) and that as much as 0.5 ft of subsidence occurred over an area covering about 1.4 mi^2. Simulated subsidence closely matched the measured subsidence in Mt. Morris, 3 mi south of the collapse, where as much as 0.3 ft of subsidence was measured. Subsidence was greatest near the collapse area and in the center of the Genesee Valley, where deposits of fine-grained sediments are thickest. A more accurate depiction of subsidence and potential rebound of the land surface could be obtained by incorporating the ISB1 package into three-dimensional model 3D-C to represent both elastic and inelastic storage.

SUMMARY

Collapses over two panels in the Retsof salt mine in the Genesee Valley in Livingston County, N.Y., in March and April 1994 allowed water from overlying aquifers to flood the mine. Subsidence of the overlying rock and sediment propagated to land surface, leaving two 300-ft-diameter sinkholes that damaged nearby structures, including a State highway bridge. Within 3 to 4 weeks of the collapses, water levels in about a dozen nearby domestic and industrial wells had declined severely, and some wells went dry. All mining

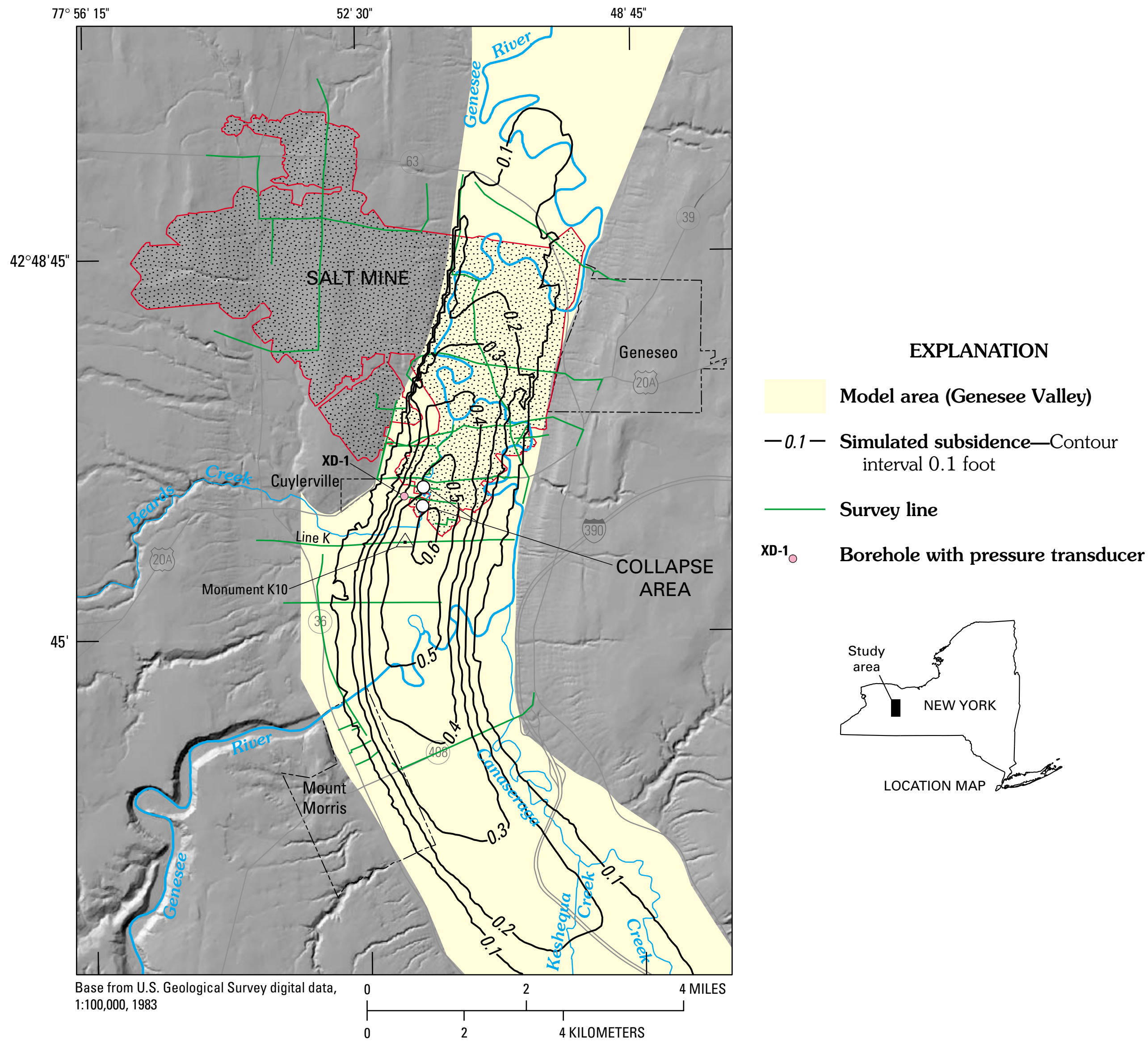


Figure 44. Distribution of land subsidence in Genesee Valley, Livingston County, N.Y. in February 1996, as computed by three-dimensional model 3D-C. (Location is shown in fig. 13.)

ceased in September 1995, and the mine was completely flooded by January 1996, by which time water levels had declined 400 ft near the collapse area and by more than 50 ft at wells 7 mi north and south of the collapse area.

Hydrogeologic Setting

The glacial aquifer system within the valley generally consists of three aquifers separated by two confining layers and is underlain by water-bearing zones in bedrock. The glacial aquifers are bounded laterally by the bedrock valley walls. The uppermost aquifer consists of alluvial sediments 20 to 60 ft thick; the middle aquifer consists of glaciofluvial sand and gravel less than 10 ft thick; and the lower aquifer consists of glaciofluvial sand and gravel about 25 ft thick overlying the bedrock valley floor. The mine lies 550 to 600 ft below the valley floor. The upper and middle aquifers are separated by an upper confining layer of lacustrine sediments and till as much as 250 ft thick, and the two confined aquifers are separated by a lower confining layer of undifferentiated glaciolacustrine sediments as much as 250 ft thick. The principal water-bearing zone in the bedrock overlying the mine consists of fractures near the contact between the Onondaga and Bertie

Limestones. The glacial aquifers are hydraulically connected at the edges of the confining layers and in subcrop zones, where water-bearing zones in the bedrock intersect the bedrock surface.

Ground water within the valley generally flowed northward prior to the mine collapse. The hydraulic head distribution in the confined aquifers under natural (precollapse) conditions is assumed to have been similar to that in the upper aquifer before the collapse, but water levels in the confined aquifers were probably above the water table beneath the valley floor. Ground water at the Onondaga/Bertie Limestone contact probably flowed northward to the Bertie Limestone subcrop in the valley north of the Fowlerville Moraine.

Water salinity in the middle aquifer is relatively low (430 to 1390 μS/cm); that in the lower aquifer ranges from 2,980 to 39,300 μS/cm, and that in the Onondaga and Bertie Limestones ranges from 72,000 to 106,000 μS/cm. The ratio of chloride to bromide concentrations in the confined aquifers is generally close to the ratio of 110 reported for brines from oil-bearing zones in the Bass Island carbonates in western New York. Ratios greater than 400 in some areas indicate, however, that halite (which has a chloride-to-bromide ratio greater than 1,000) has entered the lower aquifer as a result of solution mining operations or leakage through abandoned gas wells. The deuterium and ^{18}O content of waters from the two confined aquifers generally are similar to that of recharge in western New York, and concentrations of tritium suggest that the water is generally more than 45 years old. Water in the confined aquifers also contains biogenic gas (methane and hydrogen sulfide) that is probably derived from the Devonian shale bedrock.

Effects of Mine Collapse

The collapses and subsequent mine flooding caused land subsidence of 70 ft over the two collapsed panels; subsidence elsewhere ranged from 0.8 ft or less south of the mined area to as much 15 ft over the uncollapsed mined area. The subsidence south of the mine is attributed to compression of fine-grained sediments in the confining layers. Land subsidence also lowered the bed of the Genesee River by as much as 4.5 ft and altered the course of Beards Creek such that two ponds formed in the two sinkholes that developed over the collapses. Streamflow measurements in the Genesee River and Beards Creek did not indicate streamflow losses to the confined-aquifer system; this indicates that the water flowing into the mine did not originate as streamflow.

The collapses over the mine allowed water to drain from the lower aquifer into the mine at an altitude of 40 ft, which is 500 ft lower than the natural discharge area at the north end of the Genesee Valley. This drainage altered the hydraulic gradients in the lower aquifer, reversing the direction of flow north of the collapse area. Water levels in the lower aquifer had dropped as much as 400 ft by January 1996, when the mine was completely flooded, and about 15 domestic wells screened in the middle aquifer went dry. Drawdowns of 50 ft were recorded at wells 7 mi north of the collapse area, and drawdowns of 135 ft were recorded at wells 8 mi south of it. Water levels in upland areas and in most of the upper aquifer were unaffected by the collapse, although drawdowns of 5 ft in the upper aquifer may have resulted near the collapse area.

The reversal in the direction of ground-water flow north of the collapse appears to have caused intrusion of saline water into the lower aquifer at the north end of the valley. Salt concentrations of water from several domestic wells had become too high for human consumption within 12 months of the collapse. Ratios of chloride to bromide concentrations in ground-water samples indicate two possible sources of salinity in the lower aquifer—saline water from the Onondaga and Bertie Limestones, and halite-bearing water from an unknown source. The saline water probably flowed southward from areas near the Bertie subcrop toward the collapse area but also could have flowed upward from underlying saline-water-bearing zones in bedrock through high-angle fractures south of the Bertie subcrop under the vertical hydraulic gradient created by the drainage of water into the mine.

Declining water levels and the resulting decreases in pressure within the aquifer system caused exsolution of natural gas from water in the confined aquifers, and 2 or 3 water wells began to produce biogenic gas by July 1994. Thermogenic gas was detected within the mine and was flared from May 1995 until January 1996 through three wells drilled to the mine. Methane was detected near a plugged access shaft and around the mine perimeter at the northern end of the mine in January 1996, shortly after the mine was completely flooded, but the emissions gradually diminished through the summer 1996.

Simulation of Ground-Water Flow

Ground-water flow within the aquifer system was simulated by a three-dimensional model with MODFLOWP to represent flow conditions before and after the mine collapses. The three aquifers and two confining layers within the unconsolidated deposits were represented by five model layers. The model was calibrated to observed water levels and flow rates through nonlinear regression, and was used to calculate ground-water budgets and simulate the time required for water levels to return to precollapse conditions.

Parameter values representing aquifer properties were adjusted through steady-state simulations representing precollapse conditions and through transient-state simulations representing the flooding of the mine and water-level recovery after the mine was filled. The base flows and hydraulic-head distribution in the upper aquifer computed with the steady-state simulation matched the measured water levels and flows reasonably well, and the predicted directions of ground-water flow were consistent with the directions assumed for precollapse conditions. The standard error in heads was 8.7 ft, and computed base flows were nearly equal to the estimated base flow in Canaseraga Creek, but were only one-half the estimated base flow in the Genesee River. Optimum parameter values were obtained only for the recharge rate and hydraulic conductivity of the alluvial sediments. Neither of these values was precisely estimated because the numbers of head and flow observations were insufficient, and the sensitivity of the steady-state simulation to these measurements was limited as a result of the extensive boundaries specified in the top model layer (layer 1). Hydraulic heads computed with the steady-state simulation provided initial conditions for the transient-state simulations.

Transient-state simulations represented the 29 months during and after the collapse, including the period of drainage from the aquifer system to the mine (March 1994 through January 1996) and a period of water-level recovery after the mine had filled (January 1996 through August 1996). The computed distribution of drawdowns in January 1996 was similar to the measured distribution. The standard error in heads was 33 ft, and computed drawdowns near the collapse area (405 ft) were overpredicted by less than 10 ft, whereas drawdowns 7 mi to the north (35 ft) and 8 mi to the south (50 ft) were underpredicted by about 15 ft and 60 ft, respectively.

Computed discharges to the mine in April 1994 (20 ft^3/s) were 100 percent greater than the values estimated from the observed rate of mine flooding, and the computed discharges to the mine in September 1994 (28 ft^3/s) were 40 percent less than those estimated. The computed water budget indicated that ground water released from storage accounted for about 11 percent of the total inflow to the aquifer system and provided most of the water that discharged to the mine. Most of the drainage to the mine (58 percent) was from storage in the lower aquifer; another 15 percent was from storage from the rest of the confined system . The estimated volume of water discharged to the mine (1.1×10^9 ft^3) was only 55 to 60 percent of the mine volume, probably because drawdowns in the southern part of the aquifer system were underpredicted by as much as 60 ft.

Although the calibrated model represents the drawdown and recovery of water levels reasonably well, the simulated drawdowns in the lower aquifer were generally greater than those measured near the collapse area and less than those measured away from the collapse area. This bias indicates that the model does not accurately represent other unidentified processes or local variations in the hydraulic-conductivity distribution that affect ground-water flow. Specific storage values estimated for the lower and middle aquifers were 2.9×10^{-4} ft^{-1} and 6.9×10^{-5} ft^{-1}, respectively, much larger than the range of values (1×10^{-6} ft^{-1} to 3×10^{-6} ft^{-1}) estimated from studies of land subsidence in other regions underlain by sand and gravel aquifers. Assigning a lower value of specific storage (2.3×10^{-6} ft^{-1}) greatly increased model error, however, and no combination of the remaining parameter values was found through nonlinear regression that provided an acceptable match to the measured water levels.

Exsolution of natural gas from ground water during water-level declines lessened drawdown near the collapse area and slowed the propagation of drawdowns into areas away from the collapse, and probably accounts for the large values of specific storage estimated by regression and for part of the bias in the model. A gas-partitioning equation was developed to compute specific storage in the presence of a gas phase and indicated that specific storage was not constant, as assumed in the flow model, but varied temporally and spatially as a function of pressure. Including the effects of gas exsolution in MODFLOWP and specifying values of aquifer compressibility derived from studies of land subsidence decreased model error by about 15 percent and slightly reduced model bias.

Hydraulic heads predicted by the model indicate that water levels will return to precollapse conditions by about the year 2006. Specifying a lower specific storage value (2.3×10^{-6} ft^{-1}) for the lower aquifer speeded the simulated recovery and indicated that water levels would return to precollapse conditions by 1999. Increasing the conductance of cells representing vertical leakage through permeable deposits along the valley walls by 1 order of magnitude caused predicted water levels to return to precollapse conditions by the year 2002, whereas decreasing the conductance by that amount delayed the predicted recovery until after the year 2006.

Ground-water flow paths computed from steady-state and transient-state simulations with the model indicate that the mine collapse reversed the natural hydraulic gradient in the northern part of the valley, allowing intrusion of saline water into the lower aquifer. The estimated velocity of saline-water migration ranged from 1 to 8 ft/d, depending on the values of hydraulic conductivity, specific storage, and effective porosity specified for the lower aquifer. The most likely interpretation is that saline water traveled along preferential flow paths with a hydraulic conductivity of 2,000 ft/d and an effective porosity of 0.3. The density difference between the saline and fresh waters is relatively small and was found to have little effect on the direction of horizontal ground-water flow.

Saline water will probably continue to move southward toward the collapse area until about the year 2006, when water levels in the confined aquifers recover to precollapse conditions, and the northward hydraulic gradient becomes reestablished. The estimated magnitude of the northward gradient (3.9×10^{-4} ft/ft) is only one-third as large as the southward gradient in August 1996 (1.2×10^{-3} ft/ft); thus, the northward velocity of saline water will be only a third as rapid as the southward velocity. If saline water is assumed to migrate southward for 10 years after the collapse, about 30 years will be required for freshwater to flush the area into which the saline water has intruded.

Simulation of Land Subsidence

Water-level declines within the aquifer system that resulted from the mine collapse, and the accompanying increase in effective stress, caused compression of fine-grained sediments in the upper and lower confining layers and resulted in as much as 0.8 ft of land subsidence south of the mine. One-dimensional, transient-state simulations were conducted with MODFLOW and the ISB1 package to represent vertical flow and compression in the upper and lower confining layers at two locations. One model, which represented the collapse area, consisted of a column of 257 cells of unit volume (1 ft^3) to represent the upper confining layer at borehole XD1 containing pressure transducers, and a second model, which represented the upper and lower confining layers at survey monument K10 , about 2,800 ft south of the collapse, consisted of a column of 446 unit-volume cells. Boundaries in the upper aquifer and confined aquifers in both models were assigned hydraulic heads computed by the three-dimensional model, and subsidence was computed as the sum of the volume of water released from storage from each of the two confining layers.

Vertical hydraulic conductivity of the upper confining layer and specific storage of both confining layers were estimated through a nonlinear regression program (PEST), and the vertical hydraulic conductivity of the lower confining layer was fixed at the value estimated with the three-dimensional model (1.2×10^{-3} ft/d). Both elastic and inelastic storage were repesented in the one-dimensional models. Elastic storage was assumed, from the observed relation between compressibility and depth as derived from consolidation tests of 15 samples of confining-layer sediments, to be inversely proportional to effective stress (or depth). Inelastic storage was assumed, on the basis of published values, to be 30 times greater than elastic storage. The two models were linked in the same regression, and observations of pore pressure and subsidence were used to estimate model parameters.

Reliable values of vertical hydraulic conductivity, specific storage, and the preconsolidation head at which the transition from elastic to inelastic storage occurred were estimated through regression. Computed pressure changes in the upper confining layer were in close agreement with pressures measured with the transducers, and the maximum computed subsidence at monument K10 (0.73 ft) was only about 6 percent less than that observed. About 92 percent of the computed subsidence was the result of drainage and compression in the lower confining layer, of which 90 percent was inelastic or nonrecoverable compaction.

An alternative regression, in which subsidence was assumed to result solely from elastic compression, resulted in larger estimates of vertical hydraulic conductivity and specific storage than in the regression in which inelastic storage was assumed. Results of the alternative regression matched the observed pressure response and maximum subsidence reasonably well but indicated that the land surface would rebound during water-level recovery. The observed absence of land-surface rebound at monument K10 indicates that subsidence resulted mainly from inelastic compaction.

Values of vertical hydraulic conductivity and specific storage of confining-layer sediments estimated on the assumption of elastic compression in one-dimensional models were incorporated into the three-dimensional model, in which each confining layer was represented by a single model layer. This approximation had little effect on the simluated drainage and compression in the lower confining layer, which had a large vertical hydraulic diffusivity (150 ft^2/d) and responded relatively quickly to drawdown. Compression in the upper-confining layer was underpredicted, however, because this unit had a lower vertical hydraulic diffusivity (1.6 ft^2/d) and responded more slowly to drawdown.

The cumulative subsidence in February 1996 along survey line K, as computed by the three-dimensional model, compared favorably with the measured subsidence. Simulated subsidence closely matched the measured subsidence in Mt. Morris, 3 mi south of the collapse, where as much as 0.3 ft of subsidence was measured. As much as 0.1 ft of the simulated subsidence had occurred by February 1996 over an area covering about 16 mi^2 that extended 5 mi north of, and 7 mi south of, the collapse area. Subsidence was greatest near the collapse area and in the center of the valley, where deposits of fine-grained sediment are thickest.

REFERENCES CITED

Acres International, 1986, Sterling Mine B shaft—Geotechnical summary report: Buffalo, N.Y., 20 p. (DEC–906).

Akzo Nobel Salt, Inc., 1994–97, Retsof Mine subsidence monitoring: Clarks Summit, Pa., Akzo Nobel Salt, Inc., unpublished data on file at Geneseo Public Library, Geneseo, N.Y.

———1996, Surface and groundwater impacts report: Clarks Summit, Pa., Akzo Nobel Salt, Inc., 20 p.

Alpha Geoscience, 1995a, Well 9446 report: Albany, N.Y., Alpha Geoscience, 8 p.

———1995b, Well 9449 report: Albany, N.Y., Alpha Geoscience, 8 p.

———1995c, Well 9460 report: Albany, N.Y., Alpha Geoscience, 12 p.

———1996a, Well 9565 report: Albany, N.Y., Alpha Geoscience, 9 p.

———1996b, Well 9569 report: Albany, N.Y., Alpha Geoscience, 8 p.

———1996c, Geologic and hydrogeologic investigation of the Genesee River valley: Albany, N.Y., Alpha Geoscience, 31 p.

Baird, G.C., 1979, Sedimentary relationships of Portland Point and associated middle Devonian rocks in central and western New York: New York State Museum Bulletin 433, 14 p.

Bradley, W.H., and Pepper, J.F. 1938, Geologic structure and occurrence of gas in part of southwestern New York, part 1: U.S. Geological Survey Bulletin 899A, 68 p.

Brennan, W.J., ed., 1988, Lake Wisconsinan deglaciation of the Genesee valley: Geneseo, State University of New York, Department of Geological Sciences, Guidebook for 51st Annual Meeting, Friends of the Pleistocene, p. 29–56.

Butler, S.S., 1957, Engineering hydrology: Englewood Cliffs, N.J., Prentice-Hall, 250 p.

Coats, K.H., and Richardson, J.G., 1967, Calculation of water displacement by gas in development of aquifer storage: Society of Petroleum Engineers Journal, p. 105–112.

Cooley, R.L., and Naff, R.C., 1990, Regression modeling of ground-water flow: U.S. Geological Survey Techniques of Water-Resources Investigations Report, Book 3, Chapter B4, 232 p.

Cooper, G.A., 1930, Stratigraphy of the Hamilton Group of New York: American Journal of Science, v. 19, p. 116–134, 214–236.

Coplen, T.B. 1994. Reporting of stable hydrogen, carbon, and oxygen isotopic abundances: Pure and Applied Chemistry v. 66, p. 273–276.

Coplen, T.B., Wildman, J.D., and Chen, J. 1991. Improvements in the gaseous hydrogen-water equilibration technique for hydrogen isotope ratio analysis: Analytical Chemistry v. 63, p. 910–912.

Czolbe, Peter, 1997, Gas blocking in gas storage at simultaneous gas and water flow: American Association of Petroleum Geologists Bulletin, v. 81, p. 1,368–1,369.

Davies, P. B., 1989, Variable-density ground-water flow and paleohydrology in the Waste Isolation Pilot Plant (WIPP) region, southeastern New Mexico: U.S. Geological Survey Open-File Report 88–490, 139 p.

Davis, S.N., Whittemore, D. O., and Fabryka-Martin, J., 1998, Uses of chloride/bromide ratios in studies of potable water: Ground Water, v. 36, no. 2, p. 338–350.

Denk, J.M., Jansky, J.J., Hoch, M.T., Karabin, G.J., and Kirkwood, D. T., 1994, Accident investigation report, underground nonmetal mine, nonfatal roof fall/flooding, Akzo Nobel Salt, Inc.: U.S. Department of Labor,

Mine Safety and Health Administration, Cranberry Township, Pa., 20 p.
Dobecki Earth Sciences, Inc., 1994, Phase III seismic survey above 11 West panel, Akzo Nobel Retsof mine: Houston, Texas, Dobecki Earth Sciences, Inc., 20 p.
Doherty, John, Brebber, Lindsay, and Whyte, Peter, 1994, PEST—Model-independent parameter estimation: Ipswich, Australia, Watermark Computing, 135 p.
Dunn Geoscience Corporation, 1992, Hydrogeologic Report for the AKZO ash processing plant: Albany, N.Y., Dunn Geoscience Corporation, 44 p.
Epstein, S. and Mayeda, T.. 1953. Variation of O–18 content of water from natural sources: Geochimica Cosmologica Acta, v. 4, p. 213-224.
Freeze, R.A., and Cherry, J.A., 1979, Groundwater: Englewood Cliffs, N.J., Prentice-Hall, 604 p.
H&A of New York Consulting Engineers, 1988, Report of geologic and hydrogeologic conditions adjacent to the Retsof salt mine, Livingston County, New York: Rochester, N.Y., H&A of New York Consulting Engineers, 28 p.
Haeni, F.P., 1986, Application of continuous seismic reflection methods to hydrologic studies: Ground Water, v. 24, no. 1, p. 23–312
———1988, Application of seismic-refraction techniques to hydrologic studies: U.S. Geological Survey Techniques of Water-Resources Investigations, book 2, chap. D2, 86 p.
Hill, M.C., 1992, A computer program (MODFLOWP) for estimating parameters of a transient, three-dimensional ground-water flow model using nonlinear regression: U.S. Geological Survey Open-File Report 91–484, 382 p.
Ireland, R.L., Poland, J.F., and Riley, F.S., 1984, Land subsidence in the San Joaquin Valley, California as of 1980: U.S. Geological Survey Professional Paper 437–I, 93 p.
Isachsen, Y.P. and McKendree, W.G., 1977, Preliminary brittle structures map of New York: New York State Geological Survey, Map and Chart Series 31E, 1 sheet, scale 1:.
John T. Boyd Company, 1995, Impact analysis—Retsof mine, Akzo Nobel Salt, Inc.: Pittsburgh, Pa., John T. Boyd Company, 50 p.
Johnson, A.I., 1984, Laboratory tests for sediments of subsiding areas, *in* Poland, J.F. (ed.), Guidebook to studies of land subsidence due to ground-water withdrawal: Paris [France], United Nations Educational, Scientific and Cultural Organization Studies and Reports in Hydrology, no. 40, 305 p.
Jorgensen, D.G., 1980, Relationships between basic soils-engineering equations and basic ground-water flow equations: U.S. Geological Survey Water-Supply Paper 2064, 40 p.
Kammerer, J.C., and Hobba, W.A., 1967, The geology and availability of ground water in the Genesee River basin, New York and Pennsylvannia, Appendix I, Ground-water resources of the Genesee River basin: Albany, N.Y., New York State Water Resources Commission, 101 p.
Keller, C.K., van der Kamp, G., and Cherry, J.A., 1989, A multiscale study of the permeability of a thick clayey till: Water Resources Research, v. 25 no. 11, p. 2,299–2,318.
La Sala, A.M., 1968, Ground-water resources of the Erie-Niagara basin, New York: Albany, N.Y., New York State Water Resources Commission, Basin Planning Rerport ENB–3, 114 p.
Langill, R.F. and Associates, 1981, Geological and hydrological conditions at the Sterling mine near Cuylerville, New York: Germantown, Md., Langill, R. F. and Associates, 20 p. (DEC–922).
Leake, S.A., and Prudic, D.E., 1991, Documentation of a computer program to simulate aquifer-system compaction using the modular finite-difference ground-water flow model: U.S. Geological Survey: Techniques of Water-Resources Investigations, Book 6, chapter A2, 68 p.
Lekvam, K., and Bishnoi, P.R., 1997, Dissolution of methane in water at low temperatures and intermediate pressures: Fluid Phase Equilibria, v. 131, no.1–2, p. 297–309.
Lyford, F.P., and Cohen, A.J., 1988, Estimation of water available for recharge to sand and gravel aquifers in the glaciated northeastern United States, *in* Randall, A.D., and Johnson, A.I., eds., Regional Aquifer Systems of the United States, the Northeast glacial aquifers: American Water Resources Association, AWRA monograph Series, no. 11, p. 37–62.
MacNish, R.D., and Randall, A.D., 1982, Stratified drift aquifers in the Susquehanna River basin, New York: New York State Department of Environmental Conservation Bulletin 75, 68 p.
Mayer, S.M., 1994, Field studies of the middle Devonian Ludlowville-Moscow sequence in the Genesee and Seneca valleys, New York, *in* Brett, C.E., and Scatterday, James (eds.), Field trip guidebook, 66th annual meeting: Rochester, New York State Geological Association, p. 491–503.
McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water-Resources Investigations, Book 6, Chap. A1.
Michel, R.L., 1989, Tritium deposition in the continental United States, 1953–83: U.S. Geological Survey: Water-Resources Investigations Report 89–4072, 46 p.
Morrissey, D.J., Randall, A.D., and Williams, J.H., 1988, Upland runoff as a major source of recharge to stratified drift in the glaciated Northeast: *in* Randall, A.D., and Johnson, A.I., eds., Regional Aquifer Systems of the United States, the Northeast glacial aquifers: American

Water Resources Association, AWRA monograph Series, no. 11, p. 17–36.
Muller, E.H., Braun, D.D., Young, R.A., and Wilson, M.P., 1988, Morphogenesis of the Genesee Valley: Northeastern Geology, v. 10, no. 2, p. 112–133.
Mullins, H.T., Wellner, R.W., Petruccione, J.L., Hinchey, E.J., and Wanzer, Steven, 1991, Subsurface Geology of the Finger Lakes Region, *in* New York State Geological Association Field Trip Guidebook: 63rd annual meeting, p. 1–54.
Musetter Engineering, Inc., 1996, Evaluation of potential effects of subsidence on the Genesee River and tributaries: Fort Collins, Colo., Musetter Engineering, Inc., 100 p.
Nelson, G.L., 1982, Vertical movement of ground water under the Merrill field landfill, Anchorage, Alaska: U.S. Geological Survey Open-File Report 82–1016, 24 p.
New York State Department of Environmental Conservation, 1992, Final generic environmental impact statement on the oil, gas and solution mining regulatory program: Albany, N.Y., Division of Mineral Resources, 230 p.
———1997, Collapse and flooding of Akzo Nobel's Retsof salt mine, Livingston County, New York: Albany, N.Y., Division of Mineral Resources, 113 p.
Newland, D.H., and Leighton, Henry, 1910, Gypsum deposits of New York: New York State Museum Bulletin 143, 94 p.
Neuzil, C.E., 1986, Groundwater flow in low-permeability environments: Water Resources Research, v. 22, no. 8, p. 1,163–1,195.
Nittany Geoscience, 1995, Study of groundwater quality in the area of Retsof, New York: State College, Pa., Nittany Geoscience, 18 p.
Poeter, E.P., and Hill, M.C., 1998, Documentation of UCODE, a computer code for universal inverse modeling: U.S. Geological Survey Water-Resources Investigations Report 98–4080, 116 p.
Pollock, D.W., 1989, Documentation of computer programs to compute and display pathlines using results from the U.S. Geological Survey modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 89–381, 188 p.
Randall, A.D., 1996, Mean annual runoff, precipitation, and evapotranspiration in the glaciated northeastern United States, 1951–80: U.S. Geological Survey Open-File Report 96–395, 2 pl., scale 1:l,000,000.
Rantz, S.E., and others, 1982, Measurement and computation of streamflow, volume 1, measurement of stage and discharge, U.S. Geological Survey Water Supply Paper 2175, 284 p.
Reynolds, R.J., and Williams, J.H., 1987, Continuous seismic-reflection profiling of glacial drift along the Susquehanna, Chemung, and Chenango Rivers, South-Central New York and North-Central Pennsylvania, *in* Randall, A.D., and Johnson, A.I., (eds.) Regional aquifer systems of the United States- the Northeast glacial aquifers: American Water Resources Association, AWRA monograph series no. 11, p. 83–103.
Rickard, L.V., 1969, Stratigraphy of the upper Silurian Salina Group of New York, Pennsylvannia, Ohio, and Ontario: New York State Museum, Map and Chart Series 12, 57 p.
———1975, Correlations of Silurian and Devonian rocks in New York State: New York State Museum, Map and Chart Series 24, 16 p.
Riley, F. S., 1998, Mechanics of aquifer systems—the scientific legacy of Joseph F. Poland, *in* Land subsidence—case histories and current research: Proceedings of the Dr. Joseph F. Poland Symposium on land subsidence, Borchers, J.W., (ed.): Belmont Ca., Star Publishing Co., p. 13–27.
Schwarzenbach, P.M., Gschwend, P.M., and Imboden, D.M., 1993, Environmental Organic Chemistry: New York, John Wiley, 245 p.
Scott, J.H., Tibbetts, B.L., and Burdick, R.G., 1972, Computer analysis of seismic refraction data: U.S. Bureau of Mines Report of Investigation 7595, 95 p.
Shannon and Wilson, Inc., 1995, Task 1 Report, Retsof mine collapse, Retsof, New York: Seattle, Wash., Shannon and Wilson, Inc., 20 p.
Solley, W.B., Pierce, R. R., and Perlman, H.A., 1993, Estimated use of water in the United States, 1990: U.S. Geological Survey Circular 1081, 76 p.
Staubitz, W.W. and Miller, T.S., 1987, Geology and hydrology of the Onondaga Aquifer in Eastern Erie County, with emphasis on ground-water-level declines since 1982: U.S.Geological Survey Water-Resources Investigations Report 86-4317, 44 p.
Todd. D.K., 1980, Groundwater hydrology: New York, John Wiley, 535 p.
U.S.Geological Survey, 1996, U.S. GeoData digital elevation models: U.S.Geological Survey Fact Sheet FS–102–96, 2 p.
Van Sambeek, L.L., 1994, Predicted ground settlement over the Akzo Nobel Retsof mine: Rapid City, S.D., RE/SPEC, Inc., 33 p.
Whittemore, D.O., 1988, Bromide as a tracer in groundwater studies—geochemistry and analytical determination, *in* Proceedings of the Groundwater Geochemistry Conference, Denver, Colo., National Ground Water Asoociation, p. 339–360.
Williams, J.H., 1996, Application of borehole geophysics at the Retsof salt mine collapse site, western New York, *in* Symposium on the Application of Geophysics to Engineering and Environmental Problems, Keystone, Colo., April 28–May 1, 1996, Proceedings: Englewood, Colo., Environmental and Engineering Geophysical Society, p. 813–821.
Yager, R.M., and Kappel, W.M., 1998, Infiltration and hydraulic connections from the Niagara River to a frac-

tured-dolomite aquifer in Niagara Falls, New York: Journal of Hydrology, v. 206, p. 84–97

Young, R.A., 1975, The effects of a Late Wisconsin glacial readvance on the postglacial geology of the Genesee Valley, Livingston County, New York, *in* Geological Society of America, Abstracts with Programs, v. 7, p. 135–136.

Young, R. A., and Sitkin, L., 1994, Subsurface geology of lower Genesee River valley region: a progress report on the evidence for middle Wisconsin sediments and implications for ice sheet erosion models, *in* New York State Geological Association field trip guidebook, 66th Annual meeting, Rochester, N.Y., p. 89–125.

APPENDIX

Appendix. Wells in the monitroing network in the Genesee Valley, Livingston County, N.Y., 1997

[Locations are shown in fig. 4. x; information collected]

Well number								
USGS	Local	Owner	Land-surface elevation (feet above sea level)	Well depth (feet below land surface)	Affected by mine flooding?	Water-level hydrograph	Water sample	Aquifer
Lv91[a]	18	Barber-Ag	569.6	329	yes	x		lower
Lv203[b]	—	Croston	565.	120	unknown		x	lower
Lv301[a]	—	ANSI-9401	556.	1,113	yes		x	Retsof mine
Lv309[a]	—	ANSI-9409	554.	877	yes	x		Onondaga/Bertie Limestone
Lv340[a]	40	ANSI-9440	565.0	366	yes	x		bedrock
Lv346	88	ANSI-9446(II)[c]	559.1	499	yes	x	x	Onondaga Lime-stone
Lv348	101	ANSI-9448(II)[c]	559.5	383	yes	x		middle
Lv349	89	ANSI-9449(III)[c]	567.4	679	yes	x	x	Onondaga/Bertie Limestones
Lv350	90	ANSI-9450(III)[c]	565.8	358	yes	x	x	lower
Lv351	91	ANSI-9451(III)[c]	567.2	254	yes	x	x	middle
Lv360	52	ANSI-9460(IV)[c]	566.8	545	yes	x	x	lower
Lv361	102	ANSI-9461(IV)[c]	566.9	323	yes	x	x	middle
Lv362	81	ANSI-9462(IV)[c]	566.9	260	yes	x	x	middle
Lv365	104	ANSI-9565(V)[c]	573.1	910	no	x	x	Onondaga/Bertie Limestones
Lv366	105	ANSI-9566(V)[c]	572.9	448	yes	x	x	middle
Lv367	106	ANSI-9567(V)[c]	573.2	750	maybe	x	x	lower
Lv368	103	ANSI-9568(II)[c]	559.5	480	yes	x	x	lower
Lv369	107	ANSI-9569(I)[c]	599.9	257	maybe	x	x	Onondaga/Bertie Limestones
Lv370	108	ANSI-9570(I)[c]	599.8	207	maybe	x	x	lower and Onon-daga Limestone
Lv371	109	ANSI-9571(I)[c]	601.0	100	maybe	x	x	middle
Lv396	—	ANSI-9572	610.	1170.	yes	x		Retsof mine
Lv397[a]	75	ANSI-7901	560.0	505	yes	x		lower
Lv398	—	National Hotel	575.	132	yes			middle?
Lv399	—	Tubbs	570.	437	unknown			middle?
Lv401	1	Rolison	622.0	72	dry	x		middle
Lv402	2	Gregg	626.6	97	yes	x		middle
Lv403	3	Rugatuso	611.7	72	dry	x		middle
Lv404	4	Lovelace	643.3	13	no	x		shale bedrock
Lv405	5	Baumgardner	565.0	12	no	x		upper
Lv406	6	American Legion	564.6	32	maybe	x		upper
Lv407	7	Christiano	562.1	32	maybe	x		upper
Lv408	8	Newman	666.4	140	no	x		middle?

Well number		Owner	Land-surface elevation (feet above sea level)	Well depth (feet below land surface)	Affected by mine flooding?	Water-level hydrograph	Water sample	Aquifer
USGS	Local							
Lv409	9	Newman	675.5	116	no	x		middle?
Lv410	10	Farruggia	743.3	110	no	x		shale bedrock
Lv411	11	Hull	685.3	21	no	x		unknown
Lv413	13	Pendergast	760.	96	no	x		shale bedrock
Lv414	14	Barnhardt	628.2	78	dry	x		middle
Lv415	15	Werner	621.4	64	dry	x		middle
Lv416	16	Arend	574.8	24	dry	x		middle
Lv417	17	Falconer	639.1	42	no	x		upper?
Lv419	19	Montemarano	590.	35	no	x		upper
Lv420	20	Cipriano	580.	33	no	x		upper
Lv421[a]	21	Atochem	550.	130	yes	x		lower?
Lv422[a]	22	Atochem	550.	255	well plugged	x		unknown
Lv423	23	Bailey	610.	82	dry	x		middle
Lv424	24	Hardy	630.	115	yes	x		middle
Lv425	25	Hollinger	580.	173	yes	x	x	lower
Lv426	26	Davis #1	580.	58	dry	x		middle
Lv427	27	Banker	590.	193	yes	x		lower
Lv428	28	Koshchara	637.	54	dry	x		unknown
Lv429	29	Bryan	720.	112	no	x		shale bedrock
Lv430	30	Lloyd	645.	73	no	x		shale bedrock?
Lv431	31	Yull	755.	46	no	x		shale bedrock?
Lv432	32	Geary	830.	72	no	x		shale bedrock
Lv433	33	Scatterday	760.	99	no	x		shale bedrock
Lv434	34	private	680.	100	no	x		shale bedrock
Lv435	35	McCracken	640.	160	yes	x		middle?
Lv436	36	McCauley	685.	121	yes	x		lower ?
Lv437	37	Barrett	590.	103	dry	x		middle
Lv438	38	Colahan	690.	99	dry	x		middle
Lv439[a]	39	Cofield	600.	96	dry	x		middle
Lv441	41	private	560.	315	no	x		unknown
Lv442[a]	42	Millard	635.	166	yes	x	x	middle?
Lv443	43	Cullinan	650.	103	yes	x	x	middle
Lv444	44	Romano	670.	73	yes	x		middle
Lv445	45	Bucci	940.	125	no	x		shale bedrock
Lv446	46	Clinton	820.	295	no	x		shale bedrock
Lv447	47	Arrigenna	720.	90	no	x		unknown
Lv448	48	Jerris	660.	101	yes	x		middle

Well number		Owner	Land-surface elevation (feet above sea level)	Well depth (feet below land surface)	Affected by mine flooding?	Water-level hydrograph	Water sample	Aquifer
USGS	Local							
Lv449	49	Vozzies	910.	59	no	x		shale bedrock
Lv450	50	Morabito	585.	80	no	x		shale bedrock
Lv451	51	Dueppengisser	1,050.	90	no	x		shale bedrock
Lv453	53	Susz	670.	74	maybe	x		middle
Lv454	54	Holt/Marsh	610.	205	yes	x	x	lower
Lv455	55	private	620.	119	no	x		unknown
Lv456	56	Yull	755.	120	no	x		shale bedrock
Lv457[a]	57	Davis #2	580.	163	yes		x	lower
Lv458	58	Crandall	565.	65	maybe	x	x	middle
Lv459[a]	59	Stone	580.	120	yes	x		lower
Lv460	60	O'Neil	600.	51	no	x		Onondaga Lime-stone
Lv461	61	Wadsworth	620.	221	yes	x	x	lower
Lv462	62	Preston	655.	80	yes	x		middle
Lv463	63	Castle	720.	59	no	x		shale bedrock?
Lv464	64	Seager	770.	40	no	x		shale bedrock?
Lv465	65	Ambourgey	935.	23	no	x		shale bedrock
Lv466	66	Ambourgey	935.	23	no	x		shale bedrock
Lv467	67	Love	810.	67	no	x		shale bedrock
Lv468	68	Quakenbush	735.	47	no	x		shale bedrock?
Lv469	69	Orologio, V.	580.	137	yes	x	x	lower
Lv470	70	Amorese	935.	127	no	x		shale bedrock
Lv471[a]	71	Barrett	590.	199	yes	x		lower
Lv472	72	Waide	1,140.	33	no	x		shale bedrock
Lv473	73	Olea	770.	17	no	x		unknown
Lv474	74	Olea	770.	51	no	x		unknown
Lv476	76	Barefoot	700.	68	no	x		shale bedrock
Lv477	77	Brassie	1,140.	87	no	x		shale bedrock
Lv478[a]	78	Fuller	610.	110	yes	x		middle
Lv479	79	Ward	580.	150	yes	x	x	lower
Lv480[a]	80	Cofield	608.	208	yes	x		lower
Lv482	82	Wall	590.	36	yes	x		middle
Lv483	83	Carlson	610.	127	yes	x		middle
Lv484	84	Sanderson	610.	53	no	x		shale bedrock
Lv485	85	Young/Thorn	675.	126	maybe	x		unknown
Lv486	86	Wiley/Viola	630.	95	yes	x		middle
Lv487	87	Kane	685.	114	yes	x		lower?
Lv493	93	Kelly	600.	53	yes	x	x	middle

Well number		Owner	Land-surface elevation (feet above sea level)	Well depth (feet below land surface)	Affected by mine flooding?	Water-level hydrograph	Water sample	Aquifer
USGS	Local							
Lv494	94	Viola	610.	154	yes	x		lower
Lv496[a]	96	Carlson	610.	212	yes	x		lower
Lv498[a]	98	Davis #3	580.	174	yes	x	x	lower
Lv500	100	Baker/Wadsworth	600.	219	yes	x	x	lower
Lv510	—	Jones (house)	610.	215	yes		x	lower
Lv511	—	Jones (barn)	610.	225	yes		x	lower
Lv512	—	Halpin	580.	167	yes		x	lower
Lv513	—	McKeown	575.	85	yes		x	middle
Lv514	—	Banker	590.	193	yes		x	lower
Lv515	—	Klimasewski	580	134	unknown		x	lower
Lv516	—	Zander	580.	135	unknown		x	lower
Lv517	—	Sullivan	585.	144	unknown		x	lower
Lv518	—	Orologio, B.	580.	81	unknown		x	middle?
Lv519	—	Elliot	600.	219	yes		x	lower
Lv520	—	Morss	750.	—	unknown		x	bedrock spring
Lv521	—	Harris	610.	175	unknown		x	lower
Lv522	—	O'Connell	575.	—	unknown		x	lower?
Lv523	—	McMahon	580.	—	unknown		x	middle?
Lv524	—	Striker	580	—	unknown		x	middle?
Lv525	—	House	555.	spring	unknown		x	middle?

[a] Log available from U.S. Geological Survey, Ithaca, N.Y.
[b] Log available from Kammerer and Hobba, 1967.
[c] Azko-Nobel Salt, Inc. company number (cluster location); log available from Alpha Geoscience, 1996c.

Selected Series of U.S. Geological Survey Publications

Books and Other Publications

Professional Papers report scientific data and interpretations of lasting scientific interest that cover all facets of USGS investigations and research.

Bulletins contain significant data and interpretations that are of lasting scientific interest but are generally more limited in scope or geographic coverage than Professional Papers.

Water-Supply Papers are comprehensive reports that present significant interpretive results of hydrologic investigations of wide interest to professional geologists, hydrologists, and engineers. The series covers investigations in all phases of hydrology, including hydrogeology, availability of water, quality of water, and use of water.

Circulars are reports of programmatic or scientific information of an ephemeral nature; many present important scientific information of wide popular interest. Circulars are distributed at no cost to the public.

Fact Sheets communicate a wide variety of timely information on USGS programs, projects, and research. They commonly address issues of public interest. Fact Sheets generally are two or four pages long and are distributed at no cost to the public.

Reports in the **Digital Data Series (DDS)** distribute large amounts of data through digital media, including compact disc-read-only memory (CD-ROM). They are high-quality, interpretive publications designed as self-contained packages for viewing and interpreting data and typically contain data sets, software to view the data, and explanatory text.

Water-Resources Investigations Reports are papers of an interpretive nature made available to the public outside the formal USGS publications series. Copies are produced on request (unlike formal USGS publications) and are also available for public inspection at depositories indicated in USGS catalogs.

Open-File Reports can consist of basic data, preliminary reports, and a wide range of scientific documents on USGS investigations. Open-File Reports are designed for fast release and are available for public consultation at depositories.

Maps

Geologic Quadrangle Maps (GQ's) are multicolor geologic maps on topographic bases in 7.5- or 15-minute quadrangle formats (scales mainly 1:24,000 or 1:62,500) showing bedrock, surficial, or engineering geology. Maps generally include brief texts; some maps include structure and columnar sections only.

Geophysical Investigations Maps (GP's) are on topographic or planimetric bases at various scales. They show results of geophysical investigations using gravity, magnetic, seismic, or radioactivity surveys, which provide data on subsurface structures that are of economic or geologic significance.

Miscellaneous Investigations Series Maps or **Geologic Investigations Series (I's)** are on planimetric or topographic bases at various scales; they present a wide variety of format and subject matter. The series also incudes 7.5-minute quadrangle photogeologic maps on planimetric bases and planetary maps.

Information Periodicals

Metal Industry Indicators (MII's) is a free monthly newsletter that analyzes and forecasts the economic health of five metal industries with composite leading and coincident indexes: primary metals, steel, copper, primary and secondary aluminum, and aluminum mill products.

Mineral Industry Surveys (MIS's) are free periodic statistical and economic reports designed to provide timely statistical data on production, distribution, stocks, and consumption of significant mineral commodities. The surveys are issued monthly, quarterly, annually, or at other regular intervals, depending on the need for current data. The MIS's are published by commodity as well as by State. A series of international MIS's is also available.

Published on an annual basis, **Mineral Commodity Summaries** is the earliest Government publication to furnish estimates covering nonfuel mineral industry data. Data sheets contain information on the domestic industry structure, Government programs, tariffs, and 5-year salient statistics for more than 90 individual minerals and materials.

The Minerals Yearbook discusses the performance of the worldwide minerals and materials industry during a calendar year, and it provides background information to assist in interpreting that performance. The Minerals Yearbook consists of three volumes. Volume I, Metals and Minerals, contains chapters about virtually all metallic and industrial mineral commodities important to the U.S. economy. Volume II, Area Reports: Domestic, contains a chapter on the minerals industry of each of the 50 States and Puerto Rico and the Administered Islands. Volume III, Area Reports: International, is published as four separate reports. These reports collectively contain the latest available mineral data on more than 190 foreign countries and discuss the importance of minerals to the economies of these nations and the United States.

Permanent Catalogs

"Publications of the U.S. Geological Survey, 1879–1961" and **"Publications of the U.S. Geological Survey, 1962–1970"** are available in paperback book form and as a set of microfiche.

"Publications of the U.S. Geological Survey, 1971–1981" is available in paperback book form (two volumes, publications listing and index) and as a set of microfiche.

Annual supplements for 1982, 1983, 1984, 1985, 1986, and subsequent years are available in paperback book form.

USGS

Yager, Miller, and Kappel—Simulated Effects of 1994 Salt-Mine Collapse on Ground-Water Flow and Land Subsidence in a Glacial Aquifer System, Livingston County, New York—PP 1611

ISBN 0-607-95033-1

9 790607 950334

Printed on recycled paper