SIMULATION OF GROUND-WATER FLOW IN AQUIFERS ALONG THE SUSQUEHANNA RIVER IN COLUMBIA COUNTY, PENNSYLVANIA

By John H. Williams and Gregory E. Senko

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CONVERSION FACTORS AND ABBREVIATIONS

For the convenience of readers who may prefer to use metric (International System) units rather than the inch-pound units used in this report, values may be converted by using the following factors:

Multiply inch-pound	By	To obtain metric units
	Length	
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
gallon (gal)	3.785	liter (L)
	3.785×10^{-3}	cubic meter (m^3)
	Flow	
cubic foot per second (ft^3/s)	28.32	liter per second (L/s)
	0.02832	cubic meter per second (m^3/s)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
	6.309×10^{-5}	cubic meter per second (m^3/s)
	Hydraulic Conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)

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

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ABSTRACT

A numerical model of ground-water flow was developed for a 10.3-square-mile area along the Susquehanna River in Columbia County, east-central Pennsylvania. Ground water in the model area primarily is in secondary openings in the carbonate- and clastic-rock aquifers and primary openings in the glacial-outwash aquifer that discontinuously overlies bedrock.

The ground-water flow model was calibrated under average steady-state conditions for 1981. The simulated 1981 water budget indicates an average inflow rate of 7.24 cubic feet per second. Of this, 93 percent is recharge from precipitation and 6.6 percent is boundary flow. Sixty-two percent of the outflow is leakage to streams, 21 percent to pumpage, and 17 percent to evapotranspiration. The model was calibrated under transient conditions for December 22, 1980 through April 21, 1982. Water-level fluctuations caused by natural stresses were more successfully simulated than those caused by pumping stresses.

Three 10-year, hypothetical stress periods were simulated with the calibrated, transient model. The general impact of three pumping schemes under hypothetical drought and drought recovery conditions were simulated.

INTRODUCTION

A project was done by the U.S. Geological Survey in cooperation with the Susquehanna River Basin Commission as part of their Special Ground-Water Study funded through the U.S. Water Resources Council. The objective of the Special Ground-Water Study is to determine the availability, distribution, and quality of the ground-water resources in the Susquehanna River basin. The objectives of the study addressed in this report are to:

1. Conceptualize and quantify ground-water flow in a complex aquifer system; and

2. Provide a means of evaluating the general impact of potential stresses on the aquifer system.

The area along the Susquehanna River between Berwick and Bloomsburg in Columbia County was selected for a modeling study because: (1) it is underlain by a carbonate-rock aquifer of regional importance as a present and potential source of water for municipal, commercial, and industrial use; (2) significant amounts of ground water are presently withdrawn in the area and additional ground-water development is predicted; and (3) data collected as part of an investigation of the ground-water resources of Columbia County and surrounding area by Williams and Eckhardt (1987) could be used in model development. This report discusses the development and use of a numerical model of ground-water flow in bedrock and glacial-outwash aquifers along the Susquehanna River in Columbia County, Pennsylvania.

Location and Physiographic Setting

The study area is located along the Susquehanna River between Berwick and Bloomsburg in Columbia County, east-central Pennsylvania (fig. 1). A 10.3 mi^2 (square mile) area was modeled. The area is within the Appalachian Mountain Section of the Valley and Ridge Physiographic Province. The Appalachian Mountain Section is characterized by mountainous terrain consisting of a series of long valleys and narrow ridges.

The model area is characterized by flat terraces and steep slopes. The terrace area averages about 3,000 ft (feet) wide and ranges in altitude from 450 to 500 ft above sea level. A sharp topographic break is present between the terrace and slope areas. The slope area generally is 1,000 to 2,000 ft wide and its altitudes are as high as 700 ft above sea level. The model area is drained by the Susquehanna River, Fishing Creek, and four smaller streams (fig. 1). Population centers include Bloomsburg, Espy, Almedia, and Lime Ridge.

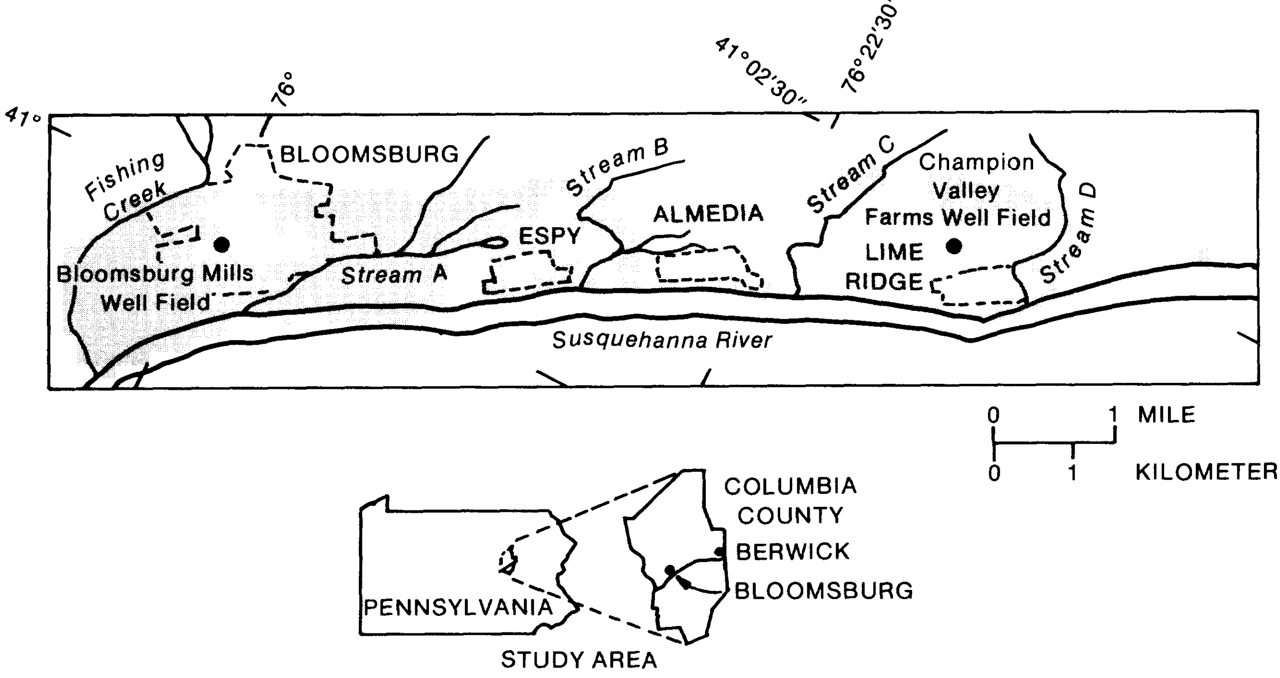


Figure 1.--Location of study and model areas, drainage, population centers, and major well fields.

Methods of Investigation

Hydrogeologic data were collected to define the characteristics of the aquifers and ground-water flow system. Locations of data-collection sites are presented on plate 1. Data collected from 83 wells and test holes included: (1) geologic and geophysical logs, (2) ground-water levels, (3) pumpage, (4) aquifer tests, and (5) field determinations of water quality. Data for wells and test holes are presented in table 1.

Data were collected from an additional 700 wells and test holes in the surrounding area (Williams and Eckhardt, 1987). As part of the U.S. Geological Survey ground-water level monitoring program, in cooperation with the Pennsylvania Geological Survey and the Susquehanna River Basin Commission, continuous water-level records have been obtained at well Co-45 since 1971 and water levels were measured weekly at well Co-1 from 1932 to 1978. Synoptic measurements of water levels were made in about 30 wells and test holes on December 22, 1980, April 29 and December 8, 1981, and April 22, 1982 (table 2). In addition to well Co-45, ground-water levels were continuously recorded at nine sites for varying lengths of time from August 1980 to June 1982. Low-flow discharge measurements were made at eight stream sites on August 2-3, 1982.

The hydrogeologic data were discretized and a numerical model of ground-water flow was constructed. The model was calibrated under average steady-state conditions for 1981 and transient conditions for December 22, 1980 to April 21, 1982. The transient calibration included matching water-level fluctuations caused by natural and pumping stresses. The calibrated transient model was used to simulate several hypothetical ground-water-development schemes under natural stress conditions.

Acknowledgments

The cooperation and assistance provided by landowners, well drillers, pump installers, industry representatives, and local and state officials are gratefully acknowledged. Special thanks goes to Gene Wieand of Wieand Brothers Drilling, James Swank of Swank and Son Pump Company, David Swank of Swank and Son Well Drilling, Robert Shober of Campbell Soup Company (Champion Valley Farms, Inc.), and William Gotshal of Bloomsburg Mills, Inc.

HYDROGEOLOGY

The study area is underlain by clastic and carbonate rocks of Silurian and Devonian age. Glacial-outwash sand and gravel of late Wisconsinan age overlie the bedrock along the Susquehanna River and Fishing Creek. The study area is on the southern limb of the Berwick anticlinorium, which trends N70E. The bedrock dips 30 to 40 degrees to the southeast. The underlying geologic units have been grouped into hydrogeologic units according to lithology (table 3). The areal distribution of the hydrogeologic units are presented in plate 2. Figure 2 presents a generalized section of the hydrogeologic units.

In the bedrock aquifers, water is stored in and flows through secondary openings such as fractures and bedding-plane separations. Weathering of carbonate material greatly enhances development of secondary permeability. In

Table 1.--Record of wells and test holes

Use: A, air conditioning; C, commerical; H, domestic and small commerical; I, irrigation; N, industrial; O, observation; P, public supply; R, recreation; S, stock; T, institution; U, test hole.

Topographic setting: H, upland; S, slope; T, terrace.

Aquifer: QGO, Glacial Outwash; DMH, Mahantango Formation; DMR, Marcellus Formation; DON, Onondaga Formation; DO, Old Port Formation; DSK, Keyser Formation; STO, Tonoloway Formation; SWC, Wills Creek Formation; SB, Bloomsburg Formation; SM, Mifflintown Formation; SRU, Upper Member of the Rose Hill Formation.

Lithology: DMSH, dolostone, limestone, and shale; LMSH, limestone and shale; LMSN, limestone and dolostone; SDGL, sand and gravel; SHLE, shale; SLSH, sandstone, limestone, and shale.

County number	Owner	Driller	Year completed	Use	Altitude of land surface (feet above sea level)	Topographic setting	Aquifer/ lithology
Co - 1	Howers			H	490	T	QGO/SDGL
45	U.S. Geological Survey		1970	O	700	H	SB/SHLE
48	Magee Carpet Co.			N	475	T	DOP/LMSH
51	Bloomsburg Mills, Inc	Kohl Brothers	1940	A	490	T	DOP/LMSH
52	Bloomsburg Mills, Inc	Kohl Brothers	1944	N	490	T	DOP/LMSH
53	Bloomsburg Mills, Inc	Kohl Brothers	1964	A	490	T	DOP/LMSH
63	Bloomsburg Packing Co.	Kohl Brothers	1946	N	480	T	DMR/SHLE
86	Scenic Knolls		1950	P	605	S	SWC/DMSH
87	Scenic Knolls		1964	P	610	S	SWC/DMSH
88	Scenic Knolls	R. R. Hornberger	1966	P	675	S	SB/SHLE
92	Johnson, J.	Stackhouse	1972	H	500	T	DMR/SHLE
98	Shrader, Don	Roy Zimmerman	1967	H	510	T	DMR/SHLE
102	Bloomsburg Carpet Ind.	R. R. Hornberger	1966	N	495	T	DSK/LMSN
103	St. Peters Chruch	R. R. Hornberger	1967	H	700	H	SB/SHLE
105	St. Peters Church	R. R. Hornberger	1966	H	500	T	DMR/SHLE
106	Dickson, D.	Stackhouse	1977	H	645	H	SWC/DMSH
108	Poloron Corp.	R. R. Hornberger	1970	N	510	T	DOP/LMSH
109	Schultz Plating	R. R. Hornberger	1973	N	705	S	SB/SHLE
110	Schultz Plating	R. R. Hornberger	1973	N	705	S	SB/SHLE
128	PA Power and Light	R. R. Hornberger	1972	C	720	S	SM/SLSH
132	Crawford, Joe	Stackhouse	1976	H	485	T	DSK/LMSN
133	Koons, Venice	Champion	1970	H	485	T	DMR/SHLE
134	Isola	Champion	1970	H	480	T	DMR/SHLE
135	Meckley, Donald		1967	H	750	S	SRU/SLSH
136	Amoco	Champion	1974	H	495	T	STO/LMSN
137	Hill, Mary	Stackhouse	1978	H	840	S	SRU/SLSH
141	U.S. Geological Survey		1979	U	500	T	QGO/SDGL
142	U.S. Geological Survey		1979	U	505	T	QGO/SDGL
146	U.S. Geological Survey		1979	U	505	T	QGO/SDGL
147	U.S. Geological Survey		1979	U	500	T	QGO/SDGL
149	U.S. Geological Survey		1979	U	510	T	QGO/SDGL
152	U.S. Geological Survey		1979	U	470	T	QGO/SAND
157	Magee, James	R. R. Hornberger	1968	H	785	S	SM/SLSH
158	Neyhard, Robert	R. R. Hornberger	1966	H	760	H	SB/SHLE
161	Bloomsburg Carpet Ind.	R. R. Hornberger	1977	N	495	T	STO/LMSN
162	Col-Mont Vo-Tech	R. R. Hornberger	1967	P	565	S	SWC/DMSH
163	Holdren, Robert	R. R. Hornberger	1966	H	500	T	STD/LMSN
164	Wolf, John		1980	H	520	S	SWC/DMSH

Table 1.--Record of wells and test holes--(continued)

Depth to water level; Depth--F, flows but head is not known; minus sign indicates above land surface.
Date--month/last two digits of year.

Reported Yield; gal/min, gallons per minute.

Specific capacity: (gal/min)/ft, gallons per minute per feet of drawdown.

Pumping rate: gal/min, gallons per minute.

Hardness: mg/L, milligrams per liter.

Specific conductance: DEG C, degrees Celsius.

Well depth (feet)	Casing Depth (feet)	Casing Diameter (inches)	Depth(s) to water bearing zones(s) (feet)			Depth to water level (feet)	Date measured (mo/yr)	Reported yield (gal/min)	Specific capacity [(gal/min)/ft]	Hardness (mg/L)	Specific conductance (µS/cm at 25°C)	pH	County well number
18						11	06/80						1
282	32	6	115	163		83	04/81	1	0.03/1		480		45
202	48	8				42	01/30		5.6 /185				48
498	94	8				25	03/58		3.2 /542				51
550	115	10				35	11/64		17 /650	462	980	7.7	52
420	77	12				30	11/64		3.8 /620				53
525	42	8				7	05/46	225					63
190								8					86
402								5					87
415	42	7	70	103	157	80	06/66	8					88
70	22	6	42	66				30		290	470	7.3	92
120	71	6						12		68	183	6.6	98
95	36	6	60	90			08/66	40		222	516	7.0	102
175	30	6				114	11/80	3		102	200	6.8	103
67	50	6	62			20	12/66	30					105
173	21	6	110	165		101	11/80			120	140	8.1	106
300	184	8	130	330	280	34	07/70	350	1.1 /200	254	600	7.9	108
390	27	6				40	06/73	9					109
495	21	6				40	06/73	5					110
200	41	6				6	10/72		.37/12	171	260	7.8	128
47	27	6	47			3	04/76	10		308	380	7.7	132
53	35	6	45			15	03/70	30					133
75	30	6	40	60		20	03/70	12					134
280	51	6	210	270		60	01/67	8		111	220	7.1	135
75	40	6	52			5	06/80	15					136
173	61	6	165			33	06/80	8	.1 /3	68	85	6.9	137
37	32	2				20	06/80						141
47	42	2				23	06/80						142
													146
													147
													149
													152
73	41	6	67			43	06/80	20		51	135	5.9	157
215	18	6	90	148	174	50	06/66	15		103	210	7.7	158
40	17	6				10	12/80						161
155	74	6	89	130	148	53	06/80		1.8 /40				162
95	22	6	45	71	90	1	08/66	12		325	360	7.3	163
25						13	06/80			307	455	7.1	164

County number	Owner	Driller	Year completed	Use	Altitude of land surface (feet above sea level)	Topographic setting	Aquifer/ lithology
Co-165	Yorty, Cindy	Alvin Swank & Son	1980	H	485	T	DMH/SHLE
166	Wagner, Claire	Stackhouse	1972	H	490	T	DMH/SHLE
167	Young, Gerald	R. R. Hornberger	1966	J	495	T	DMH/SHLE
168	Hause, Walter	Alvin Swank & Son	1979	H	520	T	DSK/LMSN
169	Horeck, John	Champion	1973	H	520	T	STO/LMSN
182	Belles, David	Virgil Huck	1975	H	525	T	DON/LMSH
183	Huber, Richard	Champion	1976	H	515	T	DOP/LMSH
186	Streater, J.	R. R. Hornberger	1968	T	470	T	DMH/SHLE
187	Wintersteen, L.			H	490	T	DMR/SHLE
189	Kawneer, Inc.	R. R. Hornberger	1966	N	470	T	DMR/SHLE
190	Kawneer, Inc.	R. R. Hornberger	1966	N	475	T	DMR/SHLE
196	Champion Valley Fms.		1963	N	500	T	DOP/LMSH
197	Champion Valley Fms.		1963	N	500	T	DOP/LMSH
198	Champion Valley Fms.		1964	N	500	T	DOP/LMSH
199	Champion Valley Fms.	R. R. Hornberger	1968	N	500	T	DOP/LMSH
202	Sweeny, Scott				505	T	QGO/SDGL
203	Sweeny, Scott			H	505	T	DMR/SHLE
204	Columbia Co. Dev. Auth.	R. E. Kresge	1970	N	510	T	SWC/DMSH
205	Columbia Co. Dev. Auth.	R. E. Kresge	1970	N	510	T	SWC/DMSH
212	Rupert, Helen			H	500	T	DMR/SHLE
213	Rupert, Helen			H	500	T	QGO/SDGL
301	U.S. Radium Corp.	Wieand Brothers	1979	U	490	T	QGO/SDGL
302	U.S. Radium Corp.	Wieand Brothers	1980	U	490	T	SGO/SDGL
303	U.S. Radium Corp.	Wieand Brothers	1979	U	490	T	QGO/SDGL
305	U.S. Geological Survey	Alvin Swank & Son	1980	O	515	T	QGO/SDGL
306	U.S. Geological Survey	Alvin Swank & Son	1980	O	490	T	DMR/SHLE
309	Lupini, H.			H	510	T	DMR/SHLE
310	Coombs, William	Alvin Swank & Son	1980	C	490	T	DSK/LMSN
311	Hudelson, Foster			H	510	T	QGO/SDGL
331	Fritz, Jeff	Stackhouse	1980	H	500	T	SM/SLSH
332	Bell Telephone Co.	Wieand Brothers	1974	C	500	T	SM/SLSH
344	Columbia Co. Dev. Auth.		1977	U	515	T	QGO/SDGL
345	Columbia Co. Dev. Auth.		1977	U	510	T	QGO/SDGL
346	Columbia Co. Dev. Auth.		1977	U	520	T	QGO/SDGL
347	Columbia Co. Dev. Auth.		1977		520	T	QGO/SDGL
348	Krum, Robert	Stackhouse	1980	H	510	T	DOP/LMSH
349	Krum, Robert			H	510	T	DOP/LMSH
355	Vance, James			H	700	H	SM/SLSH
373	Arco	Wieand Brothers	1980	O	480	T	SWC/DMSH
441	Swisher, Gary			H	500	T	DON/LMSH
446	Baker Trailer Park			P	520	S	STO/LMSH
448	Champion Valley Fms.	Wieand Brothers	1981	N	500	T	DOP/LMSH
452	Bloomsburg Water Co.	R. R. Hornberger	1967	P	500	T	DMR/SHLE
505	Champion Valley Fms.	Wieand Brothers	1981	N	500	T	STU/LMSN
573	Arco	Wieand Brothers	1980	O	480	T	SWC/DMSH

Table 1.--Record of wells and test holes--(continued)

Well depth (feet)	Casing Depth (feet)	Casing Diameter (inches)	Depth(s) to water bearing zones(s) (feet)	Depth to water level (feet)	Date measured (mo/yr)	Reported yield (gal/min)	Specific capacity [(gal/min)/ft]	Hardness (mg/L)	Specific conductance (µS/cm at 25°C)	pH	County well number
175				13	06/80	2		256	380	7.7	165
60	47	6	59			14		290	420	7.5	166
63	31	6	50	25	06/80	3		188	260	7.2	167
80	72	6		49	06/80	35		222	360	7.3	168
100	40	6	85			7					169
80	50	6	50 80			10		188	280	7.7	182
225	40	6	175	30	06/80	5		137	205	6.8	183
500	32	8	35 70 165	12	07/68		1.4 /250				186
			360 470								
86				16	07/80		18 /32	188	490		187
355	26	6	41 59 94	30	12/80		.06/20				189
415	24	10	40 68 118	6	07/80	100	0.41/20	120	2500		190
			156 308								
268	45	7		47	07/80						196
550	41	12				250					197
600	40	8				440					198
500	43	10	50 170 180	25	06/68		1.3 /218				199
34				33	07/80						202
110				38	12/80			170	400		203
273				31	10/80		7.4 /141	180	430		204
248	62	6	130 160 190	32	08/80		1.1 /136				205
120	55	6	78 93 117	32	12/80		.5 /16	137	340		212
33				20	09/80						213
35	35	6									301
35	30	6									302
37	37	6									303
68	68	6		32	12/80	50	10 /38	54	118	5.6	305
	42	6	60 74	21	06/81	25	1.7 /17	188	395		306
69	40	6		38	09/80		.44/4	86	220		309
360	40	8		21	12/80	250	3.9 /75	19	730	6.4	310
35				32	10/80						311
125	30	6		12	12/80	5		103	300		331
350	43	6	84 262	F	12/80		.13/30	68	280		332
25	23	6									344
25	20	6									345
25	24	6									346
25											347
120	85	6				6					348
54	39	6									349
				57	12/80			51	169		355
				15	04/81						373
				19	04/82			291	580		441
				14	04/82						446
155	32	8	141 152	26	08/81	150	2.2 /200	395	750		448
500	54	7	110 171	28	05/81	60	0.84/34	137	335		452
			330 361 410								
570	40	8	112 275	15	09/81	360	2.0 /280	154	300		505
			460 510								
30	30	10		14	04/82						573

Table 2.--Synoptic measurements of ground-water levels in selected wells and test holes

County number	Altitude of land surface (feet above sea level)	Depth to water level (feet below land surface)			
		December 22, 1980	April 29, 1981	December 8 1981	April 22, 1982
Co- 1	490	12.24	10.62	11.25	10.35
45	700	85.81	83.57	84.54	81.67
103	700	115.12	117.53	-	-
106	645	101.66	95.51	99.61	87.52
136	495	7.69	5.56	6.90	4.69
141	500	23.27	21.96	23.00	20.47
142	505	26.88	25.47	26.53	23.78
154	470	8.75	6.90	8.05	5.34
161	495	10.14	8.34	9.46	-
162	575	71.03	68.23	70.75	-
164	520	17.63	14.63	16.89	12.84
165	485	15.92	13.67	15.19	12.48
167	495	30.50	24.81	27.03	22.50
168	520	50.34	47.85	48.58	45.08
183	510	42.34	31.05	-	26.48
187	490	18.00	16.26	17.83	13.17
190	475	8.37	6.00	7.80	4.35
202	505	Dry	33.72	Dry	30.75
203	505	38.18	34.58	36.95	31.23
205	510	34.85	31.50	34.05	28.75
212	500	34.40	30.80	33.38	29.73
213	500	34.40	30.80	33.38	29.73
305	515	32.96	31.98	32.80	30.88
306	495	23.78	21.48	23.16	19.98
310	490	21.87	19.26	20.94	17.36
330	495	11.13	9.56	10.90	8.92
331	500	12.93	10.99	11.84	10.35
355	700	57.27	49.30	56.40	44.25
373	480	17.35	15.18	-	-
413	500	-	-	11.061	6.42
441	500	-	-	30.97	19.25
448	500	-	-	43.40	17.78
452	500	-	28.03	29.65	26.75
573	480	-	-	16.51	13.80

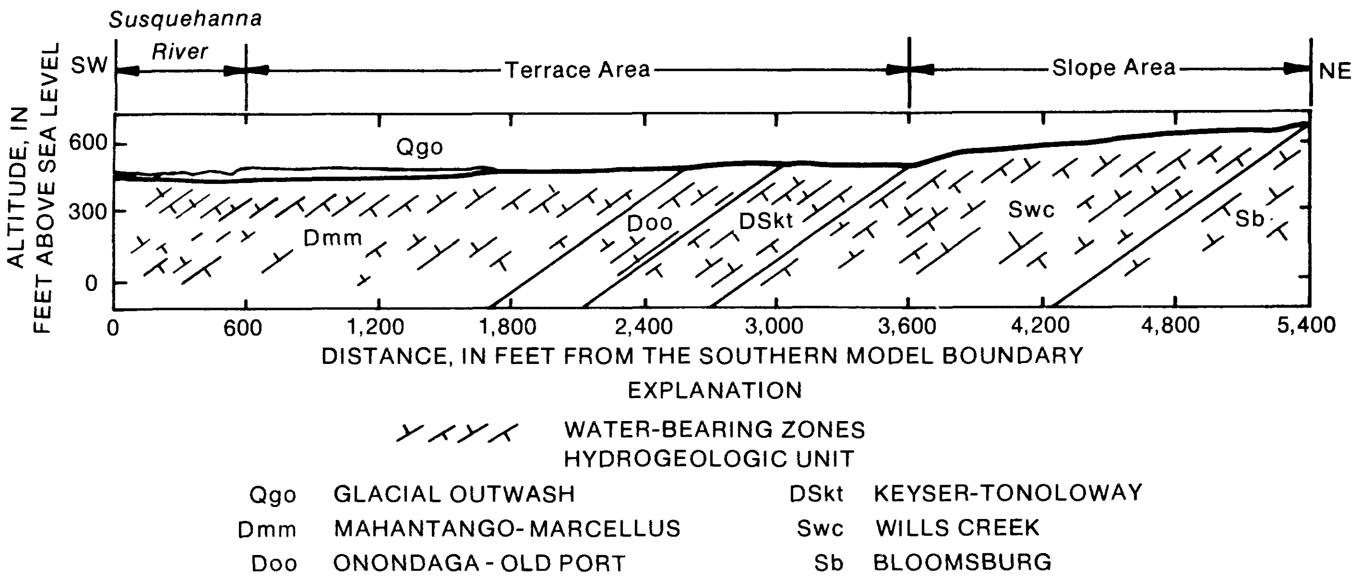


Figure 2.-- Generalized section showing the distribution of hydrogeologic units.

general, wells completed in aquifers containing carbonate rocks have specific capacities (well yields in gallons per minute per foot of drawdown) that are an order of magnitude greater than those wells completed in clastic-rock aquifers (Williams and Eckhardt, 1987).

The number and size of secondary openings in the bedrock aquifers decreases with depth because of the increase in overburden pressure and decrease in weathering activity. Temperature gradients measured in deep wells generally approach the geothermal gradient at depths greater than 300 ft, indicating that most ground-water flow occurs within 300 ft of land surface (Williams and others, 1984). Drillers' data suggest that the vertical spacing between water-bearing zones increases more rapidly with depth in clastic rocks than in aquifers containing carbonate rock. The average vertical spacing between water-bearing zones within 300 ft of land surface is about 65 ft for clastic-rock aquifers and 55 ft for interbedded clastic and carbonate rock and carbonate-rock aquifers. The average vertical spacing between zones from 300 to 600 ft below land surface is 185 ft for clastic-rock aquifers and 120 ft for aquifers with carbonate beds.

In addition to lithology and depth, topography has a major influence on the distribution of permeability in the bedrock aquifers. Median specific capacities grouped according to lithology for wells drilled in valleys are approximately 2 to 10 times greater than those for slope wells and 3 to 20 times greater than for upland wells (Williams and Eckhardt, 1987).

The bedrock aquifers have greater permeability along the direction of bedding strike than the across strike direction. Data from continuous water-level recorders indicate that cones of depression from major well fields migrate along the general bedding strike for distances of 1,000 to 3,000 ft, whereas high-yielding wells located across bedding strike as close as 500 ft apart may not show significant interference. A multiple-well aquifer test

Table 3.-- Description of hydrogeologic units

System and series		Geologic unit	Thickness (feet)	Lithologic description 1/	Hydrogeologic unit
		Glacial outwash and alluvium	0 - 70	Sand and gravel, containing some clay, silt, cobblers, and boulders.	Outwash
DEVONIAN	Upper	Trimmers Rock Formation	2,500	Predominantly interbedded gray to dark gray siltstone and shale; with considerable sandstone in the upper part and shale in the lower part.	Trimmers Rock
DEVONIAN	Upper	Harrell Formation	100	Dark-gray shale; interbedded with siltstone in the upper part.	Mahantango-Marcellus
DEVONIAN	Middle	Mahantango Formation Tully Member	50 - 60	Interbedded, argillaceous limestone and calcareous shale, dark gray, fossiliferous.	Mahantango-Marcellus
DEVONIAN	Middle	Lower Member	1,100 - 1,200	Greenish to dark-gray shale, locally calcareous; some calcareous and fossiliferous siltstone beds in the upper part.	Mahantango-Marcellus
DEVONIAN	Middle	Marcellus Formation	300	Dark-gray fissile shale, pyritic and carbonaceous.	Mahantango-Marcellus
DEVONIAN	Middle	Onondaga Formation	50 - 175	Interbedded gray agrillaceous limestone and calcareous shale in upper part gray to dark-gray noncalcareous to very calcareous shale in lower part.	Onondaga-Old Port
DEVONIAN	Lower	Old Port Formation	150	Variable lithologic sequence, consisting of dark-gray, slightly calcareous chert, locally sandy and fossiliferous in the upper part; dark-gray, calcareous shale in the middle part; dark-gray, fine- to coarse grained cherty, fossiliferous limestone at the bottom.	Onondaga-Old Port
DEVONIAN / SILURIAN	Lower / Upper	Keyser Formation	125	Gray to bluish-gray limestone, fine-to coarse-grained, thin-to thick-bedded; laminated, argillaceous and dolomitic in the upper part; coarse grained and highly fossiliferous in the middle part; nodular, argillaceous and fossiliferous in the lower part; calcareous shale interbeds crease in frequency in the upper part.	Keyser-Tonoloway
SILURIAN	Upper	Tonoloway Formation	200	Laminated, gray to dark-gray, fine-grained limestone; considerable dolomitic limestone and dolostone in the lower part; calcareous shale interbeds increase in frequency and thickness toward base.	Keyser-Tonoloway
SILURIAN	Upper	Wills Creek Formation	600 - 700	Interbedded calcareous shale, argillaceous dolostone and limestone, and calcareous siltstone; gray, yelllowish-gray and greenish-gray in the upper part; varigated greenish-gray, yellowish-gray, grayish red purple in the lower part.	Wills Creek
SILURIAN	Upper	Bloomsburg Formation	500	Grayish-red shale with interbeds of grayish-red siltstone, calcareous in part a 30 foot thick interval of grayish-red sandstone in the upper part.	Bloomsburg
SILURIAN	Middle	Mifflintown Formation	240	Dark-gray limestone and calcareous shale in the upper part; dark-gray calcareous shale with interbeds of coarse-grained limestone in the middle part; light gray quartzitic sandstone and siltstone with interbeds of greenish-gray shale in the lower part.	Mifflintown-Rose Hill
SILURIAN	Middle	Rose Hill Formation Upper member	120	Interbedded shale, limestone, and sandstone; mostly gray to greenish-gray.	Mifflintown-Rose Hill
SILURIAN	Middle	Middle member	60	Reddish-purple hematitic sandstone, with interbeds greenish-gray to reddish purple shale in the upper part.	Mifflintown-Rose Hill
SILURIAN	Middle	Lower member	720	Greenish-gray shale; with interbeds of gray, calcareous and reddish-brown, hematitic sandstone.	Mifflintown-Rose Hill

1/ Adapted from Inners (1981).

involving five closely-spaced, shallow wells completed in interbedded carbonate rock and shale showed that about six times more drawdown occurred along bedding strike than across strike.

In the outwash sand and gravel, water is stored in and flows through primary openings between individual grains. Permeability largely depends on grain size and sorting. Thicknesses of sand and gravel penetrated in wells and test holes, as shown on plate 2, are highly variable. The outwash is only locally saturated and the saturated thickness is generally less than 20 ft.

The degree of hydraulic connection between individual water-bearing zones in the bedrock and between the bedrock and sand and gravel aquifers is variable. There are no well-defined confining beds, although zones of unfractured bedrock serve as effective confining units complicating the flow system. In general, the aquifers combine to act as a single, complex, water-table system. The water table is a subdued expression of topography and ground-water divides generally follow the topographic divides.

Recharge to the aquifer is primarily from the infiltration of precipitation. Ground-water flows in the direction of decreasing water-table altitude or, in general, from areas of higher to lower altitude. The Susquehanna River and its tributaries serve as discharge areas for the ground-water flow system. Recharge occurs in all areas upgradient from these discharge areas. In areas where the water table is near to the land surface, some water may be lost as ground-water evapotranspiration.

Base-flow separation of surface-water hydrographs commonly is used to estimate ground-water recharge for drainage basins. Base flow is assumed to equal recharge. Estimates of recharge from such methods do not consider consumptive use of ground water or ground-water evapotranspiration. Estimates of ground-water discharge could not be made directly for the study area. However, estimated ground-water discharge is available for basins in the Appalachian Mountain section of south-central Pennsylvania that contain the Silurian and Devonian stratigraphic sequence. Taylor and others (1982) and Johnston (1970) estimate an average ground-water discharge of 10 in./yr (inches/year) and 8.2 in./yr for the Juniata River and Bixler Run basins, respectively.

Ground-water flow primarily occurs in localized, shallow systems. However, some water does follow deeper flow paths to discharge areas. In the slope area, the deeper water-bearing zones generally have the deepest water levels. Conversely, in the terrace area, deeper water-bearing zones have the highest water levels. This indicates the potential for downward and upward flow between deep and shallow zones in the slope and terrace areas, respectively.

The water table, as indicated by water levels measured in wells, fluctuates in response to changes in recharge to and discharge from the aquifers. Water-level fluctuations primarily are caused by seasonal variations in recharge. The highest and lowest mean monthly water levels in wells Co-1 and Co-45 occur, respectively in early spring (March-April) and early fall (September-October) (fig. 3). Water levels in some wells also show the effects of changes in discharge from the aquifers due to pumpage. Figure 4 presents a comparison of water levels observed in wells Co-190, Co-305 and Co-310. The water level in well Co-310 is affected by the pumpage of the Bloomsburg Mills well field.

The altitude of the water table was contoured on 10-ft and 20-ft intervals in the terrace and slope areas, respectively (plate 3). Averages of the 1981 synoptic water-level measurements (April 29 and December 8), single measurements, and reported levels, as well as selected stream-stage altitudes, were used for control. Most wells used for control are between 100 to 300 ft deep.

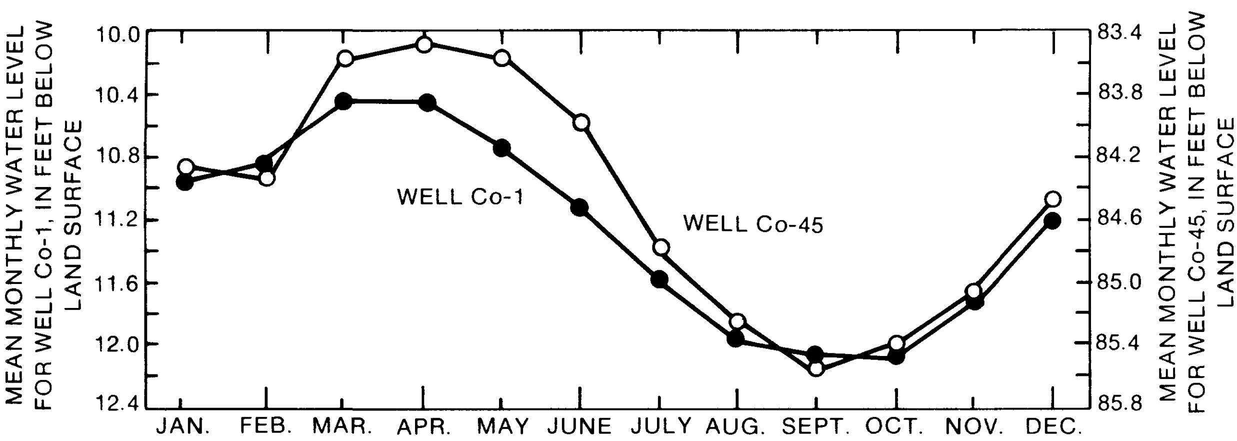


Figure 3.--Mean monthly water levels for wells Co-1 (1932-1978) and Co-45 (1971-1981).

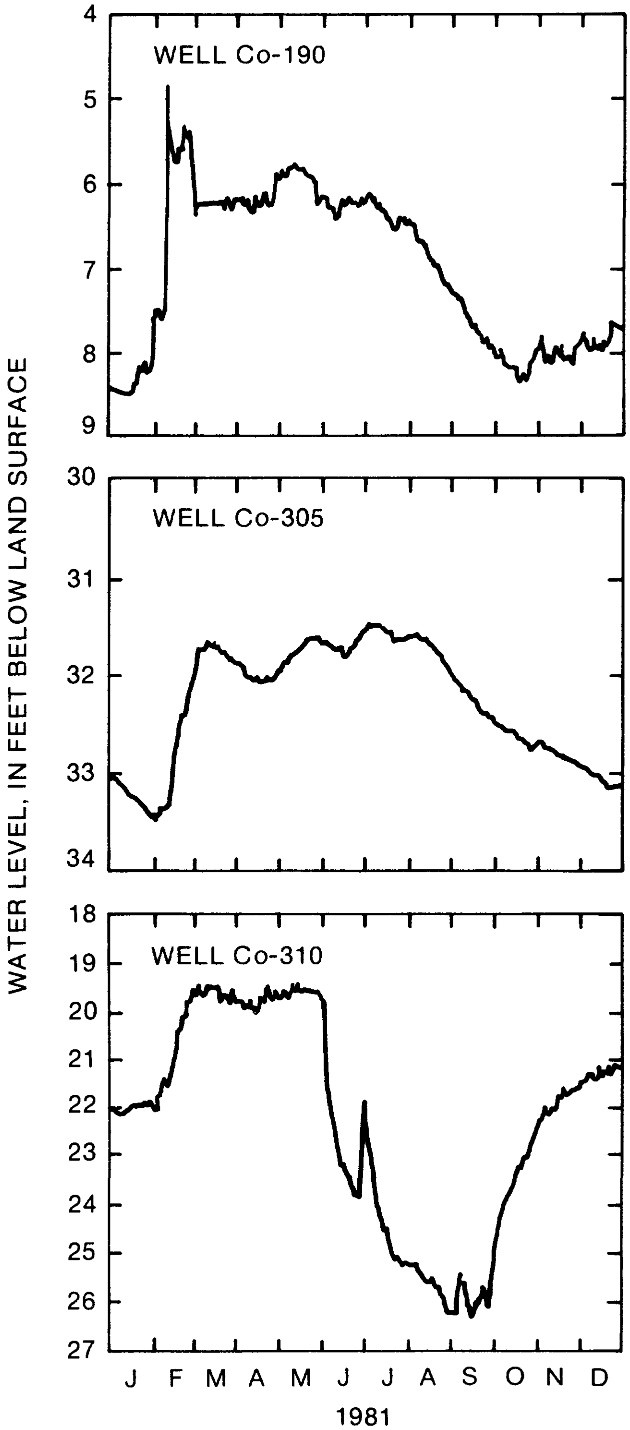


Figure 4.-- Water-level fluctuations in wells Co-190, Co-305, and Co-310 during 1981. The water level in well Co-310 is affected by pumpage from the Bloomsburg Mills well field.

MODEL DEVELOPMENT

Simplifying Assumptions

A number of simplifying assumptions were made concerning the distribution of hydraulic conductivity and storage in the aquifers during the development of the numerical model.

These include the following:

1) The relationship between the scale of modeling and the distribution of hydraulic conductivity and storage in the aquifers in such that a porous media model (Trescott, 1975) could be applied.

2) The depth of significant hydraulic conductivity and storage in the aquifers is 600 ft below land surface. The upper 300 ft of aquifer has higher hydraulic conductivity and storage values than the lower 300 ft; the ratio of upper to lower layer values decreases with increasing content of carbonate rock.

3) The ratio of hydraulic conductivity along and perpendicular to bedding strike is greater for aquifers containing carbonate rock than for shale aquifers.

4) The increase in transmissivity and specific yield due to the presence of saturated sand and gravel is proportional to its thickness and can be determined from a weighted average of sand and gravel and bedrock values.

Figure 5 presents a generalized section of the simplified distribution of the hydrogeologic units. The aquifers have been divided into two layers, each of which is 300 ft thick. The base of the aquifers is 600 ft below land surface. Bedrock within the same hydrogeologic unit and layer is assumed to have uniform hydraulic conductivity and storage properties. The Wills Creek unit is subdivided into two units to approximate the gradational increase in shale content toward its base. The glacial-outwash unit is lumped with the underlying bedrock unit.

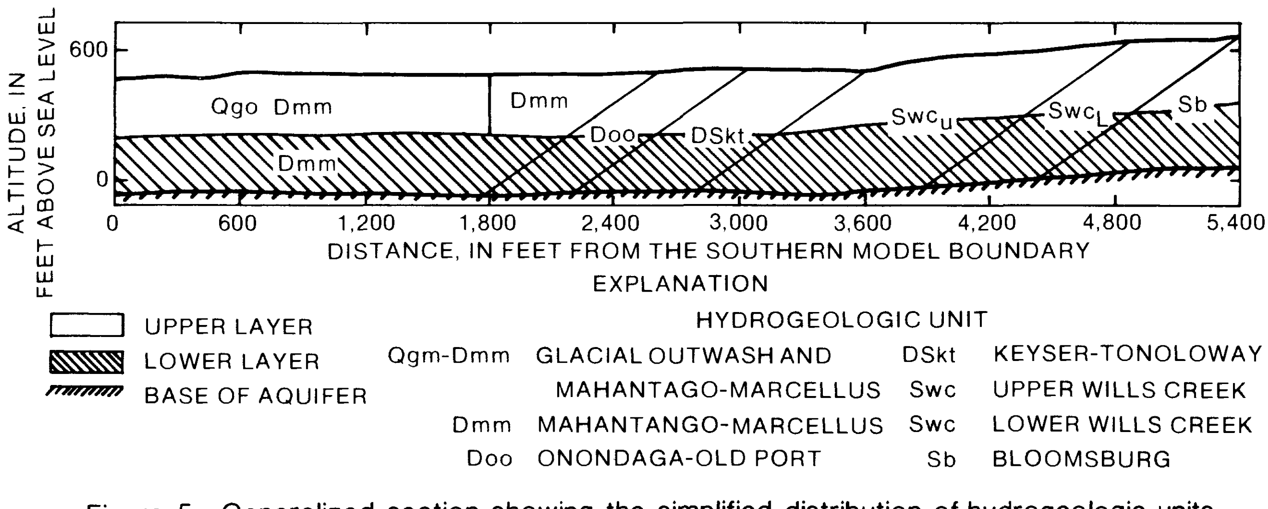


Figure 5.--Generalized section showing the simplified distribution of hydrogeologic units.

Model Selection, Design, and Construction

The three-dimensional, finite-difference model used in this study is a modified version of the model presented in Trescott (1975), Trescott and Larson (1976), and Gerhart and Lazorchick (1988). Major modifications to the original Trescott model are listed below. Additional discussion of the model modifications are included under the appropriate hydrologic variables.

1. Incorporation and extension of the input changes of Gerhart and Lazorchick (1988), which permit the entry of certain aquifer characteristics by hydrogeologic unit and their subsequent modification by topographic setting.

2. Addition of head-dependent flow conditions, including evapotranspiration, stream leakage and boundary flow. Program modifications were taken in part from Gerhart and Lazorchick (1988).

3. Inclusion of a method to simulate a thin, discontinuous layer of saturated sand and gravel. This method calculates block transmissivity and specific yield from a weighted average of the sand and gravel and bedrock aquifers values.

4. Addition of a modification that allows for recharge and evapotranspiration rates to be entered for each transient-simulation period.

5. Addition of the modification of Gerhart and Lazorchick (1988) to print out stream and boundary flow for each block.

The finite-difference grid used in the model is presented in plate 4. The grid consists of a lower and upper layer with 798 active blocks in each. The grid blocks are 600 ft on a side and 300 ft thick. Two layers were used in order to approximate vertical changes in aquifer characteristics. The grid was aligned along bedding strike so anisotropic permeability conditions could be incorporated into the model.

The northern boundary of the model area is the outcrop of the Bloomsburg shale. The boundary was simulated as a head-dependent flow boundary, which allows for steady-state flow across the model border depending on the head in the boundary blocks. No-flow, boundary conditions were used to simulate the western, eastern, and southern borders of the model area. The western and southern model borders generally correspond to the western bank of Fishing Creek and the middle of the Susquehanna River, respectively. The eastern boundary was modeled at a topographic divide in the Bloomsburg shale. The Susquehanna River and Fishing Creek were simulated as head-dependent, stream-leakage blocks. The smaller tributaries to the Susquehanna River also were simulated as head-dependent, stream-leakage blocks.

Steady-State Calibration

The model was calibrated under average, steady-state conditions for 1981. Annual-average conditions were used because at no one time can the ground-water system be considered to be at steady state. However, during a period in which the change in ground-water storage is minimal, average, steady-state conditions may be assumed. As shown by hydrographs of wells Co-190, Co-305 and Co-310 (fig. 3), the average change in ground-water storage was minimal between the beginning and end of 1981.

The calibration procedure consisted of adjusting the various hydrologic variables within certain ranges estimated from field data to produce a consistent and reasonable representation of the ground-water flow system. In general, hydraulic-conductivity values and stream-leakage coefficients were adjusted to produce the best match with observed water levels and measured low flows. The water-table altitude map (plate 3) was assumed to represent average, water-level conditions for 1981. Low flow measured on August 2-3, 1982, was assumed to generally represent steady-state, base-flow conditions in the small streams, although transient ground-water withdrawal during 1981 and 1982 do not permit a direct comparison. In addition, average base flow for 1981 is probably slightly lower because water levels measured in wells Co-45, Co-305, and Co-452 at the time of the low-flow measurements averaged about 1 ft higher than their average levels for 1981.

Hydrologic Variables

Hydrologic variables entered as data input to the model are given in table 4. Variables relating to aquifer geometry, hydraulic conductivity, stream leakage, recharge, evapotranspiration, pumpage and boundary flow are discussed in the following sections. Aquifer-storage coefficients are not used in modeling steady-state conditions and are discussed under transient calibration.

Table 4.--Hydrologic variables used in the ground-water flow model

(I=row, J=column, K=layer, N=bedrock hydrogeologic unit, M=topographic setting)

STRT(I,J)	Altitude of aquifer head in the upper and lower layers at start of simulation
IZN(I,J,K)	Bedrock hydrogeologic unit
TOPO(I,J)	Topographic setting
BOTTOM(I,J)	Altitude of the bottom of the upper layer
BOWC(I,J)	Altitude of saturated sand and gravel-bedrock contact
POWC(I,J)	Hydraulic conductivity of saturated sand and gravel
SOWC(I,J)	Specific yield of saturated sand and gravel
KXL(N)	Hydraulic conductivity in the direction perpendicular to bedding strike in the lower layer
KYL(N)	Hydraulic conductivity in the direction parallel to bedding strike in lower layer
KZL(N)	Hydraulic conductivity in the vertical direction in the lower layer

Table 4.--Hydrologic variables used in the ground-water flow model--Continued

(I=row, J=column, K=layer, N=bedrock hydrogeologic unit, M=topographic setting)

BZL	Thickness of the lower layer
SZL(N)	Storage coefficient in the lower layer
KXU(N)	Hydraulic conductivity of the bedrock in the direction perpendicular to bedding strike in the upper layer
KYU(N)	Hydraulic conductivity of the bedrock in the direction parallel to bedding strike in the upper layer
KZU(N)	Hydraulic conductivity of the bedrock in the vertical direction in the upper layer
BZU(N)	Initial saturated thickness of the bedrock in the upper layer
SZU(N)	Specific yield of the bedrock in the upper layer
PMULT(N,M)	Multiplier for topographic setting
RCG(I,J)	Stream-leakage coefficient for gaining stream conditions
RCL(I,J)	Stream-leakage coefficient for losing stream conditions
RHSS(I,J,K)	Altitude of constant stream stage
HB(I,J,K)	Altitude of stream-infiltration cutoff
QRE(I,J)	Recharge rate
ECSS(I,J)	Evapotranspiration coefficient
EHB(I,J)	Altitude at and above which evapotranspiration is maximum
EHSS(I,J)	Altitude at which evapotranspiration ceases
WELL(I,J,K)	Rate of pumpage from wells
BHSS(I,J,K)	Altitude of aquifer head at the aquifer boundary
BCSS(I,J,K)	Coefficient of flow for head-dependent flow boundary

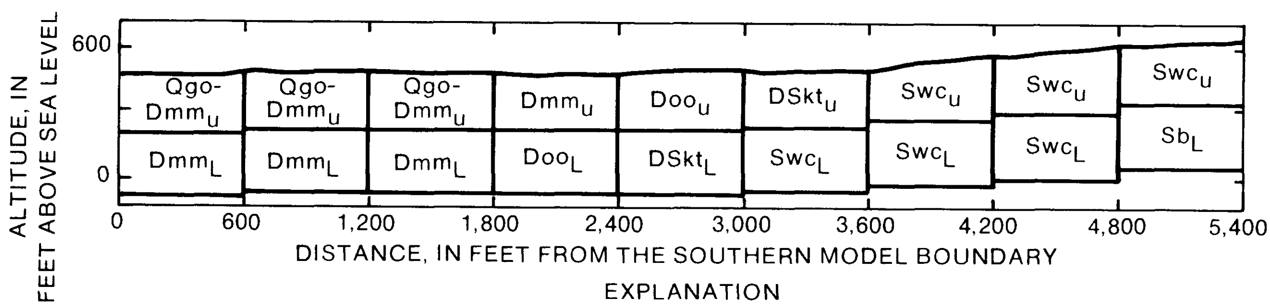
Aquifer geometry

Hydrologic variables that define the distribution and thickness of the aquifers include the following: altitude of land surface (GRND), altitude of aquifer head (STRT), bedrock hydrogeologic unit (IZN), topographic setting (TOPO), altitude of the bottom of the upper layer (BOTTOM), thickness of the lower layer (BZL), initial saturated thickness of the upper layer (BZU), and altitude of the contact between bedrock and saturated sand and gravel (BOWC).

The matrix GRND is the average of land-surface altitudes taken from U.S. Geological Survey 7-1/2 minute quadrangle maps at four equally spaced points within each grid block. Grid-block, land-surface altitudes range from 455 to 685 ft above sea level. The GRND matrix was not input directly into the model but was used to compute the BOTTOM and BOWC matrices.

The STRT matrix was discretized from the water-table altitude map (plate 3). The STRT matrix represents average, water-level conditions 1981 in the upper layer; however, for convenience, the matrix is used as the altitude of aquifer head for both the upper and lower layer at the start of simulation. Simulated aquifer heads for the upper layer were compared to the STRT matrix during calibration.

Each grid block was assigned to a bedrock hydrogeologic unit defined in the IZN matrix. The discretized bedrock unit maps of the upper and lower layers are presented in plate 5. Figure 6 presents a generalized section showing the discretized distribution of hydrogeologic units. The bedrock unit assigned to a grid block in the lower layer was not necessarily the same unit as in the upper layer because of dipping beds. Bedrock units in the upper layer were offset in the lower layer according to the structural dip.



EXPLANATION

MODEL GRID BLOCK

HYDROGEOLOGIC UNIT

Qgo-Dgm GLACIAL OUTWASH AND MAHANTAGO-MARCELLUS

Dmm MAHANTANGO-MARCELLUS

Doo ONONDAGA-OLD PORT

DSkt KEYSER-TONOLOWAY

Swc UPPER WILLS CREEK

Swc LOWER WILLS CREEK

Sb BLOOMSBURG

Subscripts U and L indicate upper and lower model layer, respectively.

Figure 6.--Generalized section showing the discretized distribution of hydrogeologic units.

Each grid block was assigned to a topographic setting (plate 5). The settings, which were stored in the TOPO matrix, included terrace, lower slope or draw, upper slope, and upland.

The altitude of the bottom of the upper layer (BOTTOM) was defined to be 300 ft below GRND. Initial saturated thickness for each bedrock hydrogeologic unit was determined by the following equation:

$$BZU(N) = 300\ FT - X(N) \qquad (1)$$

where X(N) = median of [GRND(I,J) - STRT(I,J)] for bedrock hydrogeologic unit N;
GRND(I,J) = altitude of land surface at block I,J;
STRT(I,J) = water-table altitude at block I,J; and
BZU(N) = initial saturated thickness of bedrock hydrogeologic unit N.

Initial saturated thicknesses ranged from 230 to 285 ft.

Depth-to-bedrock data (plate 3) were used to define the altitude of the saturated sand and gravel-bedrock contact (BOWC matrix) for each grid block that contains the glacial-outwash hydrogeologic unit. The discretized distribution of the outwash aquifer is shown in plate 5.

The thickness of the lower layer, BZL, was defined as 300 ft. It is assumed that significant permeability generally does not exist below a depth of 600 ft (the bottom of the lower layer).

Hydraulic conductivity

Hydraulic-conductivity variables include the following: outwash hydraulic conductivity (POWC); bedrock hydraulic conductivity perpendicular to bedding strike, parallel to bedding strike and the vertical direction in the lower layer (KXL, KYL, and KZL), and in the upper layer (KXU, KYU, and KZU); and topographic-setting multipliers (PMULT). The hydraulic conductivity was adjusted during steady-state calibration on a hydrogeologic unit and topographic basis.

Hydraulic conductivity was estimated for each hydrogeologic unit from the specific-capacity data presented in Williams and Eckhardt (1987). The specific-capacity data, which included 80 aquifer tests from one to 72 hours in duration, were adjusted to a common 24-hour pumping period using the following equation:

$$ASC = SC \bullet [0.594+0.294 \bullet \mathrm{Log}_{10}(PP)] \qquad (2)$$

where ASC = specific capacity adjusted to 24-hour pumping period, in (gal/min)/ft;
SC = specific capacity, in (gal/min)/ft; and
PP = pumping period, in hours.

The equation is based on the reduction in specific capacity observed in 11 wells that were pumped for 24 hours or longer. The transmissivity (hydraulic conductivity multiplied by saturated thickness) of each hydrogeologic unit was estimated from the median values of the adjusted specific capacity data by a method described by Walton (1970). The hydraulic conductivity of the outwash aquifer (POWC) was calculated by dividing the estimated transmissivity by the median saturated thickness of sand and gravel penetrated by wells in the outwash.

Average hydraulic conductivity of the upper layer of the bedrock units were estimated by dividing the transmissivity by the median saturated thickness for each unit. The average hydraulic conductivity of the lower layer of each bedrock unit was estimated from the upper layer value. The following upper to lower layer ratios were assumed based on the percentage of carbonate rock in the hydrogeologic unit: Mahantango and Bloomsburg, 10 to 1; Onondaga-Old Port, upper Wills Creek, lower Wills Creek, 6 to 1; and Keyser-Tonoloway 4 to 1.

Directional bedrock hydraulic conductivity of the lower layer (KXL, KYL, and KZL) and upper layer (KXU, KYU, and KZU) was estimated from the average hydraulic conductivity. A 3-to-1 ratio for hydraulic conductivities along and perpendicular to the bedding strike was assumed for shales (Mahantango and Bloomsburg) and 6 to 1 ratio for all other aquifers that contained some carbonate rock as well as shale. The directional values were calculated so as to preserve the average hydraulic conductivity of the unit. The vertical hydraulic conductivity for each bedrock aquifer was assumed to be equal to the minimum horizontal value, KXU and KXL.

The bedrock hydraulic conductivity of each block was modified according to its topographic setting (TOPO) by PMULT, a topographic setting multiplier. PMULT was estimated from the specific-capacity data grouped according to topography and lithology.

A program modification was added to simulate the presence of a thin, discontinuous outwash aquifer without adding a separate layer. In grid blocks where saturated sand and gravel were present, the block transmissivity was calculated by summing the outwash and bedrock transmissivities.

Base values of hydraulic conductivity, directional bedrock hydraulic conductivity, and topographic multipliers were adjusted during calibration on a hydrogeologic unit basis. Block to block adjustment of hydraulic conductivity was not done.

Stream leakage

Surface-water and ground-water interaction was simulated using a head-dependent stream-leakage modification. The program modification allows for stream-aquifer flow within a grid block, the magnitude of which depends on the difference between a constant stream-stage altitude and aquifer head. Model input that define stream leakage include the following: stream-leakage coefficients for gaining stream (RCG) and losing stream (RCL) conditions, altitude of constant stream stage (RHSS), and altitude at which the aquifer becomes hydraulically detached from the stream (HB). In the following discussion, stream refers to any surface-water body.

Discharge from the aquifer to streams (gaining conditions) may be directly to the stream through the streambed or in the form of seeps and springs at the streambank or adjacent to the stream. The rate of discharge, therefore, depends on streambed length and width, streambank length and height, and effective width of the stream-discharge area, as well as local aquifer characteristics and head gradient. All factors, except for the difference in head between the stream and aquifer, were incorporated in the gaining stream-leakage coefficient. RCG, the gaining coefficient was computed by the following equation:

$$RCG = ARB\cdot\left(\frac{\underline{PER}\cdot\underline{WAT}\cdot\underline{(1+\sin STR)}\cdot\underline{AQC}}{BA}\right) \qquad (3)$$

where RCG = stream-leakage coefficient for gaining stream conditions;
ARB = arbitrary multiplier adjusted during calibration, the value is the same for all grid blocks;
PER = perimeter of stream;
WAT = stream type factor (Susquehanna River, 25; Fishing Creek, 10; lakes, 5; small streams, 1);
STR = angle between stream trend and bedding strike (0°-parallel, 90°-perpendicular);
ACQ = bedrock aquifer characteristics (Mahantango and Bloomsburg, 1; lower Wills Creek, 2; upper Wills Creek, Onondaga-Old Port, 3; Keyser-Tonoloway, 5); and
BA = block area.

Under losing conditions, leakage only can occur through the streambed, and therefore, RCL, the losing-stream coefficient, must be some fraction of the RCG coefficient. A 10 to 1 ratio between RCG and RCL was assumed. The stream-leakage coefficients were adjusted during calibration by adjusting ARB.

The altitude of constant stream stage for stream-leakage blocks, RHSS, was estimated from U.S. Geological Survey 7-1/2 minute quadrangle maps. The altitude at which the aquifer becomes hydraulically detached from the stream, HB, was assumed to be 5 ft (Susquehanna River and lakes), 3 ft (Fishing Creek) and 1 foot (small streams) below RHSS. Below this altitude of stream-infiltration cutoff, the rate of flow from stream to aquifer will not increase due to decreasing head in the aquifer.

Recharge and evapotranspiration

Model input used to simulate recharge to and evapotranspiration from the ground-water flow system include the recharge rate (QRE matrix), a coefficient of evapotranspiration (ECSS), the altitude above which maximum evapotranspiration occurs (EHB matrix), and the altitude at which evapotranspiration equals zero (EHSS matrix). Precipitation for 1981 was about 5 in. below normal. Based on published ground-water discharge data and considering that some water would be lost to ground-water evapotranspiration, the average recharge rate for 1981 was estimated to be 10 in./yr for model blocks without streams. In stream-leakage blocks, the recharge rate was reduced 10 percent for small streams and lakes, 50 percent for Fishing Creek, and 100 percent for Susquehanna River to account for those parts of the blocks that are assumed to be discharge areas. Due to the large percentage of discharge area, the recharge rate is reduced to about 9 in./yr when averaged over the entire model area.

The model computes the amount of evapotranspiration from the ground-water system based on ECSS, a head-dependent flow coefficient for evapotranspiration, and the relationships between the water-table altitude and the altitudes of maximum and zero evapotranspiration, defined, respectively, by the EHB and EHSS matrices. The rate is assumed to be a maximum value whenever the simulated water table is at and above EHB. The rate decreases linearly as the water table drops from EHB to EHSS and becomes zero at EHSS. The ECSS coefficient is defined by the following equation:

$$ECSS = \frac{MET}{EHB-EHSS} \qquad (4)$$

where

ECSS = evapotranspiration coefficient;
MET = maximum evapotranspiration rate;
EHB = altitude at and above which maximum evapotranspiration rate applies; and
EHSS = altitude at which evapotranspiration rate equals zero.

The maximum evapotranspiration rate was assumed to be 12 in./yr (McGreevy and Sloto, 1980). EHB and EHSS were defined to be 1 and 5 ft, respectively, below GRND.

Pumpage

Pumpage from the ground-water system is simulated in the model by using the WELL matrix. Pumping rates are entered for each grid block with significant ground-water withdrawals. Significant ground-water withdrawals during 1981 occurred at the Champion Valley Farms and Bloomsburg Mills well fields (fig. 2). A total rate of about 350 gal/min, was pumped from wells Co-197, Co-198, and Co-199 by Champion Valley Farms. The major water-bearing zones in well CO-199 are at a depth of less than 200 ft below land surface. The pumpage from the wells was simulated by assigning the WELL rate equal to 0.26 and 0.52 ft^3/s (cubic feet per second) in upper-layer blocks 65, 10 and 65, 11, respectively. The Bloomsburg Mills well field (wells Co-51, Co-52, and Co-53) is pumped for air conditioning during the summer and early fall at an estimated rate of up to 1,000 gal/min. No information is available on the distribution of water-bearing zones at this well field. It was assumed that all pumpage was from the upper 300 ft. The total estimated pumpage for the summer was averaged over the entire year for the steady-state simulation by assigning constant WELL rates of 0.36 and 0.38 ft^3/s in upper-layer blocks 12, 11 and 13, 11, respectively.

Boundary flow

Ground-water flow across the northern border of the model area was simulated using a head-dependent, boundary-flow modification. The program modification allows for steady-state flow to occur in and out of a boundary block. The rate of flow depends on the difference between the head in the boundary block and an assigned constant head at the aquifer boundary some distance from the model border. Input for the head-dependent boundary (BHSS matrix) and a coefficient of flow (BCSS) are based on aquifer characteristics.

The ground-water divide at the crest of the Berwick anticlinorium was selected as the aquifer boundary. The BHSS matrix, the constant head at the aquifer boundary assigned to each boundary block, was estimated from water-level data in the area and in similar hydrogeologic settings. The BCSS coefficient was calculated from the following equation:

$$BCSS = \frac{AQT \cdot BHC}{DAB \cdot DELX} \qquad (5)$$

where AQT = aquifer thickness;
BHC = hydraulic conductvity;
DAB = distance to aquifer boundary; and
DELX = width of boundary block.

BHC values were adjusted during steady-state calibration in order to simulate a reasonable amount of boundary flow. Initial values and calibration adjustments of BHC were consistent with those values and adjustments of other hydraulic conductivities used in the model.

Results

The ground-water flow model was assumed to be calibrated under 1981 average, steady-state conditions when simulated heads and base flows reasonably matched corresponding estimated heads and low flows. The goodness of the match between simulated and estimated heads was judged using three different methods. Simulated heads in the upper layer and estimated heads discretized from the water-table altitude map were compared using nonparametric statistics. Table 5 presents a statistical summary of the match between simulated and estimated water-table altitude by topography and bedrock hydrogeologic unit for the calibrated steady-state model. Simulated and estimated water-table gradients along selected rows also were compared (fig. 7). Finally, average water-table altitudes for 1981 for wells in the synoptic network were compared with simulated heads of the corresponding upper layer blocks (table 6). The match between simulated and estimated head for the calibrated steady-state model was judged to be acceptable within the range of uncertainty of the estimated water-table altitude.

Due to the lack of sufficient data, it was not possible to calibrate simulated and estimated heads in the lower layer. However, simulated heads in the lower layer were compared with simulated heads in the upper layer. In the slope area, heads in the lower layer generally were lower than those in upper layer. The median difference in head was 6 ft and the greatest difference was 20 ft. In the terrace area, heads in the lower layer generally were higher than those in the upper layer. The median difference in head was 0.2 ft and the greatest difference was 6 ft. These head relationships are in general agreement with the differences observed between shallow and deep bedrock wells in the slope and terrace settings.

The following table presents a comparison between changes in the rate of low flow for stream sections estimated from the August 2-3, 1982 synoptic measurement and corresponding model-simulated base flow under average steady-state conditions for 1981:

Stream section	Estimated (ft^3/s)	Simulated (ft^3/s)
A1-A2	+0.50	+0.80
B1-B2	+0.50	+0.41
C1-C2	+0.43	+0.34
D1-D2	-0.04	-0.04

Locations of stream sections are shown on plate 1. As previously mentioned, average base flow for 1981 was expected to be lower than the estimated low flow because of the slightly higher water table on August 2-3, 1982 as compared to 1981 average conditions. In addition, summer and early fall pumpage at Bloomsburg Mills, which significantly decreased low flow in stream A, cannot be adequately modeled in an average simulation. Considering these problems of comparison, the steady-state model simulates to a reasonable degree changes in the rate of base flow along the indicated stream sections.

Final bedrock hydraulic conductivity values for the calibrated model are presented in table 7. The final calibrated value for the hydraulic conductivity of the outwash aquifer is 40 ft/d. Calibrated stream-leakage multipliers are 1.0×10^{-5} and 1.0×10^{-6} for RCG and RCL, respectively. The hydraulic conductivity values for boundary-flow coefficients are 2.7×10^{-2} and 2.7×10^{-3} ft/d for the upper and lower layers, respectively.

The model-simulated water budget for 1981 is presented in table 8. The simulated water budget indicates an average inflow rate of 7.24 ft^3/s for 1981. Of this, 93 percent is recharge from local precipitation and 6.6 percent from

Table 5.--Summary of the departure of simulated and estimated 1981 average water-table altitudes

[N, number of grid blocks; Q_3, 25 percent quantile; MED, median (50 percent); Q_1, 75 percent quantile; positive number indicates simulated value less than estimated value; negative numbers indicates simulated value greater than estimated value]

Bedrock hydrogeologic unit	Terrace				Lower slope or draw				Upper slope				Upland				All			
	N	Q_3	MED	Q_1	N	Q_3	MED	Q_1	N	Q_3	MED	Q_1	N	Q_3	MED	Q_1	N	Q_3	MED	Q_1
Mahantango-Marcellus	338	0.0	-0.8	-3.8	-	-	-	-	-	-	-	-	-	-	-	-	338	0.0	-0.8	-3.8
Onondaga-Old Port	80	1.0	-0.2	-3.7	-	-	-	-	-	-	-	-	-	-	-	-	80	1.0	-0.2	-3.7
Keyser-Tonoloway	81	4.2	1.4	0.1	1	-	-2.6	-	9	12	4.4	2.1	-	-	-	-	91	4.4	1.8	0.1
Upper Wills Creek	77	6.3	2.8	0.5	51	9.2	3.5	0.7	32	3.8	-3.8	-15	15	9.5	-2.1	-22	175	6.1	2.2	-0.8
Lower Wills Creek	7	7.4	2.9	2.0	31	14	-2.6	-22	26	16	-3.2	-12	36	16	-4.0	-20	100	13	-2.6	-17
Bloomsburg	1	-	-6.4	-	2	-	3.9	-	6	4.6	-0.7	-13	5	7.3	-4.7	-23	14	4.6	-1.0	-12
All	584	1.6	-0.1	-3.0	85	10	2.7	-4.8	73	5.0	-1.6	-10	56	9.9	-3.9	-21				

boundary flow. Stream D is the only stream that serves as source; leakage to the aquifer from the stream is 0.05 ft^3/s. Sixty-two percent of the outflow is discharge to streams, 21 percent is pumpage, and 17 percent is evapotranspiration. Less than 5 percent of the inflow (recharge and boundary flow) to the upper layer flows downward into the lower layer in the slope area. Boundary flow to the lower layer is about 12 percent of the total boundary flow. These waters flow upward into the upper layer in the terrace area.

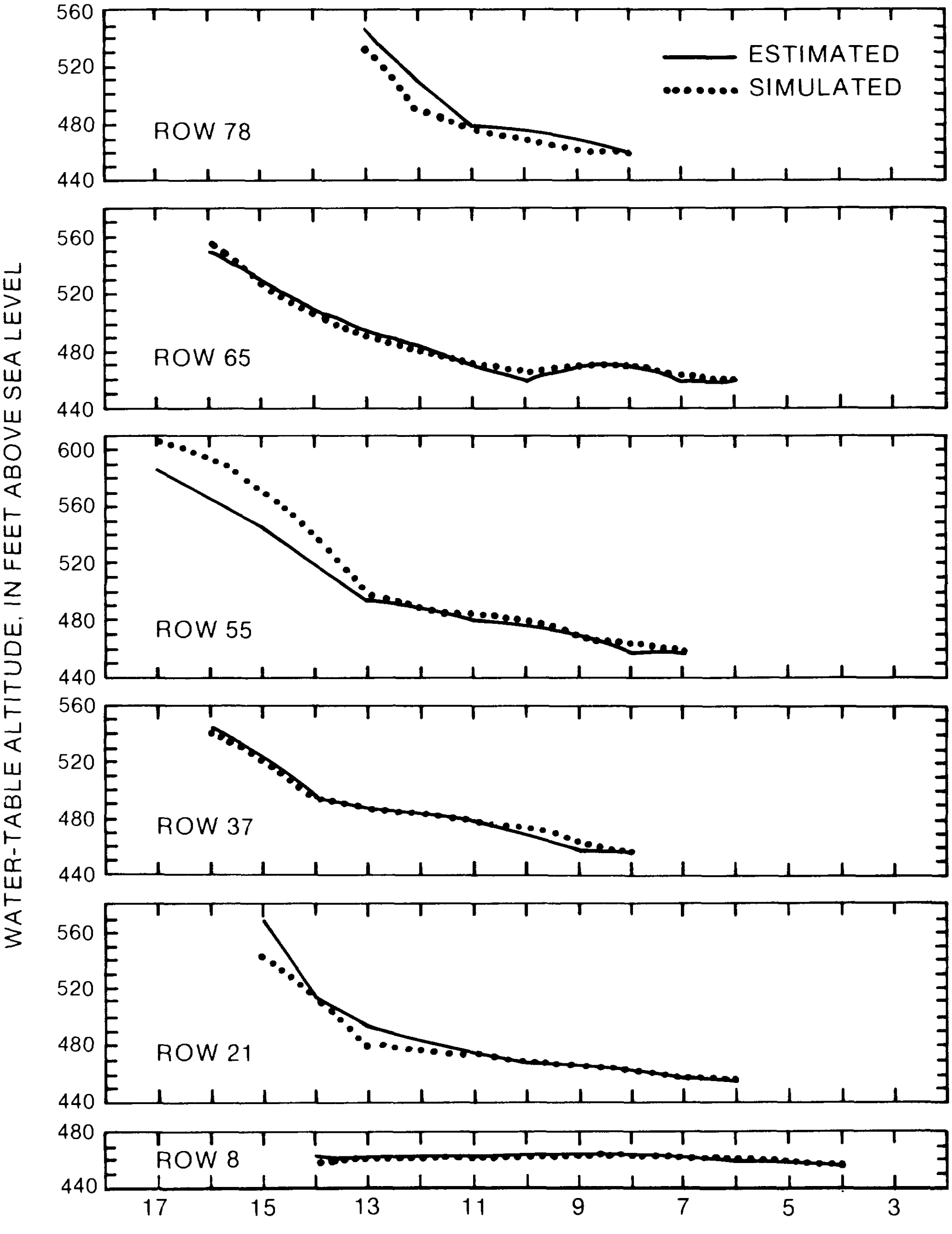


Figure 7.-- Comparison of estimated and simulated water-table altitudes along selected grid rows for the steady-state calibration simulation.

Table 6.--Comparison of simulated and observed 1981 average water-table altitudes for grid blocks with synoptic measurement wells

County number	Upper layer grid block(s)	Discretized[1/] setting	1981 average water-table altitude (feet above sea level) Observed	Simulated	Departure
Co-106	63,16	54	547	559	-12
136	25,12-25,13	31-41	489	[2/]484	5
154	6,10	11	463	463	0
161	54,12	31	486	489	- 3
162	60,13	42	506	496	10
164	55,12-56,13	42-42	504	[2/]493	11
165	60,8	11	470	466	4
167	70,8	11	469	469	0
168	77,9	31	472	464	8
183	71,10	21	475	476	-1
190	21,10	11	468	470	-2
203	67,8-67,9	11-11	469	[2/]472	-3
205	77,11	41	477	477	0
212	70,8	11	468	469	-1
306	60,9	11	473	471	2
310	17,12	31	470	471	-1
330	33,13-34,13	41-41	485	[2/]486	-1
373, 573	10,14	41	464	464	0

1/ First digit indicates bedrock hydrogeologic unit; 1, Mahantango-Marcellus; 2, Onondaga-Old Port; 3, Keyser-Tonoloway; 4, Upper Wills Creek; 5, Lower Wills Creek. Second digit indicates topographic setting; 1, terrace; 2, lower slope or draw; 3, upper slope; 4, upland. Underline indicates saturated sand and gravel.

2/ Average of two grid blocks.

Table 7.--Bedrock hydraulic-conductivity values for the calibrated steady-state model
[A dash indicates that a topographic multiplier is not needed because the unit is not found in that setting in the model area.]

Bedrock hydrogeologic unit	Upper layer hydraulic conductivity (ft/d)			Ratio of upper to lower layer hydraulic conductivity	Topographic setting multiplier			
	X1/	Y2/	Z3/		terrace	lower slope or draw	upper slope	upland
Mahantango-Marcellus	0.13	0.40	0.07	10:1	4	-	-	-
Onondaga-Old Port	.57	3.5	.29	6:1	4	-	-	-
Keyser-Tonoloway	.78	4.5	.39	4:1	4	2.7	0.5	-
Upper Wills Creek	.31	1.9	.16	6:1	4	2.8	.1	0.1
Lower Wills Creek	.24	1.2	.12	6:1	4	3	.5	.4
Bloomsburg	.09	.25	.05	10:1	4	2	.4	.4

1/ Horizontal across bedding strike

2/ Horizontal along bedding strike

3/ Vertical

Table 8.--Water budget for 1981 simulated during the steady-state calibration

Inflow (ft^3/s)		Outflow (ft^3/s)	
		Evapotranspiration	1.22
		Pumpage	1.52
Recharge	6.71	Stream leakage	4.49
Boundary flow	.48	Susquehanna River	2.18
Stream leakage		Fishing Creek	.63
Stream D	0.05	Stream A	.95
		Stream B	.40
		Stream C	.33
Total	7.24	Total	7.23

Transient Calibration

The model was calibrated under transient conditions for December 22, 1980 to April 21, 1982. The transient simulation was divided into 23 stress periods (table 9) based on the timing of recharge, pumpage, and synoptic water-level measurements. Since the ground-water system was not considered to be at steady state on December 22, 1980, a 3 month lead-in period was included prior to December 22 so that transient effects from earlier recharge and pumping conditions would be taken into account.

Hydrologic variables that changed during the transient simulation were rates of recharge (QRE), evapotranspiration (ECSS), and pumpage (WELL). Aquifer-storage variables (SOWC, SZL, and SZU) were adjusted during calibration in order to match simulated and observed water-level fluctuations. Observed fluctuations included water-level changes measured in wells with continuous recorders as well as changes measured in wells in the synoptic network.

Table 9.--Stress periods and selected hydrologic variables for the transient-calibration simulation

Stress period (mo/d/yr)	Number of days	Maximum recharge (in.)	Maximum recharge rate (in./day)	Maximum evapotranspiration rate (in./yr)	Pumpage rate (ft^3/s) in grid block1/				
					(65,10,2)	(65,11,2)	(63,11,1)	(12,11,2)	(13,11,2)
1) Lead in	90	1.39	0.015	12	0.26	0.52	0.00	0.00	0.00
2) 12/22/80-1/30/81	40	.66	.016	0	.26	.52	.00	.00	.00
3) 1/31 - 2/9	10	1.39	.139	0	.26	.52	.00	.00	.00
4) 2/10 - 2/28	19	2.90	.153	0	.26	.52	.00	.00	.00
5) 3/1 - 3/30	30	.23	.008	0	.26	.52	.00	.00	.00
6) 3/1 - 4/28	29	.54	.019	12	.26	.52	.00	.00	.00
7) 4/29 - 6/1	34	.70	.021	24	.26	.52	.00	.00	.00
8) 6/2 - 6/26	25	.71	.028	24	.26	.52	.00	1.11	1.11
9) 6/27 - 7/1	5	.14	.028	24	.26	.52	.00	.00	.00
10) 7/2 - 7/30	29	.29	.010	24	.21	.41	.00	1.11	1.11
11) 7/31 - 9/4	36	.00	.000	24	.22	.43	.00	1.11	1.11
12) 9/5 - 9/27	23	.46	.020	12	.26	.52	.00	.00	.89
13) 9/28 - 10/23	26	.12	.005	12	.26	.52	.00	.00	.00
14) 10/24 - 11/6	14	.93	.066	12	.26	.52	.00	.00	.00
15) 11/7 - 12/7	31	.58	.019	12	.26	.52	.00	.00	.00
16) 12/8 - 12/30	23	.36	.016	0	.26	.52	.00	.00	.00
17) 12/31 - 1/9/82	10	.81	.081	0	.26	.52	.00	.00	.00
18) 1/10 - 1/30	21	.56	.027	0	.26	.52	.00	.00	.00
19) 1/31 - 2/6	7	1.62	.231	0	.26	.52	.00	.00	.00
20) 2/7 - 2/26	20	.61	.030	0	.26	.52	.00	.00	00
21) 2/27 - 3/9	11	.17	.015	0	.00	.00	.78	.00	.00
22) 3/10 - 3/15	6	.96	.160	0	.00	.00	.78	.00	.00
23) 3/16 - 4/2	18	.30	.017	0	.00	.00	.78	.00	.00
24) 4/3 - 4/21	19	.81	.043	12	.00	.00	.78	.00	.00

1/ Third number refers to layer;

1, lower layer grid block; 2, upper-layer grid block.

Hydrologic Variables

Hydrologic variables in the steady-state calibration relating to aquifer geometry, except for the STRT matrix, were used in the transient calibration. Altitudes of constant stream stage, stream-infiltration cutoff, maximum and zero evapotranspiration, and head at the aquifer boundary from the steady-state simulation also were used. Hydraulic-conductivity values and stream-leakage and boundary-flow coefficients determined during the steady-state calibration were used in the transient simulation. The simulated, steady-state aquifer heads in the upper and lower layers were input as the STRT matrix at the beginning of the lead-in period for the transient simulation. The following sections discuss additional hydrologic variables of the transient calibration that were added or changed from the steady-state simulation.

Aquifer storage

Hydrologic variables that define the storage properties of the aquifers include the specific yield of the sand and gravel (SOWC), specific yield of the bedrock in the upper layer (SZU), and storage coefficient of the lower layer (SZL). Specific yields of the sand and gravel and bedrock hydrogeologic units in the upper layer were estimated from published values of Trainer and Watkins (1975), Carswell and Lloyd (1979), Becher and Root (1981), and a gravity-yield study of Appleman's Run, one mile north of the model area. The storage coefficient of the lower layer was assumed to be an order of magnitude less than the specific yield corresponding upper layer bedrock unit. A program modification was added to calculate a weighted specific yield for those blocks that contained saturated sand and gravel.

Recharge and evapotranspiration

Recharge rates and evapotranspiration coefficients were input for each stress period. Recharge was proportioned based on an evaluation of well hydrographs using a method similar to that of Rasmussen and Andreasen (1959). A maximum recharge of 10 inches was assumed for 1981. The estimated recharge and recharge rates for each period are given in table 9. The QRE rates were reduced in model blocks with streams as in the steady-state simulation.

Evapotranspiration was assumed to vary seasonally, the highest rates occurring from May to August, and no evapotranspiration occurring from December to March. Assumed maximum evapotranspiration rates for the stress periods are given in table 9. The ECSS coefficients were determined from equation 4 by using the maximum evapotranspiration rates.

Pumpage

Significant amounts of ground water were withdrawn during the transient-calibration period from the well fields of Champion Valley Farms (wells Co-197, Co-198, Co-199, and Co-505) and Bloomsburg Mills (wells Co-51, Co-52, and Co-53). In the Champion Valley Farms well field, wells Co-197, Co-198, and Co-199 were pumped during 1981 and until the beginning of March 1982. In March and April 1982, most of the water needed for the Champion Valley Farms plant was pumped from well Co-505. Pumpage from well Co-505 was assigned to the lower layer; the major water-bearing zone in this well is at a depth of 510 ft below land surface. The Bloomsburg Mills well field was pumped for air-conditioning water from June to September 1981. Pumpage rates used in the model that simulate this ground-water withdrawal are given in table 9.

Results

The model was assumed to be calibrated under transient conditions for December 22, 1980 to April 21, 1982 when simulated head changes reasonably matched corresponding observed head changes. The goodness of the match was judged by comparing simulated head changes in the upper layer with water-level changes observed in wells that were measured as part of the synoptic network.

The water-level changes observed in wells between the synoptic measurements of December 22, 1980, April 29, and December 8, 1981, and April 22, 1982 were compared with simulated changes in corresponding grid blocks for stress periods 2, 6, 15, and 24 (table 10). Observed and simulated water-level changes at the end of each stress period for selected wells and corresponding grid blocks were also compared (figs. 8 and 9).

Transient calibration was considered complete when further adjustment of storage coefficients did not improve the match between observed and simulated water-level changes. The storage input was adjusted during transient calibration by topographic setting as well as by hydrogeologic unit. Adjustment of storage based on topography is conceptually justifiable and helped to improve the calibration. Final storage values for the calibrated transient model are presented in table 11. Calibration of the model with water-level changes caused by variations in recharge was more successful than the calibration with changes caused by both pumpage and recharge. The variation in recharge is widespread and integrated over the entire model area; however, pumpage is mostly from discrete fractures. Simulated water-level changes under pumping conditions in grid blocks corresponding to wells Co-154, Co-310, and Co-448 could not be matched in detail with those observed in the wells without significant local adjustments of hydraulic conductivity and storage. Such local adjustment was not done because it is not known how well the pumping effects observed in the wells are representative of average block conditions.

Simulated rates of inflow and outflow for the stress periods in the transient calibration are presented in figure 10. Rates of recharge and pumpage correspond to values presented in table 9. Simulated evapotranspiration rates for the stress periods ranged from 0 to 2.5 ft^3/s. The maximum rate was simulated for the stress periods between April 29 and September 4, 1981. Simulated rates for water taken into storage in the aquifers were as much as 36 ft^3/s and simulated rates for water withdrawn from storage in the aquifers were as high as 5 ft^3/s. The minimum and maximum rates of water taken into or from storage in the aquifers were simulated for the stress periods between January 31 to February 6, 1982 and July 31 to September 4, 1982, respectively. Boundary flow into the aquifers for the stress periods was about 0.45 ft^3/s and showed minimal variation. Simulated stream-leakage rates ranged from 2.3 to 9.3 ft^3/s. The minimum rate was simulated for the period between July 31 to September 4, 1981. The maximum rate of discharge was simulated for the period between February 10 to February 28, 1981.

Table 10.--Comparison of simulated and observed water-level changes for the transient-calibration simulation

Well number	Upper layer grid block(s)	Water-level change, in feet[1/]: 12/22/80-4/29/81 Transient simulation period 2-6			4/29/81-12/8/81 Transient simulation period 6-15			12/8/81-4/22/82 Transient simulation period 15-24		
		Observed	Simulated	Departure	Observed	Simulated	Departure	Observed	Simulated	Departure
Co-106	63,16	6.2	5.7	-0.5	-4.1	-5.0	-0.9	12	7.8	-4.2
136	[2/]25,12-25,13	2.1	2.2	.1	-1.3	-2.0	- .7	2.2	2.7	.5
154	6,10	1.9	2.8	.9	-1.2	-3.6	-2.4	2.7	4.0	1.3
161	54,12	1.8	.8	-1.0	-1.1	- .6	.5	-	-	-
162	60,13	2.8	1.9	- .9	-2.5	-1.5	1.0	-	-	-
164	[2/]55,12-56,13	3.0	.5	-2.5	-2.3	- .3	2.0	4.1	.6	-3.5
165	60,8	2.3	1.4	- .9	-1.6	-1.0	.6	2.7	1.8	- .9
167	70,8	5.7	1.7	-4.0	-2.2	-1.1	1.1	4.5	2.6	-1.9
168	77,9	2.5	.7	-1.8	- .7	- .5	.2	3.5	1.0	-2.5
183	71,10	11	1.9	-9.1	-	-	-	-	-	-
190	21,10	2.4	1.1	-1.3	-1.8	-1.2	.6	3.4	1.6	-1.8
203	[2/]67,8-67,9	3.6	2.2	-1.4	-2.4	-1.5	.9	5.8	3.8	-2.0
205	77,11	3.4	2.6	- .8	-2.6	-2.2	.4	5.3	3.5	-1.7
212	70,8	3.6	1.7	-1.9	-2.6	-1.1	1.5	3.7	2.6	-1.1
306	60,9	2.3	1.6	- .7	-1.7	-1.1	.6	3.2	1.9	-1.3
310	17,12	2.6	4.1	1.5	-1.7	-6.1	-4.4	3.6	6.0	2.4
330	[2/]33,13-34,13	1.6	1.3	- .3	-1.4	-1.1	.3	2.0	1.8	- .2
373,573	10,14	2.2	2.0	- .2	-1.4	-2.5	-1.1	2.8	3.0	.2
		Median		-1.0	Median		.5	Median		-1.3

[1/] Positive value indicates water-level rise; negative value indicates water-level decline.

[2/] Simulated change is the average for two grid blocks.

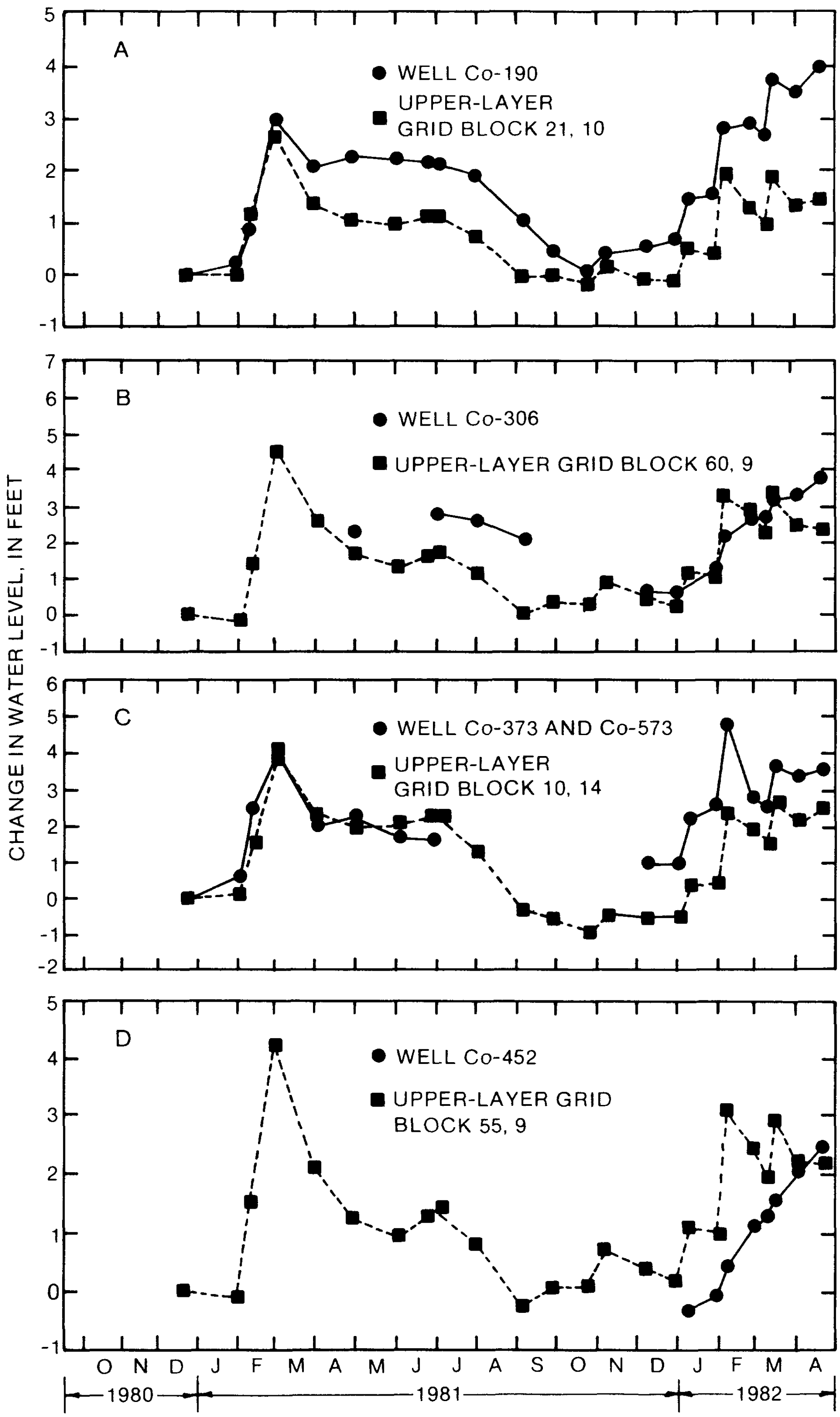


Figure 8.-- Comparison of observed and simulated water-level changes for selected wells and corresponding grid blocks.

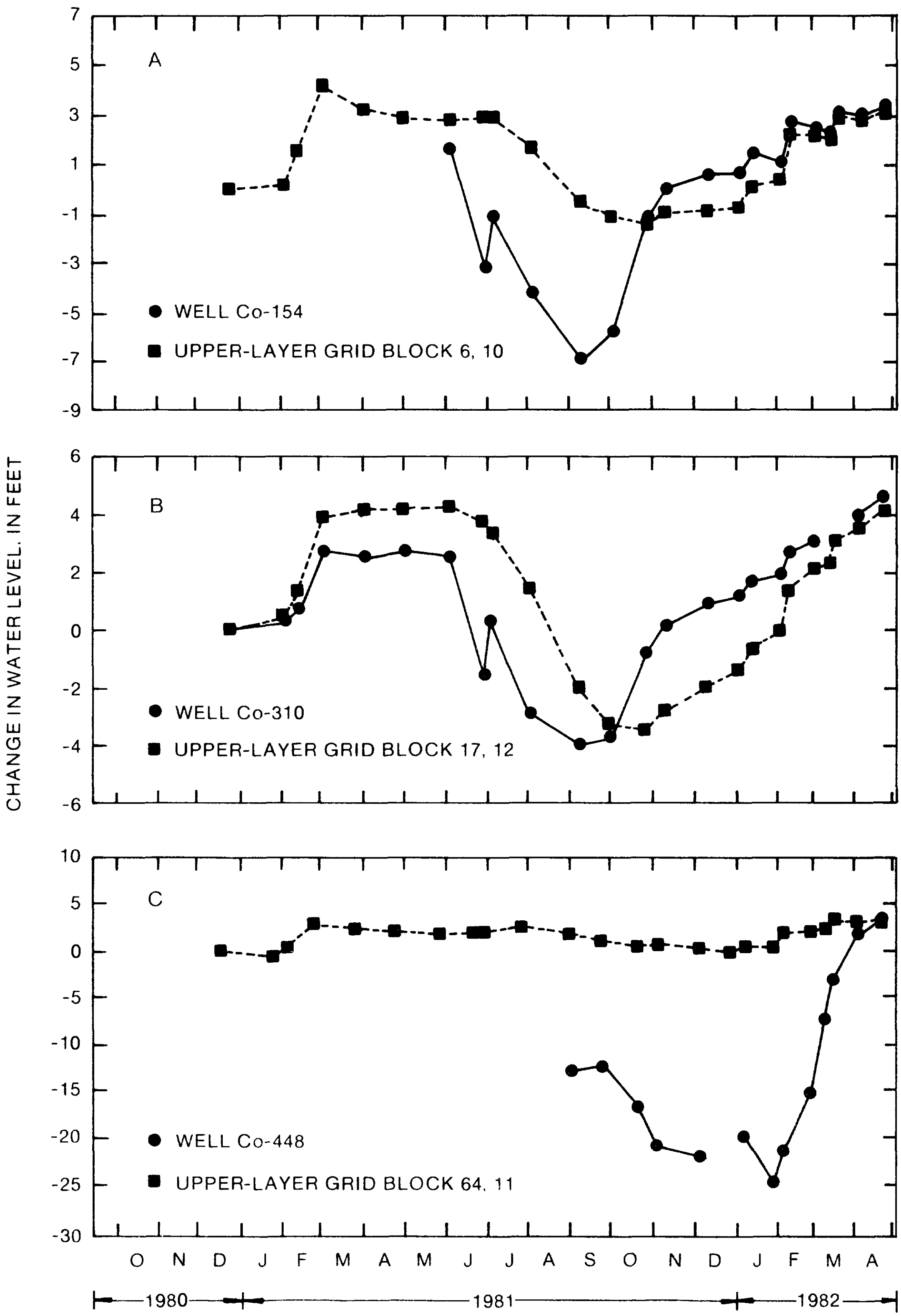


Figure 9.--Comparison of observed and simulated water-level changes affected by pumpage for selected wells and corresponding grid blocks.

Table 11.--Aquifer-storage values for the calibrated transient model

	Upper layer, specific yield[1]	
Hydrogeologic unit	Terrace-lower slope or draw	Upper slope-upland
Glacial outwash	0.20	
Mahantango-Marcellus	.02	-
Onondaga-Old Port	.06	-
Keyser-Tonoloway	.08	0.04
Upper Wills Creek	.04	.02
Lower Wills Creek	.03	.02
Bloomsburg	.015	.01

1/ Lower layer storage coefficient is one-tenth upper-layer specific yield.

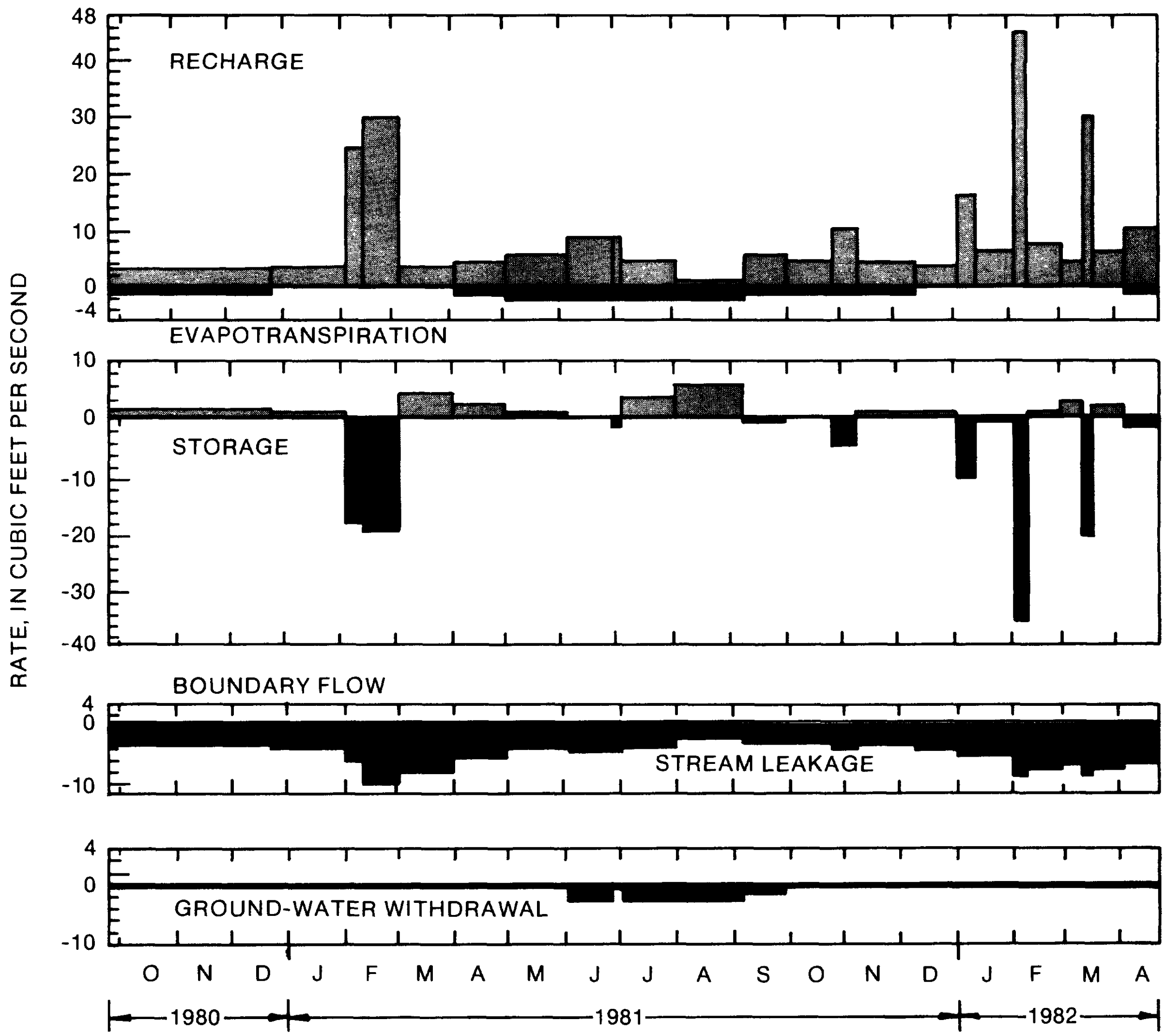


Figure 10.--Simulated rates of inflow (positive rate) and outflow (negative rate) for the transient-calibration simulation.

MODEL USE

The model incorporates many generalizations and assumptions about the flow of ground water in a complex aquifer system. However, within limitations, the flow model can be used as a tool to guide the development and management of ground water in the model area. The model can be used to simulate, in a general way, the effect of both natural and artificial stresses on the ground-water system. Given a stress or combination of stresses, typically a change in ground-water recharge or withdrawal, the model can be used to simulate the corresponding average effects on water-table altitudes and stream leakage. Stresses that could be simulated include: (1) drought, (2) drought recovery, (3) reduced recharge due to urbanization, (4) increased recharge due to spray irrigation or recharge basins, (5) pumpage, (6) well injection, and (7) excavation dewatering. The more widespread the stress, the greater the probability that it can be successfully simulated by the model.

The transient model is generally more useful than the steady-state model in simulating the impact of stresses because recharge and, in many cases, ground-water withdrawals are transient in nature, varying seasonally and annually. The ground-water flow system typically never reaches steady-state conditions because the stresses are changing constantly. Examples of transient simulations under hypothetical stress conditions are presented in the following section.

Three, 10-year simulations were made using the transient model under hypothetical stress conditions in order to demonstrate model use. The effects of three, ground-water development schemes were simulated. The natural stress simulated, recharge varying on a seasonal and annual basis, was kept constant for all three examples.

Hydrologic Variables used in Example Simulations

Recharge, evapotranspiration, and pumpage variables were input into the calibrated transient model. Maximum annual recharge was varied from year to year in order to simulate times of decreased (drought) and increased recharge (drought recovery). A four-year drought of increasing severity was simulated in years 4, 5, 6, and 7. Hypothetical annual recharge for the example simulations is presented in the following table:

	Year									
Maximum annual	1	2	3	4	5	6	7	8	9	10
recharge (in.)	13	12	12	10	9	8	8	12	14	12

Based on long-term hydrographs from wells Co-1 and Co-45 (fig. 3), the example simulations were divided into two stress periods per year. The highest recharge and lowest maximum evapotranspiration rates were input for the October to March simulation period. The simulated, annual-high water-table altitude and base flow occur at the end of the October to March period. The lowest recharge and highest maximum evapotranspiration rates were input for the April to September simulation period. The simulated, annual-low water-table altitude and base flow occur at the end of the April to September period.

Recharge was distributed as a percentage of the total annual recharge per 6-month period based on the distribution of recharge in 1981 used in the transient calibration. In the transient-calibration simulation about 70 percent of the recharge for 1981 occurred during the October to March period and 30 percent during April to September period. Maximum evapotranspiration rates for the periods were the weighted averages of the respective values from the transient calibration simulation. Maximum evapotranspiration rates of 4 and 20 in./yr were input for the October to March and April to September periods, respectively.

Three hypothetical ground-water development schemes were modeled in the example simulations (table 12). In simulation I, pumpage was continued unchanged from that in 1982. In simulation II, pumpage was increased at the Champion Valley Farms well field. In addition, pumpage was increased at the Columbia County Development Authority well field (wells Co-204 and Co-205), which had insignificant withdrawal in 1982. In simulation III, pumpage from three hypothetical well fields was added to the pumpage in simulation II. The simulated pumpage from two of the hypothetical well fields was seasonal. Year-round pumpage was increased about 165 percent between simulations I and II. Year-round and seasonal pumpage was increased about 25 and 180 percent, respectively, between simulations II and III.

Results

The impact of the natural and pumping stresses of the example simulations on water-table altitudes and base flows was evaluated by observing the simulated effect on water levels in selected grid blocks (fig. 11) and leakage to selected streams (figs. 12 and 13). In addition, the simulated differences in water-table altitudes between the three pumpage schemes at the end of the hypothetical drought was contoured (plate 6).

As indicated by figure 13 and plate 6a, the increased pumpage between simulations I and II had a significant effect on water-table altitudes and base flow in the eastern part of the model area. Increased drawdown exceeded 20 and 35 ft between and near the eastern well fields at the end of the drought, respectively. In simulation I, stream C became a losing stream only during April to September in years 6 and 7. In simulation II, stream C was a losing stream in both seasonal periods in years 5, 6, and 7.

Additional pumpage in simulation III had less of an impact on water-table altitudes and base flow than the increase in pumpage between simulations I and II (fig. 13 and plate 6b). Increased drawdown was only slightly greater than 15 ft near the eastern well fields; and 10 ft near the central and western well fields. Simulated leakage rates for stream C were virtually unchanged between simulation II and III. Stream A became a losing stream during April to September in year 7 in simulation III.

The impact of the overall increased pumpage between simulations I and III are shown on plate 6c. Increased drawdown exceeded 20 ft between the three western most well fields. Increased drawdown near the middle of these three exceeded 40 ft.

Table 12.--Stress periods and hypothetical pumpage for the example simulations

Example simulation	Stress period	Pumpage rate (ft^3/s) for years 1-10 in grid block (63,11,1)	(12,11,2)	(13,11,2)	(65,11,2)	(77,11,2)	(25,12,2)	(39,11,2)	(53,12,2)
I	Oct-Mar	0.78	-	-	-	-	-	-	-
	Apr-Sep	.78	0.56	0.56	-	-	-	-	-
II	Oct-Mar	.78	-	-	0.78	0.50	-	-	-
	Apr-Sep	.78	.56	.56	.78	.50	-	-	-
III	Oct-Mar	.78	-	-	.78	.50	0.50	-	-
	Apr-Sep	.78	.56	.56	.78	.50	.50	0.50	0.50

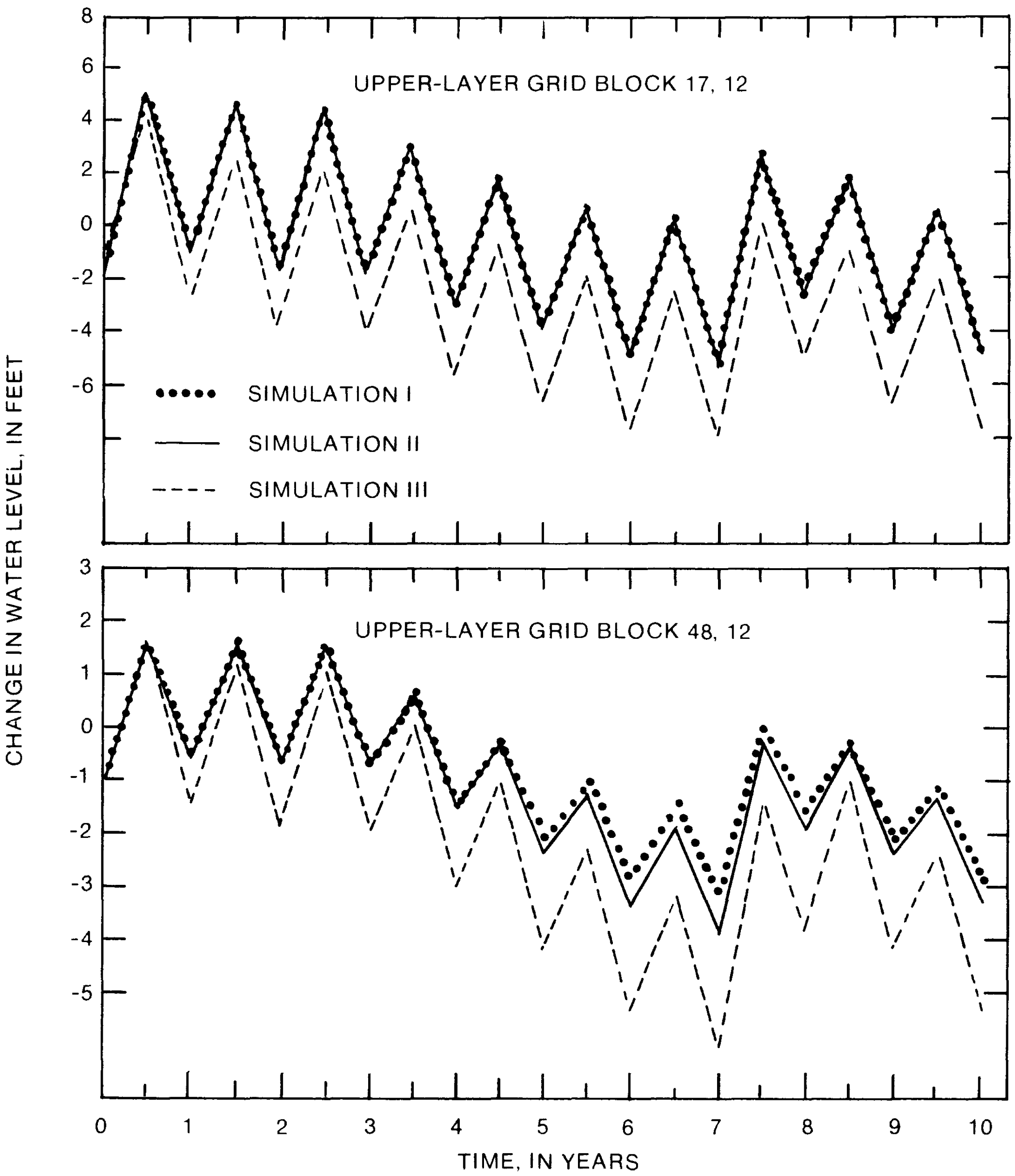


Figure 11.--Water-level changes simulated during example simulations I, II, and III for selected grid blocks.

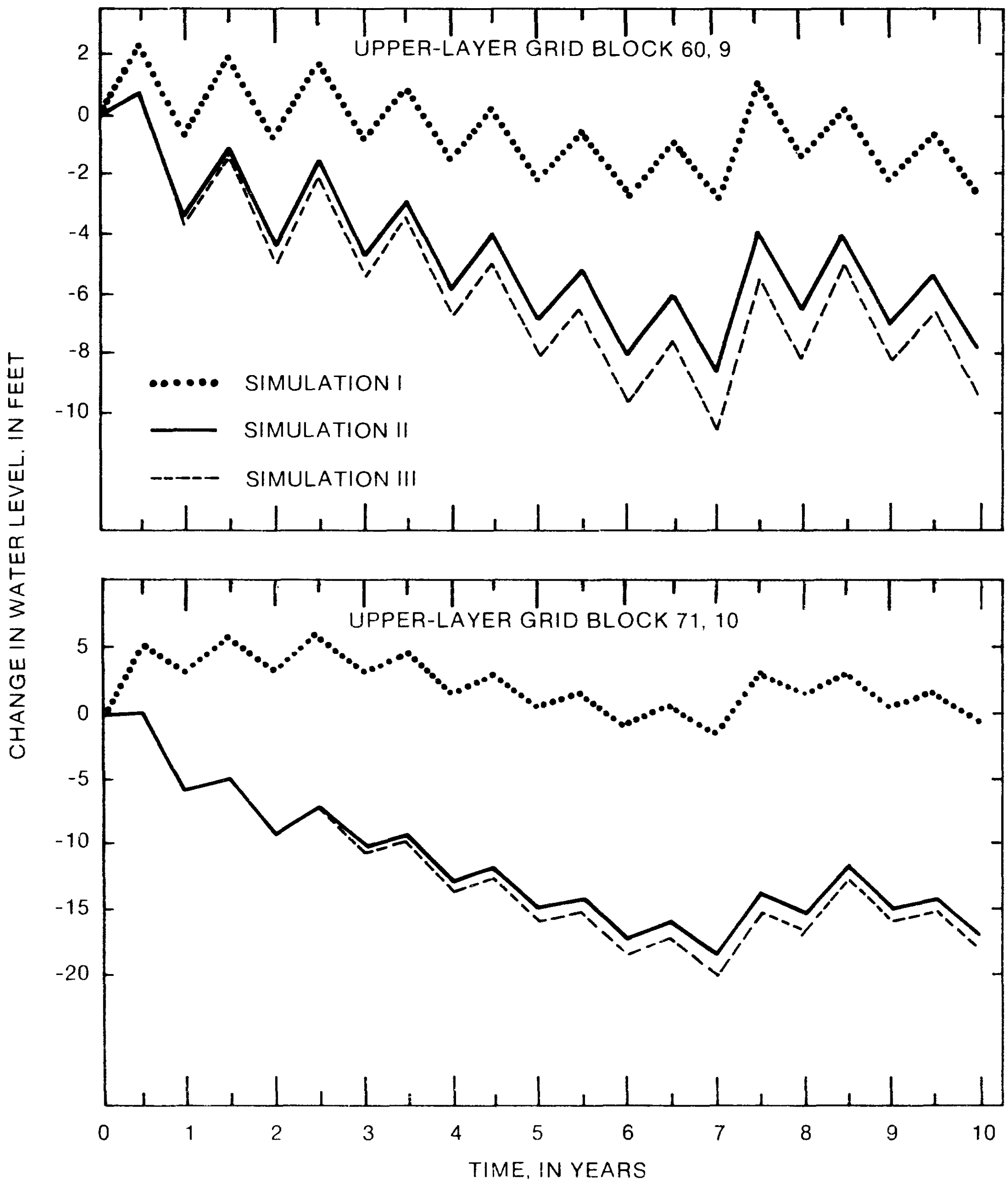


Figure 11.--Water-level changes simulated during examples I, II, and III for selected grid blocks--Continued.

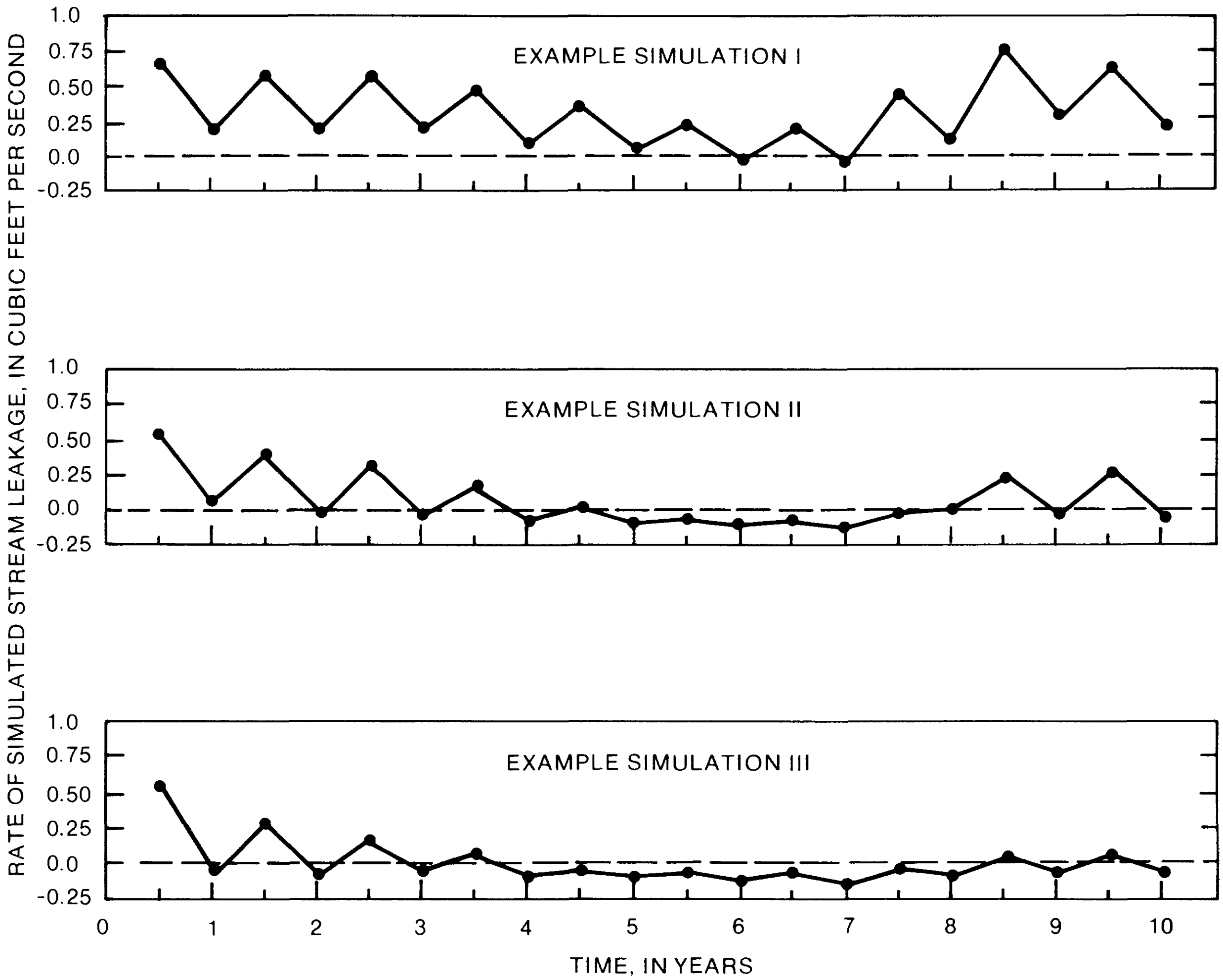


Figure 12.--Simulated stream leakage to stream C for the example simulations.

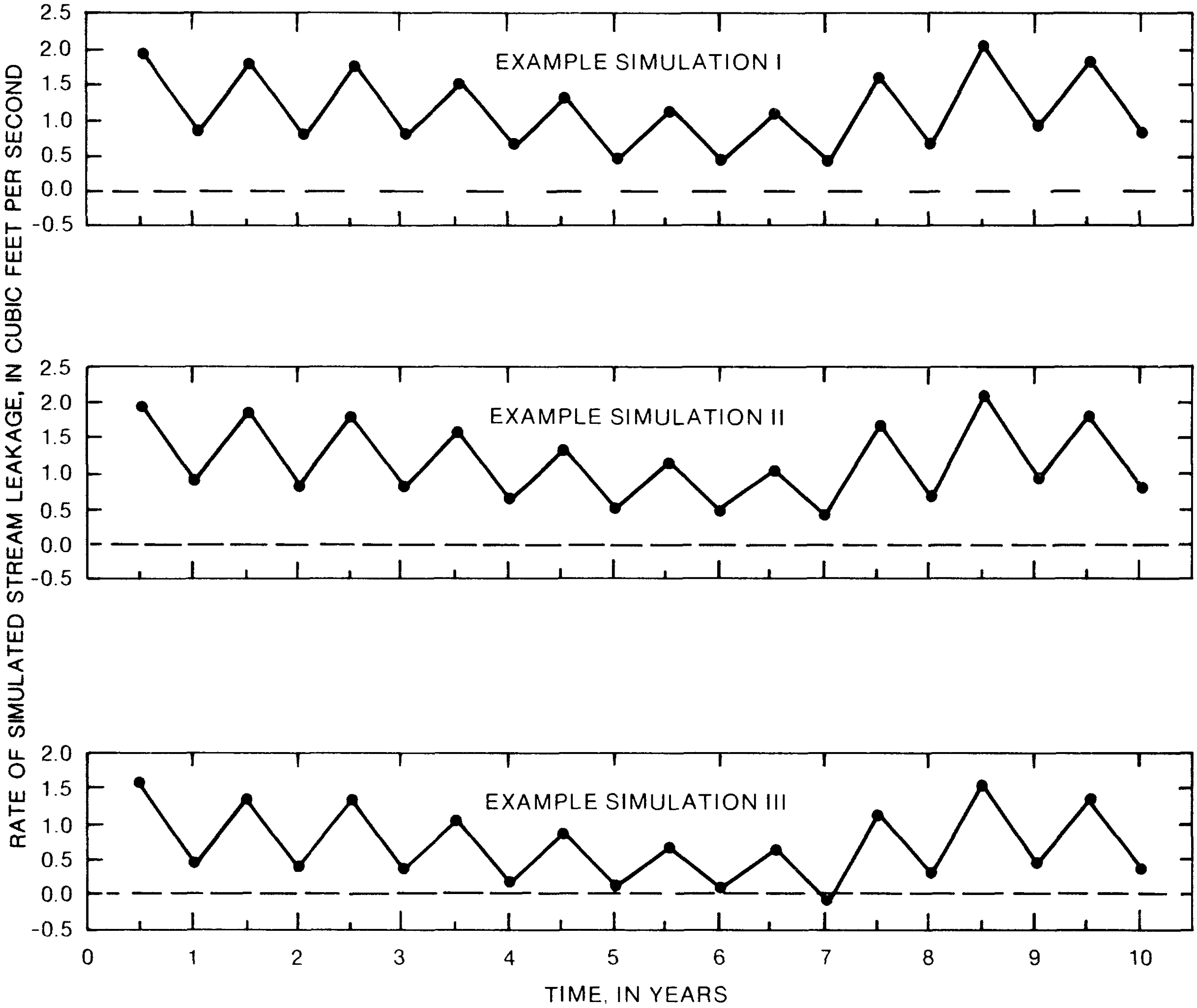


Figure 13.--Simulated stream leakage to stream A for the example simulations.

SUMMARY

A numerical model of ground-water flow was developed for the bedrock and glacial-outwash aquifers along the Susquehanna River in Columbia County, Pennsylvania. The 10.3-mi^2 model area, located on the north side of the Susquehanna River between Berwick and Bloomsburg, east-central Pennsylvania, is underlain by clastic and carbonate bedrock and glacial-outwash sand and gravel. The bedrock and sand and gravel aquifers act as a single complex, water-table system.

The two-layered model developed during the project simulates ground-water flow based on the following hydrologic conditions: (1) hydraulic conductivity differs according to hydrogeologic unit, structure, topography, and depth below land surface; (2) aquifer storage differs according to hydrogeologic unit, topography, and depth below land surface; (3) head-dependent stream leakage differs according to stream type and trend, hydrogeologic unit, and type of leakage (gaining or losing conditions); and (4) head-dependent boundary flow differs according to distance to the aquifer boundary and hydraulic conductivity.

The flow model was calibrated under average steady-state conditions for 1981 and transient conditions from December 22, 1980 to April 21, 1982. In the steady-state calibration, hydraulic conductivity and stream leakage variables were adjusted in order to simulate estimated water-table altitudes and low flows. The simulated steady-state water budget for 1981 indicates an average inflow rate of 7.24 ft^3/s. Ninety-three percent of inflow is recharge from precipitation and 7 percent is boundary flow. Outflow is through leakage to streams (62 percent); pumpage (21 percent); and evapotranspiration (17 percent). In the transient calibration, aquifer-storage variables were adjusted in order to simulate observed water-level changes caused by natural and pumping stresses. The transient model was more successful in matching the natural water-level fluctuations observed in wells due to variations in recharge than those changes caused by pumping stresses. Recharge is a more widespread stress than ground-water withdrawal, which occurs, to a large degree, along discrete bedrock fractures.

Three 10-year, hypothetical stress schemes were simulated with the calibrated transient model to demonstrate model use. The general impact of three pumpage schemes on water-table altitudes and base flows were evaluated under hypothetical natural stress conditions (drought and drought recovery). In simulation I, pumpage was continued unchanged from that in 1982. Year-round pumpage was increased about 165 percent between simulation I and II. Between simulations II and III, year-round and seasonal pumpage was increased about 25 and 180 percent, respectively. Increased pumpage between simulation I and II increased drawdown more than 20 ft in the eastern part of model area. In simulation I, a stream in the eastern part became a losing stream in the dry seasonal period during two years of the drought. In simulation II, the stream became a losing stream in both seasonal periods during three years of the drought. Increased pumpage between simulations II and III increased drawdown about 15 ft in the eastern part and 10 ft in the central and western parts. In simulation III, a stream in the western part became a losing stream in the dry seasonal period during one year of the drought.

REFERENCES CITED

Becher, A.E. and Root, S.I., 1981, Groundwater and geology of the Cumberland Valley, Cumberland County, Pennsylvania: Pennsylvania Geological Survey 4th ser., Water Resource Report 50, 95 p.

Carswell, L.D. and Lloyd, O.B., 1979, Geology and ground-water resources of Monroe County, Pennsylvania: Pennsylvania Geological Survey 4th ser., Water Resource Report 47, 61 p.

Gerhart, J.M. and Lazorchick, G.J., 1988, Evaluation of the ground-water resources of the lower Susquehanna River basin, Pennsylvania and Maryland: U.S. Geological Survey Water-Supply Paper 2284, 128 p.

Inners, J.D., 1981, Geology and mineral resources of the Bloomsburg and Mifflinville quadrangles and part of the Catawissa quadrangle, Columbia County, Pennsylvania: Pennsylvania Geologic Survey, 4th ser., Atlas 164cd, 106 p.

Johnston, H.E., 1970, Ground-water resources of the Loysville-Mifflintown quadrangles, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 27, 96 p.

McGreevey, L.J. and Sloto, R.A., 1980, Development of a digital model of ground-water flow in deeply weathered crystalline rock, Chester County, Pennsylvania: U.S. Geological Survey Water-Resources Investigations Report 80-2, 42 p.

Rasmussen, W.C. and Andreasen, G.E., 1959, Hydrologic budget of Beaverdam Creek basin, Maryland: U.S. Geological Survey Water-Supply Paper 1472, 106 p.

Taylor, L.E., Werkheiser, W.H., DuPont, N.S., and Kriz, M.L., 1982, Ground-water resources of the Juniata River Basin, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 54, 131 p.

Trainer, F.W. and Watkins, F.A., 1975, Geohydrologic reconnaissance of the upper Potomac River Basin: U.S. Geological Survey Water-Supply Paper 2035, 68 p.

Trescott, P.D., 1975, Documentation of finite-difference model for simulation of three-dimensional ground-water flow: U.S. Geological Survey Open-File Report 75-438, 32 p.

Trescott, P.D. and Larson, S.P., 1976, Documentation of finite-difference model for simulation of three-dimensional ground-water flow: U.S. Geological Survey Open-File Report 76-591, 21 p.

Walton, W.C., 1970, Groundwater resource evaluation: McGraw-Hill Series in Water Resources and Environmental Engineering, 664 p.

Williams, J.H., Carswell, L.D., Lloyd, O.B., and Roth, W.C., 1984, Characterization of ground-water circulation in selected fractured-rock aquifers using borehole temperature and flow logs: Proceedings of NWWA/EPA Conference on Surface and Borehole Geophysical Methods in Ground-Water Investigations, San Antonio, Texas, p. 842-852.

Williams, J.H. and Eckhardt, D.A., 1987, Ground-water resources of the Berwick-Bloomsburg-Danville area, east-central Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 60, 80 p.