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BASIN, OHIO.

The Ohio State University,
Ph.D., 1978
Geology

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HYDROGEOLOGY AND DIGITAL MODELING OF THE
BURIED-VALLEY AQUIFER AND THE SILURIAN-DEVONIAN
CARBONATE AQUIFER IN THE SCIOTO RIVER BASIN, OHIO

DISSERATION

Presented in Partial Fulfillment of the Requirements for
the Degree Doctor of Philosophy in the Graduate
School of The Ohio State University

By

Mustafa Ahmad Ukayli, B.S., M.S.

* * * * *

The Ohio State University

1978

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I. INTRODUCTION

Area of Study

The Scioto River Basin, one of the principal basins in Central Ohio, comprises an area of about 6510 square miles, or more than one-sixth of the total area of the State (fig. 1). The basin, approximately rectangular, extends southward through central Ohio for about 130 miles. It is about 55 miles wide in the upper part and narrows to about 25 miles in the southern third of the basin.

The Scioto River, the main stream in the basin, is about 231 miles long. It originates on till plains in the northwestern part of the basin, flows southeastward for about 60 miles, and then turns abruptly southward near Green Camp for about 171 miles, eventually emptying into the Ohio River at Portsmouth. Major southwestward and southward flowing tributaries include the Olentangy River, Alum, Big Walnut, Walnut, Salt, and Beaver Creeks. Southeastward and eastward flowing tributaries include Darby, Mill, Dear, Paint, Sunfish, and Scioto Brush Creeks. Columbus, with a population (1974) of about 741,000, is the major population center. Other significant cities include Marion, Delaware, Circleville, Chillicothe, and Portsmouth. Included in the basin, are parts or all of the following counties, listed from north to south; Crawford, Morrow, Marion, Hardin, Allen,
PLEASE NOTE:

This dissertation contains colored photographs, which will not reproduce well. Several pages have very small print. Filmed in best way possible.

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Figure 1. Map of Ohio Showing Outline of the Scioto River Basin and the physiographic provinces.
Auglaize, Logan, Union, Delaware, Knox, Licking, Franklin, Madison, Champaign, Clark, Fayette, Pickaway, Fairfield, Hocking, Ross, Greene, Clinton, Highland, Vinton, Jackson, Pike, Adams, and Scioto County.

Purpose and Scope of Study

The growing demand for water for domestic, municipal, and industrial use in the Scioto River Basin has led to increased emphasis on this major resource and on the management of aquifers throughout the area. Anticipated growth and population increase necessitates a dependable water source. Ground water in the Scioto River Basin is expected to meet this need.

Increasing industrialization in the basin and rising public demands for water necessitates proper management of the aquifers to satisfy the needs for good quality water. The expected increase in population from 1.1 million in 1974 to about 2 million by the year 2020 will focus the attention of planners on the Scioto River Basin as a source for public water supply (O.D.N.R., 1977, Table 28). Industrial water use will be greatly influenced by the cost of obtaining the supply and by pollution abatement requirements; under these restrictions, ground water is seen in the future as the most favored water source.

The purpose of this study is to evaluate the ground water resources of selected parts of the basin and to develop digital hydrologic models that can provide detailed projections of the ground-water situation to the year 2020. These models should enable
one to predict the configuration of the water table under different management techniques over specified periods of time.

This evaluation required the extensive use of a computer in order to analyze and process the required data, including:

1. Analyses of streamflow data which permit the determination and evaluation of baseflow.

2. Interpretation of well field data in order to determine transmissivity (T), discharge rate (Q), storage coefficient (S), head (H), and infiltration rates, both induced and natural.

The above data were processed by a modified version of a computer program published by Prickett and Lonnquist (1971). Potentiometric surface maps representing existing (1976) conditions were prepared for two representative aquifers, part of the glacial outwash aquifer along the Scioto River and the limestone aquifer that underlies the Western part of the basin. The model for the buried valley aquifer was based on data obtained from a series of aquifer tests in the vicinity of Piketon. The second model, which includes most of the carbonate aquifer within the Scioto River Basin, was based on wells, test holes, and aquifer tests conducted at widely scattered points.

After calibration and verification of the models, predictions of future management techniques that would tend to guarantee continuous ground-water availability were made. Estimates of the future ground-water withdrawals were incorporated and simulations of the predicted water table configurations and responses for 1980, 1990, 2000 and 2020 were generated.
Physiography and Topography

The Scioto River Basin lies within two subsections of the Central Lowland Physiographic Province (fig. 1). The Till Plains subsection, which occurs in the northernmost two thirds of the basin, is characterized by a youthful surface that, for the most part, is nearly flat to gently rolling. This region is covered by clay-rich till generally less than 50 feet in thickness and locally absent along many stream channels. Low relief end moraines, kames, eskers, and outwash deposits are scattered throughout the area.

The Till Plains, which occupy the major part of the glaciated region north of Circleville, slope southward from the northern boundaries of the basin at a rate of about seven feet per mile (Sharp, 1932). For the most part, the surface is nearly featureless except for stream channels and floodplains. The valleys of the Scioto and Olentangy Rivers are the most conspicuous (fig. 2). North of Columbus these valleys rarely exceed a half mile in width and, for the most part, their floors are so narrow that they are entirely occupied by the stream. The southward slope of the Till Plains is interrupted by a series of low end moraines, such as the St. Johns, Broadway, Powell, London, and Bloomingburg moraines (fig. 23). The moraines, which represent a sequence of recessional positions of the Wisconsinan ice margin, loop across the basin from west to east. They are convex to the south and are ridges of irregular width and usually low relief. Many of the larger tributaries of the Scioto River flow along the distal margins of these end moraines. Kames and eskers are also present and appear as long
narrow ridges and rounded hills of water-lain silt, sand, and gravel. Though not common in the basin, locally they form prominent topographic features in Pickaway, Fairfield, Franklin, Delaware, and Morrow Counties. Their local relief is generally less than 60 feet.

Glaciated and unglaciated plateaus of the Appalachian Plateau subprovince lie to the south and east of the Scioto basin. A major west-facing escarpment consisting of north trending scarps and glacial terraces marks the boundary between the Till Plains and the Appalachian Plateaus. The glaciated Plateau is sparingly capped with thin Illinoian-age till which forms a narrow belt about 6 miles wide and crosses the basin the vicinity of Chillicothe. Generally this region is less rugged than the adjacent unglaciated area. The unglaciated section of the Appalachian Plateaus is a rugged forest-covered region characterized by narrow valleys and steep hillsides. It is dissected and well-drained by mature streams that have cut their channels deep into the steep hillsides. In some places the streams have cut deep gorges that are partly filled with alluvial sand, gravel, and silt. The exposed bedrock is predominantly shale of Devonian age, overlain locally by Berea and Cuyahoga Sandstones of Mississippian age and Sharon conglomerate of Pennsylvanian age. The sandstone and conglomerate form the cap rock of many of the hills in the unglaciated plateau and are responsible for their higher relief.

The rugged topography and steep-sided valleys in this part of the basin enhance runoff from precipitation and hence contribute the greatest amount of flow to the streams in the basin.
Relief

The highest elevations occur along the northwestern margin of the basin just east of Bellefontaine in Logan County, where several hills rise to elevations of 1,434 to 1,550 feet. The lowest area, approximately 480 feet above sea level, occurs in the southernmost part of the basin at Portsmouth where the Scioto drains into the Ohio River (fig. 2). The maximum difference in elevation within the basin is approximately 1,070 feet.

Drainage

The Scioto River is the longest stream in the basin. In fact, the Scioto is one of the longest tributaries of the Ohio River. The river rises west of Kenton in the northwestern corner of the basin and flows southeastwards for about 60 miles along the distal margin of the Wabash Moraine and then bends to flow south for about 171 miles to its confluence with the Ohio River at Portsmouth. Between Kenton and Prospect its average gradient is about 1.5 feet per mile. The river then flows through a shallow gorge between Prospect and Columbus and the gradient increases to about 5 feet per mile. Below Columbus the gradient flattens to about 1.7 feet per mile and continues at this grade to its confluence with the Ohio River.

All the major tributaries of the Scioto River are consequent upon the southward slope of the Till Plains and occupied their present courses as soon as the ice retreated (Sharp, 1932, p. 39). The courses of most of these streams are governed to
Figure 2. Map of the Scioto River Basin showing relief and drainage pattern. Numbers show approximate elevation in feet above sea level. (Modified after O.D.N.R., 1963).
a certain degree by the position of end moraines. Because the Till Plains slope southward, the major tributaries also flow in this general direction in valleys that are almost parallel for considerable distances before they gradually converge to flow into the Scioto. The streams thus have a poorly developed dendritic pattern with a near-parallel alignment of the major streams (fig. 2).

South of Chillicothe, the Scioto River enters the rugged unglaciated Plateau Province where tributary streams flow in deep narrow valleys that lie as much as 500 feet below the uplands; they have little or no floodplains. In this region the Scioto Valley becomes slightly narrower and the floodplain is one to two miles wide as compared to 2 miles wide further upstream.

The total number of perennial and intermittent streams in the basin is 489 and they have a total length of about 4,200 miles (O.D.N.R., 1960). The drainage efficiency of the basin is thus relatively poor due to the presence of relatively long streams with low gradients. Table 1 shows some characteristics of the principal streams in the basin.

**Climate**

Ohio has a continental climate that is relatively homogeneous from one season to the next. Summers are moderately warm and humid, whereas winters are moderately cold and cloudy.
TABLE 1: Major Streams in the Scioto River Basin, their lengths, gradients, and drainage areas. (Data from Report No. 12, Ohio Water Plan Inventory, Division of Water, 1960).

<table>
<thead>
<tr>
<th>Name of Stream</th>
<th>Length (Miles)</th>
<th>Elevation at Source</th>
<th>Elevation at Mouth</th>
<th>Average Gradient (ft. per mile)</th>
<th>Drainage Area (sq. miles)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Scioto River</td>
<td>230.8</td>
<td>1010</td>
<td>481</td>
<td>2.3</td>
<td>6509.9</td>
</tr>
<tr>
<td>Olentangy River</td>
<td>88.5</td>
<td>1189</td>
<td>702</td>
<td>5.5</td>
<td>536.3</td>
</tr>
<tr>
<td>Alum Creek</td>
<td>55.8</td>
<td>1120</td>
<td>715</td>
<td>7.4</td>
<td>200.7</td>
</tr>
<tr>
<td>Big Walnut Creek</td>
<td>74.2</td>
<td>1165</td>
<td>667</td>
<td>7.0</td>
<td>556.7</td>
</tr>
<tr>
<td>Walnut Creek</td>
<td>49.8</td>
<td>1120</td>
<td>652</td>
<td>9.4</td>
<td>280.7</td>
</tr>
<tr>
<td>Salt Creek</td>
<td>45.4</td>
<td>1003</td>
<td>559</td>
<td>9.7</td>
<td>553.4</td>
</tr>
<tr>
<td>Mill Creek</td>
<td>37.8</td>
<td>1074</td>
<td>840</td>
<td>6.2</td>
<td>185.5</td>
</tr>
<tr>
<td>Darby Creek</td>
<td>78.7</td>
<td>1170</td>
<td>643</td>
<td>6.8</td>
<td>556.6</td>
</tr>
<tr>
<td>Deer Creek</td>
<td>67.1</td>
<td>1130</td>
<td>621</td>
<td>7.6</td>
<td>408.4</td>
</tr>
<tr>
<td>Paint Creek</td>
<td>94.7</td>
<td>1120</td>
<td>586</td>
<td>5.6</td>
<td>1142.7</td>
</tr>
<tr>
<td>Sunfish Creek</td>
<td>26.5</td>
<td>795</td>
<td>530</td>
<td>10.0</td>
<td>144.6</td>
</tr>
<tr>
<td>Scioto Brush Creek</td>
<td>36.0</td>
<td>781</td>
<td>506</td>
<td>7.6</td>
<td>273.5</td>
</tr>
</tbody>
</table>
The mean daily maximum air temperature in July ranges from 84 degrees F. in the northern portion of the Scioto basin to 88 degrees F. in the southern portion (Anderson and King, 1976). The mean daily minimum air temperature in January ranges from 19 degrees F. in the north to 25 degrees F. in the south.

The growing season starts in late May in the north to mid-May in the south and it ends around mid-September in the north to late September in the south.

Precipitation, which increases southward in the basin, reflects a trend opposite that of temperature (fig. 3). Portsmouth has an average of 42 inches of precipitation per year, and the amount decreases northward to an average of 34 inches per year in Hardin County.

The average annual streamflow in the basin ranges between 10 and 15 inches per year (fig. 4). It is lowest in the upper part of the basin when precipitation is generally less than 37 inches. Here, the lower streamflow may also be related to the absence of buried outwash, which tend to contribute a substantial amount of groundwater runoff to streamflow. The mean flow increases southward (downstream) due to increase in precipitation (as much as 44 inches per year) and to the presence of extensive outwash deposits, both buried and surficial, that lie along the main stream and many of its tributaries.

There is also a correlation between areas of ground-water recharge and higher precipitation. This is especially evident in Highland County where both precipitation (fig. 3) and ground-water runoff are high.
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Figure 3. Mean annual precipitation in the Scioto River Basin (After Ohio Dept. Nat. Res., 1963).
Figure 4. Average streamflow in the Scioto River Basin, in ches (based on data from Walker et al., 1963 Table 13).
II. GENERAL GEOLOGY

Pleistocene Geology

Glaciers covered central and northern Ohio during four different stages within the last million years, but definite evidence of the earliest two (Nebraskan and Kansan) is difficult to establish in the Scioto Basin. Deposits of questionable Nebraskan age have been described in southwestern Ohio (Leverett, 1929; Ireland, 1943; Merrill, 1953; Teller, 1970) and northern Kentucky (Leverett, 1929; Lessig, 1961; Campbell et al., 1974). These deposits include a variety of features ranging from a deeply weathered outwash in eastern Ohio, leached gravel rich in clay and manganese southwest of Cincinnati, and crystalline erratics in northern Kentucky. Teller (1970) described a poorly developed, truncated paleosol on till underlying a Kansan till in Decatur County, Indiana. Kansan-age deposits in the Cincinnati region have been described by Teller (1970) as a deeply-weathered till found in valleys and on interstream uplands. Possible pre-Illinoian drift was reported by Rosengreen (1970) in central Highland County, Ohio. It consists of a till unit overlying a paleosol developed in sand and gravel. Quinn (1974) identified two truncated Yarmouthian (?) soil profiles in Ross County.
Because both soils directly overlie the bedrock surface they indicate that pre-Illinoian drift may be absent from Ross County.

Illinoian drift in the Scioto Basin consists of thin, patchy ground moraine that extends northeast-southwest across the basin from the northwestern corner of Hocking County to the east-central part of Highland County (fig. 5). The till is relatively thin, ranging in thickness from an average of 34 feet in Ross County to 53 feet in Highland County.

The Illinoian till is highly jointed and commonly contains secondary clay accumulations along joint margins in the lower part of the oxidized zone (Quinn, 1974). Where exposed, the depth of oxidation averages 12 feet. The oxidized till is yellowish-brown whereas unoxidized till is dark-gray.

Illinoian kames and eskers are present locally. Kames are most plentiful in Highland County where they occur in groups consisting of several connected hills forming kame and kettle topography in which the highest kame rises to a height of about 200 feet above the surrounding plain. These eskers and kames generally consist of permeable gravel and sand, rendering them good sources of ground-water supply, and ground-water runoff to the streams that flow from their margins.

Kempton and Goldthwait (1959) distinguished two levels of Illinoian outwash in the Scioto Valley south of Chillicothe. The higher outwash terrace is composed predominantly of silty sand with some gravel, whereas the lower terrace contains much
Figure 5. Glacial map of Ohio showing the glacial deposits of the Scioto River Basin.
coarser and cleaner sand and gravel.

Most of the glaciated part of the Scioto Basin is covered by Wisconsinan drift, which forms broad areas of ground moraine separated by a succession of end moraines. Ground moraine is the most widespread glacial feature of the basin; it forms low and flat surfaces in the north and hilly terrain in the south where it occurs as a thin cover of till on the dissected bedrock. In these areas, ground moraine is composed of till that contains an unsorted mixture of 10-40 percent sand, 30-50 percent silt, and 15-40 percent clay (Goldthwait, 1969). The till is of low permeability, but in many places it contains lenses of sand and gravel that vary both in thickness and areal extent. These lenses are major sources of ground water for many communities and farms. Ground moraine ranges in thickness from a feather edge to more than 293 feet, and averages about 50 feet.

The end moraines are characterized by thicker and more hummocky deposits of drift. They loop across the basin varying in width from less than a mile to more than 2.5 miles (fig. 5). They range from 10 to 300 feet in thickness and may rise a few tens of feet above the surrounding ground moraine. End moraines usually consist of till that is interbedded with stratified lenses of sand and gravel. Since end moraines are more permeable than ground moraine, they are major sources of ground-water runoff, at least locally.

Wisconsinan till weathers to a calcareous clay-silt loam soil that is yellowish-brown to dark. The oxidized zone may
extend 10 to 20 feet below the surface. Unoxidized till is blue gray, blocky, and relatively hard. It is usually reported as blue clay or "hardpan" by water well drillers.

The sparsely scattered kames and eskers form long ridges and knobs, consist predominantly of sand and gravel, and are important sources of water. Many are mined for sand and gravel. Eskers and kames generally stand high above the surface of the till plain, particularly in northwestern Delaware County, eastern Franklin County, and along the buried Scioto valley at Circleville in Pickway County.

Ancient stream channels that were cut into the bedrock during the Teays and Deep Stage drainage systems were filled with sand and gravel during interglacial periods. The outwash deposits were formed by melt-water flowing away from the ice margins during the glacial advances of the Illinoian and Wisconsin stages (figs. 6 and 7). These buried valleys constitute one of the most promising source of groundwater in the Scioto Basin. They form large reservoirs for storage of groundwater and, at the same time, can receive direct recharge from adjacent streams when head conditions permit.

The thickest and most extensive outwash deposit in a buried valley extends along the existing Scioto River channel between Columbus and Portsmouth (fig. 7). Major tributaries include Big Darby and Deer Creeks. In its northern part, the buried valley is about eight miles wide and, in some places, contains as much as 250 feet of outwash (fig. 8). The valley narrows considerably southward, from 2.5 to 1.5 miles in width,
Figure 6. Map showing outwash deposits and locations of cross sections (fig. 8) along the buried valley. (Adapted from Goldthwait et al., 1960.)
Figure 7. Map showing configuration of the buried valley along the Scioto River. (Adapted from Walker et al., 1965.)
Cross-section in northern Pickaway County just south of Lockbourne Air Force Base.

Cross-section in Ross County. Green Township.

Cross-section in Scioto County Valley Township.

Figure 8. Representative cross-sections of the buried valley across the Scioto River (after Walker et al., 1965).
and may contain 70 to 80 feet of outwash. The outwash in the buried valley is very porous and permeable and has a relatively high recharge potential.

**Bedrock Geology**

**Regional Structural Setting**

The Scioto River drainage basin lies between three major structural basins - the Michigan, Appalachian, and Illinois basins (fig. 9). These structural basins are separated by the northward plunging Cincinnati Arch, which splits in northwestern Ohio into two major limbs, the Findlay and Kankakee Arches. The Findlay limb trends slightly east of north along the western margin of the Michigan basin and influences the dip of the beds in central Ohio. The Kankakee limb trends to the northwest, separating the Michigan and Illinois structural basins. The Scioto River drainage basin lies on the eastern flank of the Cincinnati Arch but near its crest. The rocks dip gently to the east-southeast at a slope of about 20 to 30 feet per mile and thicken gradually towards the axis of the Appalachian Basin. Hence the oldest bedrock, predominantly shale of late Ordovician age, subcrops beneath the drift in the deeper parts of some buried valleys in the southwestern part of the basin. The youngest sedimentary bedrock, largely sandstones of Pennsylvanian age, occurs along the southeastern border as ridges capping the hilltops of the Allegheny Plateau Province.
Figure 9. Regional structure map showing study area, Basin and Arch Systems, and generalized structure on Precambrian Surface (After Ulteig, 1964).
Bedrock Stratigraphy

Exclusive of glacial and younger deposits, the stratigraphic sequence of formations exposed in the Scioto Basin consists, in ascending order, of Upper Ordovician, Silurian, Devonian, Mississippian, and Pennsylvanian rocks (figs. 10 and 11). The lithostratigraphic succession of formations described in this section is based on the work of many previous investigators, such as Stout and Lamey (1940), Stout et al. (1943), Carman (1947), Norris and Spicer (1958), Schmidt (1958), Wolfe et al (1962), Ulteig (1964), and Owens (1970).

The Upper Ordovician System

In the southwestern margin of the Scioto Basin, Upper Ordovician rocks subcrop beneath the glacial drift in the buried valleys in central and western Madison County. These strata consist of a thick sequence of shale and interbedded thin layers of limestone. The shale is soft, greenish-blue and calcareous; the limestone layers are fossiliferous, hard, argillaceous, and average 1 to 5 inches in thickness. The Ordovician shales are generally of very low permeability and are not significant sources of water. Water is stored only near the surface in fractures and along the bedding planes. Water obtained from wells tapping these shaly deposits is likely to be highly mineralized with excessive concentration of iron, sulfate, total dissolved solids, and with a high hardness.
Figure 10. Geologic map of Ohio showing the general geology in the Scioto River Basin (After Ohio Geological Survey).
### Figure 11. Generalized Stratigraphic Sequence of Rock Formations in the Scioto River Basin.
The Silurian System

Silurian strata subcrop beneath glacial till and crop out along many streams in much of the western half of the Scioto Basin. They consist of a thick series of massive limestones and dolomites that are separated near the base by the Osgood Shale (fig. 11). The upper part of the Brassfield Limestone is massive, hard, and crystalline towards the top. The basal portion of the Brassfield is massive and has a coarse granular texture. The Brassfield limestone is generally unimportant as an aquifer and yields from wells drilled in localized areas can be 10 to 60 gpm. However, the water is sulfurous. The Brassfield Limestone generally marks the lower stratigraphic limit of ground-water supplies.

The Osgood shale, soft and calcareous, contains a few thin layers of dolomite. It forms a moderately thick (about 45 feet) aquiclude that separates the Lockport dolomite of the Niagara Group from the underlying Brassfield.

Dolomites in the Lockport and Bass Islands Group constitute the bedrock in the western half of the Basin and form the lower part of the extensive Silurian-Devonian carbonate aquifer. The Lockport is a light gray to white, fine to coarsely grained, bedded dolomite that is pure. It crops out as an irregular band in the southwestern part of the basin, mainly in Adams and Highland Counties. Where thicker than a hundred feet, it is usually characterized by reef-like structures and loses its discernible bedding character. The Lockport is generally permeable where exposed and where weathering has widened the solution cavities in
the rocks. Permeability decreases where the Lockport is overlain by the Bass Islands Group.

The Bass Islands Group consists of about 400 feet of thin to medium-bedded argillaceous dolomite which varies in color from brown to bluish-gray. This group contains thin beds of gypsum and anhydrite randomly distributed near the middle of the section. Of major significance is a widespread zone of generally high permeability that occurs in the lower part of the Bass Islands, just above the Lockport. This zone, about 10 to 15 feet thick, is characterized by large cavities and solution channels. It has been identified in many places by drillers as a major source of water in many deep wells in areas including Plain City and Harrisburg, Ohio (Norris, 1956). Known as the "Newburg zone", it is believed to have been developed as a result of solution by percolating ground water. Caliper logs of test wells in southwest Ohio show the especially prominent zone as about 15 feet thick with the base lying from 3 to 5 feet above the Lockport (fig. 12).

The Devonian System

The Columbus Limestone of Middle Devonian age is separated from the underlying Bass Islands Group by a major disconformity. The disconformity represents a break in the sequence whereby Lower Devonian rocks, the Oriskany Sandstone and Detroit River Group, have been removed by erosion. This represents an extended erosional period during Early Devonian time. Generally, the disconformity is recognized by a thin layer of black, tarry
Figure 12. Caliper logs of selected wells in western Ohio showing cavity zone (Newburg) above the Lockport Dolomite. Datum is top of Lockport Dolomite. (After Norris and Fidler, 1973.)
bitumen (Stout and Lamey, 1940), but locally it is recognized by a conglomerate at the base of the Columbus Limestone (Schmidt, 1958).

The lower portion of the Columbus Limestone is massive, brown, and contains a high percentage of magnesium. It grades upward into a massive bluish-gray limestone that has a low magnesium content. The formation is jointed and commonly the joints open to fissures that lead to sinkholes and underground passages (Stout and Lamey, 1940). The Columbus Limestone is the principal consolidated aquifer in central Ohio and wells tapping it commonly yield appreciable amounts of water. Yields vary from 100 to 500 gpm, as seen in Figure 32, which shows ground-water yields in the Scioto Basin.

The Delaware Limestone conformably overlies the Columbus Limestone. The Delaware is an argillaceous, cherty, evenly-bedded limestone that is rather hard and dense. It is generally considered a poor source of ground water.

Together with the Silurian and Devonian carbonates, the Columbus and Delaware limestones constitute the main carbonate aquifer in the Scioto Basin. They are collectively referred to as the "Big Lime" by the drillers in Ohio, and are the most uniformly productive of the consolidated aquifers in the basin.

The Ohio and Olentangy Shales of Devonian age overlie the Delaware, directly underlie glacial till, and crop out along valleys in a belt about 10 miles wide that extends from Ross County in the south to Crawford County in the north. These
deposits overlap the Delaware and Columbus limestone southward to Pickaway County. The Olentangy Shale is a bluish-gray soft argillaceous shale that contains limestone concretions near its base. The overlying Ohio Shale is black in the upper part but becomes brownish toward the bottom of the section. It grades in structure from massive near the top to thinly laminated or fissile towards the bottom. The lower part also contains pyrite which occurs as nodules.

Regionally, the Olentangy and Ohio Shales are aquicludes and therefore very poor sources of ground water. These aquicludes blanket the underlying carbonates in the eastern half of the basin and impede recharge.

The Mississippian System

In the eastern part of the Scioto Basin rocks of Mississippian age crop out in a belt about 20 to 25 miles wide that trends north-south. In general these rocks can be grouped into five different formations which are, in ascending order, the Bedford Shale, Berea Sandstone, Sunbury Shale, Cuyahoga Formation and the Logan Formation.

The Bedford Shale is a gray to brown, soft, laminated, argillaceous shale. It gradually changes to siltstone and thins southward from Delaware. In the northern part of the basin it is about 90 feet thick (Westgate, 1926). The Berea Sandstone lies unconformably on the Bedford, and consists in the northeastern and central part of the basin of moderately cemented fine-grained sandstone,
but changes to a siltstone southward. Its estimated thickness in the state varies from 5 to 225 feet but averages about 45 feet (Stout et al., 1943). Schmidt (1958) estimated that its thickness in Franklin County varies from 55 feet in the north to about 5 feet in the south. Sandstones in the Berea Formation, at least locally, yield fair amounts of ground water (10 to 70 gpm). The Sunbury Shale is a brown to black fissile shale that is highly carbonaceous. Its thickness ranges between 12 and 35 feet. The Lower Cuyahoga Formation occurs only along the extreme southeastern margin of the basin (Jefferson County) where it consists of a series of alternating thin layers of sandy shale and thin to massive fine-grained sandstones. Locally, the massive sandstones yield fair to good (5 to 50 gpm) amounts of ground water, particularly near their outcrops. The Blackhand member of the Cuyahoga Formation subcrops in a narrow belt in the central eastern edge of the basin in eastern Delaware, Franklin, Fairfield, Pickaway and Hocking counties. It consists of massive deposits of coarse-grained sandstones that are highly cross-bedded and grade into rather pebbly conglomerates. The Blackhand ranges in thickness from 50 to 200 feet and often forms one great ledge. In areas where it subcrops below the glacial drift fracturing is more developed and wells drilled in such areas can yield 10 to 125 gpm of good quality water. The Blackhand thickens to the northeast and becomes a major source of brine water, and often becomes a reservoir for oil and gas in many areas outside the basin boundaries.
The Logan Formation is an undifferentiated group of shales, sandstones, and conglomerates. They are generally a poor source of ground water and occur well above the regional water table in the basin.

**The Pennsylvanian System**

Only three of the four formations that comprise the Pennsylvanian system are present in the area. The Pottsville and Allegheny Formations crop out in eastern Scioto and central Jackson Counties. The Pottsville ranges in thickness from 0 to 140 feet, and thins westward, whereas the Allegheny is as much as 80 feet thick. The sandstone, shale, clay, and coal of the lower Pottsville Formation cover the largest portion of these areas whereas the sandstone, shale, clay, coal and limestone of the Allegheny Formation are confined to the hilltops in the extreme southeastern parts of the area. The Conemaugh Formation consists of an undifferentiated sequence of thin layers of sandstone, shale, and coal and is confined to some of the higher hilltops in the extreme southeastern margin of the basin.
III. SURFACE-WATER HYDROLOGY

Stream Flow

The stream-flow characteristics of a drainage basin generally depend on the amount of precipitation, temperature, wind velocity, area and shape of the basin, slope, amount and kind of vegetation, the type of soil, and the lithologic character of material adjacent to the stream. Nearly one third of the rainfall in Ohio flows as surface runoff. Total runoff is a combination of surface runoff, derived from precipitation, and ground-water runoff from the underlying shallow aquifers in hydrologic connection with the stream.

Surface runoff is that part of the precipitation that falls on the ground and stream surfaces and finds its way into the stream channel without infiltrating into the soil. Ground-water runoff is defined as that part of the precipitation that infiltrates the soil and reaches the ground-water table, than at a later period percolates into the stream channel.

The average annual stream flow in the Scioto River Basin ranges from 10 to 15 inches (fig. 4). It is low in the upper part of the basin, largely because of the low precipitation but also because of the absence of buried glacial outwash, which would tend to contribute ground-water runoff. This part of the basin is covered by a thick deposit of clay-rich till of low permeability. The average flow
increases southward as does precipitation. The presence of extensive outwash deposits of very permeable sand and gravel contributes extensively to each stream's runoff. Some tributaries pick up additional water from ground-water seepage along the margins of end moraines because they commonly contain layers of sand and gravel. This is especially evident in Highland County where both precipitation and ground-water runoff are high (fig. 3).

Stream Hydrographs and Flow Duration Curves

Stream hydrographs and flow-duration curves can be used to analyze the characteristics of the flow of streams. A stream hydrograph is a graphical representation of fluctuations in flow arranged in chronological order. The hydrographs of the two neighboring streams of equal drainage area in a given geographical locality may differ considerably. The differences can be attributed to geologic structure, permeability throughout the basin and particularly of the material adjacent to the stream channel, size of the drainage area, topography, vegetative cover, intensity and duration of rainfall, type of soil, antecedent soil moisture, and land use. If the drainage area is composed largely of low permeability material, such as till or shale, then the stream would tend to be flashy with high peak flows, and even floods, because of the low storage capacity of the basin. However, where the deposits adjacent to a river are composed of permeable material, such as sand and gravel, or the drainage area is characterized by abundant permeable material, such as glacial outwash or limestone with many solution openings, then
the stream hydrograph would tend to be more uniform. Flooding occurs less often and the storage capacity of the basin is high.

The flow-duration curve shows the frequency of occurrence of various rates of stream flow (Cross and Hedges, 1959). It is a cumulative frequency curve prepared by arranging all discharges of record for a particular stream gaging station in order of magnitude and then dividing them according to percentage of time during which specific flows are equalled or exceeded. The discharge rates are then plotted against the percent of time they are equalled or exceeded on a logarithmic probability paper and a flow-duration curve is obtained.

Flow-duration curves are helpful in estimating the yield of a stream, and in comparing the flow characteristics of different streams. They are also useful in evaluating the storage capacities, because low flows may be derived entirely from ground-water runoff. The shape of the duration curve gives a clear picture of the nature of the flow. If the curve is gently sloping the stream is said to be uniform and has few floods and a large ground-water discharge. A steep curve indicates that the stream is flashy and storage effects are very low.

**Flow Characteristics**

Using daily stream flow data for a nearly normal year (1967)* for

*1967 was a near normal year because streamflow in most basins in the state was about equal to the long term average and generally the greatest since water year 1961. There was no extremely low flow, nor were there any damaging floods in Ohio during water year 1967. (Tuller, 1975)*
the major streams in the basin and by applying computer separation techniques, several hydrographs and flow-duration curves were obtained. A discussion of the hydrograph separation techniques and listing of the computer programs used to develop the hydrographs and flow-duration curves appear in Appendix A.

The hydrographs and flow-duration curves were used in conjunction with a knowledge of the geology of the watershed in order to compare the characteristics of flow among the streams and in order to analyze the changes in stream regimen (figs. 13 to 30). Table 2 presents a summary of the data on which the analysis was based. Deviations in the average flow shown in table 2 and figure 4 are due to the fact that the data in table 2 represent one year of record whereas the data in figure 4 represent an average of 30 years of records.

In general, the hydrographs and flow-duration curves reflect the heterogeneous nature of the flow characteristics in the basin. Wide differences in streamflow characteristics exist between adjacent sub-basins as well as within the same sub-basin but at different gaging stations. Such differences are caused by the presence or absence of buried glacial outwash deposits underlying the stream bed that contribute large quantities of groundwater to streamflow.

In the upper part of the basin, the Scioto River drains an area covered by a relatively thick layer of glacial drift that is underlain by limestone to the west and shale to the east (fig. 31). The stream channel is cut into the till only and receives most of its ground-water from ground moraine. Likewise, the hydrograph of the Little Scioto
### Table 2. Summary of Streamflow and Flow Duration Data for Representative Streams in the Scioto River Basin Based on 1967 Streamflow Data.

<table>
<thead>
<tr>
<th>Gaging Station Location</th>
<th>Drainage Area in sq. mi.</th>
<th>High Flow (Q10)</th>
<th>Mean Flow (Q25)</th>
<th>Median Flow (Q50)</th>
<th>Average Flow (Q75)</th>
<th>Low Flow (Q90)</th>
<th>(Q10/Q90) x 100</th>
<th>(Q25/Q75) x 100</th>
<th>Recharge Rate (gpd/sq. mi.)</th>
<th>% of Flow as Ground Water</th>
</tr>
</thead>
<tbody>
<tr>
<td>Little Scioto above Marion</td>
<td>72.4</td>
<td>2.804</td>
<td>0.981</td>
<td>0.262</td>
<td>0.026</td>
<td>0.009</td>
<td>17.65</td>
<td>6.14</td>
<td>222,000</td>
<td>38.7</td>
</tr>
<tr>
<td>Scioto River at Columbus</td>
<td>1629</td>
<td>3.250</td>
<td>1.007</td>
<td>0.374</td>
<td>0.374</td>
<td>0.104</td>
<td>5.62</td>
<td>2.77</td>
<td>237,000</td>
<td>35.4</td>
</tr>
<tr>
<td>Scioto River at Chillicothe</td>
<td>3849</td>
<td>2.910</td>
<td>1.107</td>
<td>0.405</td>
<td>0.102</td>
<td>0.117</td>
<td>4.99</td>
<td>2.75</td>
<td>267,000</td>
<td>43.2</td>
</tr>
<tr>
<td>Olentangy River at Clarksdale</td>
<td>157</td>
<td>3.427</td>
<td>0.975</td>
<td>0.255</td>
<td>0.040</td>
<td>0.022</td>
<td>12.48</td>
<td>4.94</td>
<td>243,000</td>
<td>32.7</td>
</tr>
<tr>
<td>Olentangy River nr. Delaware</td>
<td>393</td>
<td>3.104</td>
<td>0.885</td>
<td>0.226</td>
<td>0.071</td>
<td>0.028</td>
<td>10.53</td>
<td>3.53</td>
<td>123,000</td>
<td>18.9</td>
</tr>
<tr>
<td>Whetstone Cr. nr. Ashley</td>
<td>98.7</td>
<td>2.796</td>
<td>0.922</td>
<td>0.284</td>
<td>0.075</td>
<td>0.044</td>
<td>7.97</td>
<td>3.51</td>
<td>281,000</td>
<td>41.0</td>
</tr>
<tr>
<td>Big Walnut Cr. @ Central College</td>
<td>180</td>
<td>2.400</td>
<td>0.947</td>
<td>0.421</td>
<td>0.363</td>
<td>0.342</td>
<td>2.65</td>
<td>1.62</td>
<td>351,000</td>
<td>49.7</td>
</tr>
<tr>
<td>Big Walnut Cr. @ Rees</td>
<td>544</td>
<td>2.776</td>
<td>0.820</td>
<td>0.287</td>
<td>0.140</td>
<td>0.074</td>
<td>6.12</td>
<td>2.42</td>
<td>332,000</td>
<td>36.1</td>
</tr>
<tr>
<td>Alum Cr. @ Africa</td>
<td>122</td>
<td>3.697</td>
<td>0.918</td>
<td>0.246</td>
<td>0.066</td>
<td>0.017</td>
<td>14.75</td>
<td>3.73</td>
<td>270,000</td>
<td>35.8</td>
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<tr>
<td>Alum Cr. @ Columbus</td>
<td>189</td>
<td>2.603</td>
<td>0.915</td>
<td>0.291</td>
<td>0.132</td>
<td>0.058</td>
<td>6.70</td>
<td>2.63</td>
<td>279,000</td>
<td>39.8</td>
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<tr>
<td>Deer Cr. @ Mt. Sterling</td>
<td>228</td>
<td>2.395</td>
<td>1.004</td>
<td>0.306</td>
<td>0.118</td>
<td>0.061</td>
<td>6.27</td>
<td>2.92</td>
<td>289,000</td>
<td>47.3</td>
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<td>Deer Cr. @ Williamsport</td>
<td>333</td>
<td>2.646</td>
<td>0.925</td>
<td>0.266</td>
<td>0.109</td>
<td>0.045</td>
<td>7.67</td>
<td>3.21</td>
<td>200,000</td>
<td>45.6</td>
</tr>
<tr>
<td>Paint Creek nr. Greenfield</td>
<td>249</td>
<td>2.019</td>
<td>0.960</td>
<td>0.309</td>
<td>0.048</td>
<td>0.012</td>
<td>11.59</td>
<td>4.47</td>
<td>294,000</td>
<td>48.8</td>
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<tr>
<td>Rocky Fork nr. Berretts Mills</td>
<td>140</td>
<td>2.229</td>
<td>1.171</td>
<td>0.329</td>
<td>0.061</td>
<td>0.032</td>
<td>8.34</td>
<td>4.38</td>
<td>309,000</td>
<td>59.8</td>
</tr>
<tr>
<td>Tar Hollow Cr. @ Tar Hollow St. Park</td>
<td>1.4</td>
<td>3.185</td>
<td>1.037</td>
<td>0.123</td>
<td>0.000</td>
<td>0.000</td>
<td>-</td>
<td>-</td>
<td>311,000</td>
<td>45.8</td>
</tr>
</tbody>
</table>

(1) 10% Flow is high flow. Flow that is equalled or exceeded 10 percent of the time.
(2) Q 25% Meanflow. An index of basin storage (Cross and Hedges, 1959, p. 9).
(3) 50% flow is Median flow, flow that is equalled or exceed 50 percent of the time.
(4) 75% flow is Average flow, flow that is equalled or exceeded 75 percent of the time.
(5) 90% flow, flow that is equalled or exceeded 90 percent of the time. An arbitrary index of the dry weather flow (base flow) and indicates the relative amount of ground water storage that is effective in maintaining dry weather flow (Cross and Hedges, 1959, p. 5).
(6) $Q_{10}/Q_{90}$ Both ratios describe the slope of the flow duration curve and hence the character of the stream flow.
(7) $Q_{25}/Q_{75}$ If the ratio approaches unity the stream is then characterized by a uniform flow and has a steady supply. If the ratio is greater than unity then the flow is much more variable.
River above Marion displays relatively high flows following heavy rainfall, but the peak flow decreases rapidly because of the lack of basin storage (fig. 13). The flow-duration curve of the Little Scioto has a steep slope and small rates of low flow (.009 cfs/mi² at 90 percent) (fig. 14). This is a reflection of the low permeability of the basin material. The ratio of the high flow to low flow \( \left( Q_{10}/Q_{90} \right)^{1/2} \) supports this conclusion, because the Little Scioto has the highest ratio in the basin (Table 2). The low flow index \( Q_{90\%} \) indicates a very low base flow and relatively small amounts of groundwater runoff.

Between Delaware and Columbus, the Scioto River flows through a gorge cut into limestone and gains some groundwater runoff from the limestone. Below Columbus the stream channel is cut into an extensive wide, and thick deposit of outwash that is relatively permeable and, in this stretch, the contribution of groundwater runoff to the stream flow increases markedly. The stream hydrograph and flow-duration curve for the gage at Chillicothe emphasize these changes in lithology in that the base flow of the river is very high south of Columbus, and therefore the storage capacity of the underlying glacial outwash aquifer (figs. 15 and 16). The flow-duration curve flattens towards the lower limb (around the 80 per cent value) indicating a very high base flow. The Scioto River continues to flow over the extensive outwash aquifer to the Ohio River, and the stream discharge reflects a much higher flow rate and sustained flow than it does upbasin.
Figure 13. Hydrograph of Little Scioto River above Marion, Ohio.
Figure 14. Flow-duration curve of Little Scioto River above Marion, Ohio.
231500 SCIOTO RIVER AT CHILLICOTHE, OHIO 1967

11 DAY LOCAL MINIMA

3849.0 SQ.MI.

Figure 15. Hydrograph of the Scioto River at Chillicothe, Ohio.
Figure 16. Flow-duration curve of the Scioto River at Chillicothe, Ohio.
The Olentangy River flows along the eastern margin of the Broadway Moraine. Its channel is underlain by either shale or limestone. Contribution from the thick dense layer of till is relatively small due to its low permeability. Outwash is absent in this area with the exception of a few kames and eskers in the upper reaches along the eastern loop of the Broadway Moraine. Consequently the Olentangy River is characterized by a low base flow. This is also reflected by the high index of the high flow to low flow ratio $(Q_{10}/Q_{90})^{1/2}$. Inspection of the hydrographs and the flow-duration curves for two different locations, one near the headwaters of the stream at Claridon, and the other near the lower end in the vicinity of Delaware shows that there is no significant increase in the nature of flow, and the 90 percent flow is almost equal at both locations (figs. 17 to 20). This implies that there is but little ground-water runoff from the till.

Wide differences in stream flow characteristics can exist not only in adjacent areas but even within the same sub-basin. In the northern half of its drainage area, Alum Creek has the same low-natural flow characteristics as the Olentangy. Alum Creek flows several miles over an area of predominately thin layer of till underlain by dense shale and its sustained flow is low as reflected by the hydrograph and flow-duration curve at Africa gage (figs. 21 and 22). The base flow of Alum Creek at the downstream station at Columbus shows a three fold increase over that at Africa (figs. 23 and 24). This substantial increase is due to ground-water runoff from thick and
223000 OLENTANGY RIVER AT CLARIDON, OHIO 1967

5 DAY LOCAL MINIMA

TOTAL DISCHARGE 5.534E9 CF OR 15.17 INCHES
GROUND WATER RUNOFF 1.066E9 CF OR 5.12 INCHES
GROUND WATER AS % 33.7
RECHARGE RATE 243000 GPD /SQ. MI.

Figure 17. Hydrograph of the Olentangy River at Claridon, Ohio.
PERCENT OF TIME DISCHARGE IS EQUALLED OR EXCEEDED

Figure 18. Flow-duration curve of the Olentangy River at Claridon, Ohio.
225500 OLENTANGY RIVER NEAR DELAWARE, OHIO 1967

TOTAL DISCHARGE   1.254E 10  CF OR 13.74  INCHES
GROUND WATER RUNOFF 2.366E 9  CF OR 2.59  INCHES
GROUND WATER AS %   18.9
RECHARGE RATE    123000  GPD /SQ. MI.

Figure 19. Hydrograph of the Olentangy River near Delaware, Ohio.
Figure 20. Flow-duration curve of the Olentangy River near Delaware, Ohio.
Figure 21. Hydrograph of Alum Creek at Africa, Ohio.
Figure 22. Flow-duration curve of Alum Creek at Africa, Ohio.
Figure 23. Hydrograph of Alum Creek at Columbus, Ohio.
Figure 24. Flow-duration curve of Alum Creek at Columbus, Ohio.
permeable glacial deposits in the lower reach of the stream. Similarly, Big Walnut Creek at Columbus is affected by the buried glacial outwash that increase the base flow and storage capacity of the stream channel.

Deer and Darby Creeks exhibit a striking similarity in flow characteristics. They both drain areas of similar surface features, as they flow through areas covered with relatively thick glacial drift that is underlain by limestone. Poorly sorted glacial outwash, covered by a few feet of alluvium, is present along the lower reaches of both streams. Both hydrographs and flow-duration curves express these similarities (figs. 25 to 28). The ratio of high to low flow \((Q_{10}/Q_{90})^{1/2}\) of both streams are nearly alike and the shape of the flow duration curves flatten up towards the lower limbs indicating significant base flow.

In the unglaciated portion of the basin, stream hydrographs indicate rapid runoff following precipitation, and low sustained flows. This is due to the rugged terrain, including steep valley sides that induce rapid surface runoff. The flow also reflects the low permeability of the bedrock in this unglaciated area. Tar Hollow Creek flows across an area underlain by thin-bedded Mississippian sandstone. Its hydrograph (fig. 29) expresses the flashy nature of the stream. Because the flow dissipates quickly after a storm, the flow-duration curve is very steep. Furthermore, Tar Hollow Creek is commonly dry during the late summer (fig. 30).

From analyses of geologic and stream flow-characteristics in the basin it is evident that the streams fall under one of three
Figure 25. Hydrograph of Big Darby Creek at Darbyville, Ohio.
Figure 26. Flow-duration curve of Big Darby Creek at Darbyville, Ohio.
231000 DEER CREEK AT WILLIAMSPORT, OHIO 1967

TOTAL DISCHARGE 1.035E 10 CF OR 13.38 INCHES
GROUND WATER RUNOFF 4.718E 9 CF OR 6.10 INCHES
GROUND WATER AS % 45.6
RECHARGE RATE 290000 GPD /SQ. MI.

Figure 27. Hydrograph of Deer Creek at Williamsport, Ohio.
Figure 28. Flow-duration curve of Deer Creek at Williamsport, Ohio.
235500 TAR HOLLOW CREEK AT TAR HOLLOW STATE PARK, OHIO 1967

3 DAY LOCAL MINIMA

TOTAL DISCHARGE 4.477E7 CF OR 14.27 INCHES
GROUND WATER RUNOFF 2.051E7 CF OR 6.54 INCHES
GROUND WATER AS % 45.8
RECHARGE RATE 311000 GPD /SQ. MI.

Figure 29. Hydrograph of Tar Hollow Creek at Tar Hollow State Park, Ohio.
Figure 30. Flow-duration curve of Tar Hollow Creek at Tar Hollow State Park, Ohio.
categories. Those that flow in unglaciated terrains are generally characterized by a very high ratio of high to low flow index \((Q_{10}/Q_{90})^{1/2}\) and very low base flow. The bedrock has a very low storage capacity and permeability. The second category of streams includes those streams that flow over the relatively low permeability glacial drift and lacks the permeable outwash material. These streams differ from those in the first category in that they have higher sustained flows and lower ratios of high to low flow. The till stores large amounts of water but releases it very slowly, resulting in low-flow indices that are relatively higher than those in the unglaciated category. The third category includes those streams that flow on a till-covered terrain associated with surficial or shallow buried outwash deposits that are very permeable and have appreciable storage capacities. They are characterized by very high low-flow indices and low ratios of high to low flow indices.

Streams that fall into the first category include Salt Creek, Sunfish Creek, and Scioto Brush Creek; those in the second are Scioto River above Marion, Olentangy River, and Alum Creek above Africa; those in the third category includes Scioto River below Columbus, Big Walnut Creek, Alum Creek, and Paint Creek.
IV. SOURCES OF GROUND WATER IN THE SCIOTO RIVER BASIN

Ground Water in Consolidated Aquifers

Ground-water in the Scioto River Basin occurs in two principal types of aquifers; consolidated or bedrock aquifers and unconsolidated aquifers such as till, alluvium and outwash. Consolidated aquifers occur in the Silurian-Devonian carbonate rocks that underlie most of the western half of the basin, and occur along the eastern edge of the basin in the Mississippian sandstones (fig. 31). Silurian-Devonian carbonates are covered by glacial deposits in the western half of the basin and, to the east, by a succession of younger Paleozoic formations.

The principal carbonate aquifer in the Scioto River Basin is the Bass Islands Group of the Late Silurian age. The Lockport Dolomite of Middle Silurian age becomes an important component of the aquifer where it crops out beneath the glacial drift. Towards the central part of the basin the Columbus and Delaware Limestones of Devonian age complete this extensive carbonate aquifer. The aquifer, in areas where it is the main source of ground water, ranges from 150 to 450 feet in thickness and the Bass Islands Group is the thickest and most productive.

At or near the crest of the Cincinnati Arch, which lies west of the basin boundary, the beds are flat-lying. To the east of the crustal area the beds dip gently eastward. When the Cincinnati Arch was uplifted in Late Paleozoic-Early Mesozoic time, erosion removed the
Figure 31. Generalized bedrock geology of the Scioto River basin.
younger formations from the crest of the arch and exposed the older strata, hence the Columbus and Delaware Limestones along with the Bass Islands Group are absent towards the crest and the Lockport Dolomite is either exposed or covered by till in the southwestern part of the basin.

Fracturing and jointing of the rocks took place, perhaps consequent to uplift, thus developing secondary permeability. Increased permeability resulted from solution by percolating ground-water, which dissolved the carbonate rocks, enlarged the joints and fractures, and created local solution channels. Generally, rocks of the Bass Islands Group are more permeable than those of the Lockport Dolomite, especially towards the base of the section where a widespread zone of high permeability caused by solution openings occurs above the top of the Lockport (figs. 11 and 12). This zone is called the "Newburg Zone" (Norris, 1957). Yields from the Newburg Zone far exceed those commonly obtained from wells drilled into limestone. Along the eastern margin of the aquifer the Newburg Zone occurs at greater depths than those commonly reached by drillers and, has not yet been explored.

The carbonate aquifer is one of the most important sources of ground-water in central and northwestern Ohio. It provides approximately 35 mgd (million gallons per day) to municipal and industrial users in northwestern Ohio alone. It is the most uniformly productive of all the consolidated aquifers in the basin (fig. 32). A more detailed discussion of the aquifer characteristics and its potential is given in Chapter VI.
Figure 32. Ground water yields from individual wells drilled in the Scioto River Basin (O.D.N.R. Div. of Water).
The Berea Sandstone and Cuyahoga Formation, both of Mississippian age, are subordinate aquifers in the basin. They are not well developed and their exploitation is largely limited to domestic and farm use.

The Berea subcrops beneath the glacial till in a belt approximately two miles wide in an area adjacent to the eastern limits of the Devonian rocks (fig. 31). This belt extends northward from the vicinity of Chillicothe to the northern boundary of the basin in Crawford County. The Berea consists mainly of sandstone that alternates with thin beds of shale. The sandy layers form aquifers that range from a few inches to several feet thick and are composed of well cemented fine-grained quartz sandstone. The total thickness of the Berea ranges from 55 feet in the north to about 5 feet in the south.

The Black Hand Member of the Cuyahoga Formation consists of coarse-grained massive channel-like conglomeratic sandstone deposits. The Black Hand subcrops along a narrow belt in eastern Delaware, Franklin, Fairfield and Pickaway Counties.

Most of the water in the Berea and Black Hand occurs along joints and bedding planes. Yields from wells drilled into these aquifers are highly variable but the maximum rate is generally less than 15 gpm. Wells that tap the Berea in Franklin County yield 9 to 70 gpm (Schmidt, 1958) and wells in the Black Hand Sandstone in Fairfield County yield 10 to 125 gpm (Wolfe and others, 1962). The best quality water is obtained from the shallower depths and the quality deteriorates downward, increasing in both iron and dissolved solids.
Ground Water in the Unconsolidated Aquifers

Till is the most widespread of the glacial deposits in the basin. It is far less permeable than outwash deposits and is not considered to be an important source. Very low yields are obtained from dug wells in rural and farming communities in the basin. Till generally impedes infiltration to the water table and underlying aquifers and, hence, acts as an aquiclude. It is of great hydrologic significance, however, because of its considerable storage potential.

Isolated stringers of sand and gravel of variable thickness are distributed randomly with the till throughout most of the glaciated area. They are relatively permeable and yield appreciable amounts of water for domestic and farm use.

By far the most important ground-water sources in the unconsolidated deposits are surficial outwash and the buried valley aquifers that underlie or are adjacent to drainage channels of present streams. Glacial outwash in the old valleys of the pre-glacial Teays and Deep Stage drainage systems forms a high-yielding aquifer. Pre-Illinoian glaciation partially filled these valleys with well-sorted layers of sand and gravel, interbedded in some cases with silt. The Wisconsin glaciation completed the deposition of these materials, but one cannot be sure to which any given buried valley belongs. In general, the upper portion of the valley fill consists of undifferentiated till and alluvium that cover the more porous and permeable sand and gravel deposits. Recharge through the till is relatively small except where the sand and gravel are exposed at the surface, as in the case of the Scioto River south of
Columbus near Shadeville. Recharge to the buried-valley aquifer occurs in these places when the water table is lower than the stream level. Leakage from the aquifer to the stream occurs when the water table is higher than the stream level.

Beyond the drift border, south of Chillicothe, well sorted sand and gravel fill the Scioto River channel and, in places, reach a thickness of as much as 300 feet. The distribution of the coarse valley-fill varies greatly, both horizontally and vertically (fig. 8). Some of the thickest and most permeable valley-fill deposits occur in the Scioto River valley in the vicinity of Piketon. Here, the outwash deposits can be recharged by induced infiltration from the river and wells are capable of yielding more than 1000 gpm (Norris, 1963). The ERDA Uranium enrichment plant, for example, pumps 13 mgd from 15 wells that tap the valley-fill deposits along the river channel just west of the city (fig. 41).

The buried-valley aquifer is the most high yielding and dependable source of ground water of good quality in the basin. The attention of many industries and municipalities in the region has been focused on this aquifer as a future source to satisfy their growing needs for water. For this reason, a more detailed discussion of the aquifer characteristics and its potential as a major source of ground-water supply is presented in Chapter V. Also, a digital model of a section of the aquifer at Piketon is described in order to demonstrate the relevancy of computer simulation techniques to the solution of problems of water supplies for the various municipalities and industries in the area.
V. HYDROGEOLOGY OF THE BURIED VALLEY AQUIFER

The buried-valley aquifer in the Scioto River Basin lies adjacent to the Scioto River and extends from the proximity of southern Columbus in Franklin County, through Pickaway, Ross, Pike and Scioto Counties (fig. 7). The length of the aquifer, between Columbus in the north and Portsmouth in the south, is about 90 miles. The valley is widest in the north near Ashville, where it is approximately eight miles wide and contains an average thickness of 250 feet of sand and gravel. The narrowest part is near Wakefield, about eight miles south of Piketon, where the valley is about 1.25 miles wide and contains about 73 feet of sand and gravel. The aquifer underlies approximately 440 square miles.

The aquifer supplies the largest ground-water yield and forms the most important aquifer in the basin. It is a major source of ground water in Ohio and has considerable potential for additional development in the future. It is the main source of water for many municipalities and industries. The estimated average daily demands of ground water from this aquifer for public use in 1974 was about 17 mgd and industrial use averaged about 45 mgd. Increasing demands for good-quality water resulting from industrialization of the area and rising cost of treatment of surface water certainly make additional use of water from this aquifer very attractive.

Present development of the aquifer is centered in areas of
large population and industry. The local plastic and glass industries near Circleville pump about 8.5 mgd from this aquifer, and paper companies near Chillicothe pump about 12 mgd. The Energy and Resources Development Administration (ERDA) Uranium Enrichment Plant at Piketon requires more than 13 mgd of ground water, which is pumped from 15 wells along the Scioto River just south of Piketon.

Walker et al. (1965) estimated that the potential supply in the aquifer could be as much as one and a half billion gallons per day in the area adjacent to the Scioto River. Bloyd (1974) estimated the water available from storage in the outwash and alluvial aquifer to be 748 billion gallons. The increase in public and industrial demand for water and the increase in agriculture combined with the rise in the cost of surface-water treatment will undoubtedly result in more extensive development of the aquifer as a major source of groundwater.

The average public demand for water in Columbus was approximately 102 mgd in 1974. Future needs are estimated to be 122.9 mgd in 1980 and 241.7 mgd in 2020 (O.D.N.R. Table 39, in preparation). These projected increases in water demand have forced the city of Columbus to evaluate additional sources. Among these is the development of a well field with a capacity of 50 mgd in the Scioto outwash in the southern part of Franklin County.

The ERDA plant at Piketon has recently increased well production from 3 to 13 mgd. The plant is undergoing expansion and, thus, water requirements are expected to increase even more. Other
potential expansions are foreseen for industrial plants in the Circleville and Chillicothe areas. The population of many cities and villages is increasing and new communities continue to appear (Table 3). Consequently, public water-supply systems are being expanded and new ones are frequently required, thus increasing the reliance on the buried valley aquifer as the source of water supply.

The areas of highest yield of the aquifer lie in a strip along the buried valley and immediately adjacent to both sides of the Scioto River. Wells in this zone generally produce 500 to 1000 gpm (fig. 32). Yields of wells in the aquifer generally increase with an increase in the saturated thickness of the aquifer. Wells drilled in the northern two-thirds of the aquifer where the aquifer thickness averages about 100 feet, generally yield more than 1000 gpm. The reason for the high yield is the closeness of the wells to the river, which allows induced infiltration. Farther from the river, well yields decrease and drawdowns are somewhat greater due to the decreased effect of infiltration and the larger spread of the cone of depression.

Statistical summaries of the total populations being served by public water systems in the basin (O.D.N.R. Table 39, in preparation) show that most of these systems obtain their supply from the aquifers in the area. As shown in their projections, the number of persons served by public systems will increase from approximately 0.8 million in 1974 to 1.6 million in 2020. This increase in population alone will undoubtedly raise demands on the buried-valley aquifer as a
TABLE 3

Public Water Demand Projections Obtained from the Buried Valley Aquifer in million gallons per day

<table>
<thead>
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<th></th>
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<th></th>
<th></th>
<th></th>
<th></th>
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<td>Ashville</td>
<td>0.17(^1)</td>
<td>0.28</td>
<td>0.34</td>
<td>0.57</td>
<td>0.77</td>
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<td>Chillicothe</td>
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<td>4.77</td>
<td>6.01</td>
<td>7.78</td>
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<td>8.0</td>
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<td>91.83</td>
<td>114.0</td>
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<td>0.66</td>
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<td>Groveport</td>
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<tr>
<td>Waverly</td>
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<td>0.94</td>
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<tr>
<td>Lake White</td>
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<td>0.03</td>
<td>0.04</td>
<td>0.05</td>
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</tr>
<tr>
<td>Piketon</td>
<td>0.16</td>
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<td>0.29</td>
<td>0.31</td>
</tr>
<tr>
<td>Lucasville</td>
<td>-</td>
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<td>1.46</td>
<td>2.28</td>
<td>2.80</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>14.89</strong></td>
<td><strong>16.71</strong></td>
<td><strong>68.88</strong></td>
<td><strong>107.45</strong></td>
<td><strong>134.46</strong></td>
</tr>
</tbody>
</table>

\(^1\) Data are partly from O.D.N.R. Table 39 (in preparation).
\(^2\) Data are from O.D.N.R. Table 19 (1963).
source of water. Daily demand for public water supply in the year 2020 may increase about eight times that of 1974 (Table 3). The need for more ground water will become imperative with the increase of industry in the area.

Table 4 projects the use of industrial water to the year 2020. The future withdrawal of ground water for industrial use will be greatly influenced by the cost of obtaining surface water and by pollution abatement requirements in order to meet surface water quality standards. Under these growing restrictions, ground water becomes more favored than surface water, since, in most instances, ground water is much cheaper and less polluted than surface water, and thus it is anticipated that withdrawals from the aquifer will increase tremendously.

At present, industries in the area are concentrated in three places. In the Circleville area are plastics, glass, and chemical plants that now require a total of approximately 8.3 mgd. The paper plants in Chillicothe use large quantities of ground water; their average daily demand is about 28.6 mgd. The ERDA Plant at Piketon requires a daily average of about 13 mgd which are withdrawn from 15 wells. Further expansion of the plant will increase the demand to about 25 mgd. Other nearby industries include sand and gravel operations, food and dairy processing, tool and machinery making, and transportation.
TABLE 4. Projections of the demand of Industrial Water in mgd

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<td>46.82</td>
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<td>8.9</td>
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<td>Pike (excluding 25 mgd forecasted for ERDA)</td>
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<td>7.16</td>
<td>7.39</td>
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<tr>
<td>Pike (total)</td>
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<td>32.16</td>
<td>32.39</td>
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<td>Ross</td>
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<td>44.49</td>
<td>45.61</td>
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<td>TOTAL</td>
<td>45.55</td>
<td>140.19</td>
<td>170.42</td>
<td>186.40</td>
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</table>
Character of Material in the Buried-Valley Aquifer

The wide valley that was cut into the bedrock by the Teays and Deep Stage streams was later filled with sand, gravel, and minor clay. These sediments were transported downstream by the melt-waters issuing from the ice front and were deposited as a valley train. Clay occurs as lenses and layers interbedded with the sand and gravel. Lenses of till also occur in the valley fill north of the glacial limits.

Logs of wells in the aquifer show that the sand, gravel, and clay deposits vary greatly, both vertically and horizontally. This is attributed to the variable conditions that existed during the periods of deposition. The layers are generally discontinuous and difficult to trace from one well to another. In some places, thick deposits of sand and gravel mixed with clay indicate that the material was deposited with little sorting. Elsewhere coarse gravel, containing little sand and clay is present.

Part of the outwash aquifer north of the glacial boundary is covered by till that ranges in thickness from 5 to 150 feet. The till generally consists of an unsorted mixture of clay, silt, sand, and gravel. Because glacial till retards the infiltration of water, it controls, to a certain degree, the availability of water in the aquifer. The permeability of till ranges between 0.0002 and 0.9 gpd/ft$^2$ (Norris, 1962). The average permeability as reported by tests made in Illinois, South Dakota, and Ohio was estimated to be 0.175 gpd/ft$^2$ (Norris, 1962).
End moraines contribute more recharge to the aquifer than till. Sand and gravel lenses are generally associated with end moraines and these contain appreciable amounts of water that recharge the aquifer.

Beyond the glacial limits south of Chillicothe, the aquifer is overlain by 5 to 10 feet of soil and Recent alluvium. The material that makes up the aquifer consists in large part of coarse sand and medium gravel. Finer-grained material, consisting of fine sand and silt, are scattered randomly throughout the aquifer. The greatest amount of recharge occurs in the unglaciated part where coarse material extends to the surface.

Bedrock underlying the aquifer varies from carbonates of Devonian age (Columbus and Delaware Limestones) to shales of Devonian age (Ohio and Olentangy Shales). The carbonates underlie the aquifer from the vicinity of Columbus southward to the confluence of the Scioto River and Little Darby Creek, and are part of the widespread Silurian-Devonian aquifer. Evidence of leakage from the carbonate aquifer to the buried valley aquifer is supported by the potentiometric surface of the carbonate aquifer (fig. 65), which lies at higher elevations than the general water table levels in the buried valley aquifer (Plate 1). Because flow occurs from high to lower head, water flows from the carbonate aquifer to the buried valley aquifer.

South of Ashville, the aquifer is underlain by shales of Devonian and Mississippian age. They are generally impermeable and leakage from them is negligible.
Geomorphic Development of the Aquifer

The Scioto River buried-valley aquifer lies in the drainage channel of the preglacial Teays River; the oldest preserved drainage system in the basin. The Teays River originated in the Piedmont Plateau of North Carolina and Virginia, flowed north westward to the Ohio Valley at Huntington, West Virginia, where it occupied what is now part of the present Ohio River Valley between Huntington and Wheelersburg. From Wheelersburg the Teays River turned northward toward Waverly and flowed in a broad valley, now abandoned, that is roughly parallel to the course of the present Scioto River and then on to the vicinity of Chillicothe where the valley passes under the mantle of the glacial drift (fig. 33). The course of the Teays River north of the glacial boundary has been established by well logs, test holes, and geophysical methods. From Chillicothe the river flowed northwestward through Pickaway, Madison, Clark, and Champaign counties to the Indiana border in Mercer County. The river then continued westward through Indiana and Illinois to join the ancestral Mississippi River. Recent studies (Norris and Spicer, 1958; Schmidt, 1958; and Walker et al., 1965) have locally modified the position of the Teays River as proposed by Stout et al. (1943).

The average gradient of the Teays River (Stout, Ver Steeg, and Lamb, 1943) is about 10.76 inches per mile. The width of the valley in southern Ohio varies from 1.25 to 8 miles (Walker et al., 1965), but it averages about 1.45 miles, especially in areas least
Figure 33. Teays Stage drainage in the Scioto River basin (after Stout et al., 1943).
Figure 34. Deep Stage drainage in the Scioto River Basin (after Stout et al., 1943).
disturbed by later erosion. The width of the valley, however, increases uniformly downstream due to the addition of major tributaries. The depth of the valley ranged from 25 to 250 feet and cross sections in the Scioto valley indicate that the valley had deep narrow channels and, locally, bedrock highs (fig. 8).

The low gradient of the Teays River and the uniform broad valley in which it flowed, as well as the well-rounded hills along its course, indicate that the river had reached erosional maturity before the onset of continental glaciation.

The Teays drainage system was terminated when the Pre-Illinoian (Kansan or Pre-Kansan) ice sheet advanced southward into Central Ohio. The ice dammed the Teays valley and its tributaries, creating long finger lakes in which clay and silt deposits, given the name "Minford Silt" by Stout and Schaaf (1931), accumulated and filled the valleys. The continued damming caused the waters in the lakes to break over low divides to establish a new drainage system (fig. 34), known as the Deep Stage (Stout, et al., 1943). The Newark River, a major tributary to the master Cincinnati River, drained a large part of central and south-central Ohio. It rose in northeastern Ohio, flowed southward and entered the Scioto basin in eastern Fairfield County. From Fairfield County, the Newark River continued into Franklin County then turned southward, entering Pickaway County. From Pickaway County southward, the river cut its channel along the course of the present Scioto River. The gradient of the Newark River as estimated by Stout et al. (1943) was about 23 feet per mile between Franklin County and the confluence of the
river with the Cincinnati River.

Walker et al. (1965) indicate that the Deep Stage drainage is characterized by much deeper erosion than was the earlier Teays Stage. The valley floor of the Deep Stage is 90 feet below that of the Teays stage at Circleville; 185 feet at Piketon and more than 195 feet at Portsmouth. The deep, wide channel was later filled with sand, gravel, silt, and undifferentiated glacial drift during the advance of the Illinoian and Wisconsin glacial stages. The outwash material was deposited as sorted sand and gravel in the now buried channels of the Deep Stage.

The Deep Stage drainage period was terminated when the channels of the rivers were partly filled with sand and gravel. The channels were filled to elevations of 560-580 feet at Piketon (Norris and Fidler, 1969). The drainage channel of the present Scioto River, which has its course in the same buried channel, and which came into existence during the last retreat of the glacier in Wisconsinan times, has cut 20-30 feet of the outwash material deposited previously.

**Hydrologic Properties of the Aquifer**

The hydrologic properties of aquifers can best be determined by conducting aquifer tests in the areas where development is planned. Pump tests of certain amounts of discharge and durations are especially valuable in determining the yielding properties of aquifers. Many of these aquifer tests have been conducted along the buried-valley aquifer (fig. 35) and the results are summarized
Figure 35. Locations of aquifer tests in the Buried-valley aquifer.
in Table 5. The coefficient of permeability (P) of the aquifer is a unit measure of the ability of the aquifer to transmit water. It is defined as the rate of flow of water in gallons per day through a cross-sectional area of one square foot of the aquifer under a hydraulic gradient of 1 foot per foot. It depends primarily on the nature of the pore spaces in the aquifer, the degree of sorting of the material, and the degree of roundness and size of the individual grains of the material composing the aquifer.

Permeability ranges from 456 gpd/ft\(^2\) to 4,600 gpd/ft\(^2\) and averages 3,250 gpd/ft\(^2\), as seen from Table 5. The variations in the results are due mainly to the distance of the pumped well from the Scioto River, which in turn affects the infiltration rate from the stream. Wells that are drilled close to the stream have higher transmissivity and permeability values than those that are away from the stream. The test at Circleville was about 1.5 miles from the river, hence the distance to the source of recharge is greater and the transmissivity and permeability values are lower. Variations are also due to the variability in the composition of the aquifer material and the presence or absence of clay lenses that are scattered randomly through the aquifer. The results further show that the aquifer can yield large quantities of water and that these quantities are sustained by the induced infiltration from the streamflow.

The coefficient of transmissivity (T) is defined as the rate of flow in gallons per day through a vertical section of the aquifer one foot wide that extend through the entire saturated thickness of
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<th>Test # on Fug.</th>
<th>Location</th>
<th>Date</th>
<th>Duration of Test, hr.</th>
<th>Pumping Rate (gpm)</th>
<th>Aquifer Transmissivity (gpd/ft)</th>
<th>Permeability ( P ) (gpd/ft(^2))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1(^a)</td>
<td>Franklin</td>
<td>Hamilton</td>
<td>1975</td>
<td>72</td>
<td>946</td>
<td>70</td>
</tr>
<tr>
<td>2(^b)</td>
<td>Pickaway</td>
<td>Circleville</td>
<td>1973</td>
<td>-</td>
<td>2200</td>
<td>100</td>
</tr>
<tr>
<td>3(^c)</td>
<td>Ross</td>
<td>Union</td>
<td>1964</td>
<td>96</td>
<td>115</td>
<td>68</td>
</tr>
<tr>
<td>4(^c)</td>
<td>Ross</td>
<td>Scioto</td>
<td>1953</td>
<td>88</td>
<td>415</td>
<td>70</td>
</tr>
<tr>
<td>5(^c)</td>
<td>Ross</td>
<td>Scioto</td>
<td>1958</td>
<td>74</td>
<td>800</td>
<td>68</td>
</tr>
<tr>
<td>6(^c)</td>
<td>Ross</td>
<td>Scioto</td>
<td>1958</td>
<td>72</td>
<td>800</td>
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</tr>
<tr>
<td>7(^c)</td>
<td>Ross</td>
<td>Scioto</td>
<td>1965</td>
<td>50</td>
<td>505</td>
<td>49</td>
</tr>
<tr>
<td>8(^c)</td>
<td>Pike</td>
<td>Seal</td>
<td>1964</td>
<td>216</td>
<td>1000</td>
<td>63</td>
</tr>
<tr>
<td>9(^c)</td>
<td>Pike</td>
<td>Seal</td>
<td>1952</td>
<td>60</td>
<td>795</td>
<td>41</td>
</tr>
</tbody>
</table>

\( a \) After Stilson, 1976

\( b \) After Norris, 1975

\( c \) After Walker and Others, 1965
the aquifer. The transmissivity represents the water transmitting capability of the aquifer. Since T is a product of permeability and thickness, the factors that affect permeability are also applicable to transmissivity.

The transmissivity of the buried valley, as determined from aquifer tests, ranges from 46,000 to 322,000 gpd/ft, and averages about 215,000 gpd/ft.

The infiltration rate through the stream bed is defined as the rate of flow, in gallons per day per square foot of stream bed per foot of differential head between the stream and the water table, when the water level is lowered below the stream level. The infiltration rate depends primarily on the permeability of the stream bed, the difference in head between the stream stage and the ground-water level, and the temperature of water in the stream. Norris and Fidler (1969) evaluated the infiltration rate through the bed of the Scioto River by means of an aquifer test conducted at Piketon. They considered the results (6.2 gpd/ft²/ft) a minimum rate. Tests conducted in the Columbus South Well field (Klaer and Assoc., 1976) indicated infiltration rates that ranged from 2.41 to 3.50 gpd/ft²/ft in similar aquifer material. The variations can be attributed to a lowered permeability of the stream bed at Piketon. Here a layer of silt, mud, and organic debris, penetrating only a few inches into the sediments, tends to decrease the stream bed permeability (Norris and Fidler, 1969). An increased ground-water gradient, perhaps due to nearby pumping wells, would tend to increase the rate. It is estimated that about 41.6 mgd of water can be
intercepted by infiltration from the Scioto River at Columbus' South Well field by lowering the ground-water level about four feet under the stream level (Klaer and Assoc., 1976).

Typical municipal, commercial, and industrial wells drilled into the buried valley deposits along the Scioto River range in depth from 45 to 200 feet and yield 100 to 3000 gpm. The largest recorded yield is 2500 gpm from a collector well at the Mead Paper Co. at Chillicothe. Most municipal wells tapping the aquifer yield 500 to 1000 gpm.

Industrial wells, especially those operated by the Mead Paper Company at Chillicothe yield 1000 to 2500 gpm. Collector wells drilled for the same company produce a total of 40 mgd. The ERDA Plant at Piketon uses an average of 15 mgd, which is obtained from a group of wells along the river west and southwest of Piketon.

Industrial wells in the Circleville area range from about 60 to 180 feet deep and yield a total of 8 mgd. Tests conducted in Columbus' South Well field indicated that, under average conditions of river stage and water temperature, the total capacities of 5 collectors installed along the Scioto River would yield a minimum of 45.7 mgd (Klaer and Assoc., 1976).

According to the projections made by the Ohio Department of Natural Resources (in preparation), considerable industrial, economic, and agricultural growth is expected to occur well into the next century. This increase in growth will undoubtedly be coupled with an increase in the water supply requirements. The buried valley aquifer system is virtually the only dependable source to meet such
increases. It has been only moderately developed so far, has a very high potential yield, and conditions for development of ground water supplies along the aquifer anywhere from south of Columbus to Portsmouth are as favorable as any in the state.

There are three principal sources of recharge to the buried-valley aquifer. The most important of these is the direct infiltration from streams. The other two sources of recharge are percolation of direct precipitation over the valley itself and the infiltration of water through the till-covered bedrock uplands into the Scioto River Valley. Contribution to recharge as underflow from the consolidated rocks is localized and is significant only where the aquifer is underlain by carbonate bedrock. Underflow from the shale bedrock is relatively small. Recharge due to precipitation varies seasonally and occurs mainly between late fall and early spring. Recharge by induced infiltration occurs only in those places where withdrawals from the aquifer lower the ground-water levels below the stream level.

Ground water entering the buried-valley aquifer both from upland runoff and direct precipitation moves under low gradients to discharge into the Scioto River. Plate 1 shows the generalized configuration of the water table surface in the buried valley aquifer and is based on the elevation of water levels measured in wells that are in the files of Ohio Division of Water. The map was constructed with a contour interval of 10 feet. Water level elevations at selected well locations (Table 20) were plotted on a series of topographic maps at a scale of 1:63000. Surface elevations were estimated from
U.S. Geologic Survey 7.5 minute quadrangle maps. In order to obtain sufficient areal coverage of well locations, information covering the period of 1946-1976 was used. Hence the water level map obtained represents a composite level of the water table for this period of record. The map does not, however, represent the existing water level conditions except in areas where major developments and increase in withdrawal rates altered the original contour pattern. Inspection of water level hydrographs of representative observation wells (figs. 36 and 37 show that average seasonal variations in water levels occur within ± 4 feet from the average. The average depth to water level in feet below land surface is generally less than 40 in the glaciated area, mainly due to the thickness of overlying till material. The average depth to the water in the unglaciated area varies from 5 to 30 feet and averages approximately 20 feet.

The horizontal distance between the contours, indicate that the gradient along the flow lines perpendicular to the axis of the river varies from 10 to 40 feet per mile. The gradient increases toward the edges of the buried valley because of the changes in aquifer composition and the increase in the finer fraction away from the central axis of the buried valley.

The Scioto River is a gaining stream along its entire course from Columbus to Portsmouth, as can be seen from the contour and flow lines. All the contours converge upstream and intersect the
Figure 36. Hydrograph of observation well Pk-3 at Circleville showing seasonal fluctuations of the water level in the Buried-valley aquifer between 1959-1977 (After Ohio Div. of Water).
Figure 37. Hydrograph of observation well Pi-2 at Piketon showing seasonal fluctuations of the water level in the Buried-Valley aquifer between 1969 and 1977 (after Ohio Div. of Water).
stream pointing upward, a characteristic of gaining streams. The flow lines all converge towards the stream channel, which acts as a discharge line. Discharge areas, mainly due to heavy pumping, occur around major developed parts of the aquifer. These are characterized by cones of influence especially in Columbus, Circleville, and Chillicothe (Plate 1).

The rate of movement of ground water has been estimated to be in the order of 1-2 feet per day (Norris and Fidler, 1969).

The interaction between ground water and surface water systems and the hydraulic connection between these systems can best be demonstrated by estimating the amount of ground water discharge to the stream. A small percentage of the discharge of ground water into the Scioto River can be estimated by computing the difference in low flow (flow equaled or exceeded 90% of the time, generally assumed to be all ground-water runoff) between two gaging stations. For example, low flow at Chillicothe gage for 1967 was 450 cfs, and downstream at Higby it was 616 cfs. This is a increase of 166 cfs, but because the contribution of Paint Creek (27 cfs) is included in the flow at Higby, the actual increase along the valley is therefore 139 cfs or about 90 million gallons per day. The distance between Chillicothe and Higby, if measured along the axis of the valley, is 11 miles. The discharge of ground water into the river is approximately 8.2 mgd/mile of valley length (6.2 gpd/sq. ft. of streambed). Norris and Fidler (1969) gave an estimate of 1.8 mgd per mile of valley length. Contribution of ground water is impeded
or reversed in areas of concentrated withdrawals. Outside the areas of heavy pumping, the head of the ground-water table is higher than that of the stream and thus ground-water discharge into the stream. This usually peaks during the dry season when the river stage is lowest. When the river stage rises during the season of high precipitation, the situation may reverse and surface water recharges the ground-water table.

Chemical Composition of Ground Water in the Aquifer

Table 6 shows recent analysis of water samples collected from public water supply wells tapping the buried valley aquifer at different places along the aquifer. Most of the samples were analyzed by the Ohio Department of Health and their records are filed by the Ohio Environmental Protection Agency (OEPA). Four of these wells are used by OEPA to monitor the chemical composition of the aquifer and are sampled every six months. The wells range from 60 to 135 feet in depth; their locations are shown on Figure 38.

There seems to be no observable downstream trend in the concentrations of the major chemical constituents in water from the aquifer and the water quality varies from one well to another. This is related to the fact that ground-water flow in the aquifer is mostly laterally toward the Scioto River. The concentration of chemical constituents in water derived from the buried valley aquifer is generally influenced by the mineral composition of the unconsolidated material in the aquifer and in the rocks surrounding it. This is well expressed by the degree of hardness, which ranges from 288 to 519 mg/l
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1 EPA Quality Monitoring Well.  
2 Records from EPA.  
3 U.S. Geological Survey.  
4 Ohio Division of Water.
Figure 38. Locations of chemical quality analysis along the Brucied-valley aquifer.
(milligrams per liter), and dissolved solids, which range from 367 to 762 mg/l. The water is very hard and softening is required for many purposes. Dissolved solids and hardness can be related to the mineral composition of the material in the aquifer, which consists predominantly of limestone and dolomite pebbles derived from the carbonate terranes in the northern and western parts of the Scioto River Basin. Water in contact with such material will dissolve some of the constituents of these rocks, such as calcium and magnesium. Hardness due to high calcium and magnesium bicarbonates is considered temporary and may be removed by heating the water.

The higher concentrations of hardness and dissolved solids at Shadeville are attributed to the upward leakage of ground water from the carbonate aquifer that underlies the buried valley aquifer in that area. Allong (1971) reported dissolved solids that exceeded 1000 mg/l at Five Points in Pickaway County, and attributed it to leakage from the Silurian-Devonian aquifer. Further evidence of such leakage is shown by the distinct presence of hydrogen sulfide during aquifer tests conducted in the proposed Columbus South Well Field near Shadeville (Stilson and Associates, 1976).

The degree of hardness of the water is dependent on the concentration of calcium (Ca) and magnesium (Mg). Both cause boiler scale and encrustation of well screens.

The presence of high concentrations of magnesium in combination with chloride produces a water that is corrosive and damaging to boilers and other heating equipment. High concentrations of magnesium
beyond the safe drinking standards, 125 mg/1, can be cathartic to humans.

Iron concentrations range from 0.6 to 3.3 mg/1. The amount of iron in all samples exceed the limit recommended in the Federal drinking water standards of 0.3 mg/1. Although 0.6 to 3.3 mg/1 of iron does not pose any health hazards, this amount in public water supplies is objectionable because of its effects in staining clothes and fixtures, and in the process of food preservation. Iron also forms encrustations in pipes and on well screens as a result of bacterial activity, causing screens to clog and hence lowering the well capacity. Partial removal of iron in ground water can be accomplished by aeration to precipitate iron oxides, and by chlorinating the well during the high flood seasons in order to kill the bacteria and prohibit encrustation on the screen. High concentrations of iron are commonly related to leakage from the shale and sandstones that underlie or surround the aquifer.

Manganese (Mn) is present in concentrations that exceed the recommended standards (0.05 mg/1). The Mn concentrations range from 0.03 to 1.06 mg/1. In concentrations greater than 0.05 mg/1 it causes an undesirable taste, and stains laundry. It becomes objectionable for use by industries in concentrations greater than 0.2 mg/1.

Sulfate concentrations (SO₄) are less than 150 mg/1 except in the Shadeville well where the concentrations (257 mg/1) exceed the recommended limit of 250 mg/1. The high concentration is due to leakage through gypsum and evaporite deposits in the carbonate rocks that underlie the aquifer, and possibly from leaching gypsum fragments.
in the overlying till in the vicinity of Shadeville. High concentrations produce a bitter taste and may be cathartic to humans. Sulfate also combines with calcium to form scale that is especially resistant to heat and thus would require an increased use of energy to heat water in industrial boilers.

Chloride (Cl) concentrations are generally less than 50 mg/l and far less than the 250 mg/l limit recommended by Federal standards. Such low concentrations indicate no contamination to the aquifer from oil field brines or sewage effluent. Cases of ground-water contamination from oil-field brine pits were reported in Delaware and Morrow Counties by Pettyjohn (1971) and Baranovic (1975).

Other chemical constituents that are present in only minor concentrations include nitrates (NO$_3$), potassium (K), and fluoride (F). Their presence in such small concentrations poses no threat to the quality of ground water in the buried valley aquifer.

The presence of arsenic (As) has been reported in one analysis of water from the Waverly municipal well field. Although arsenic is present in the sample in concentrations below the recommended limits, its presence in public supply is alarming and indicative of potential ground-water pollution. Its presence may be related to the use of pesticides and herbicide compounds that contain arsenic oxides; in a public supply it is a potential danger because concentrations exceeding the recommended limit of 0.05 mg/l for long periods of time would be poisonous to humans and could cause lung, skin, and liver cancer.
The persistence of traces of ammonia and phosphate in most of the samples is an indication of a potential contamination to the aquifer. These compounds in ground water may be related to the use of fertilizers in agriculture. Ammonia and phosphate are used as fertilizers in the farming areas adjacent to the aquifer and are leached from the soil by infiltrating water.

Traces of barium have been detected in samples from Obetz, Circleville, Waverly, and Piketon. Barium additives are mixed in diesel fuel to reduce smoke emissions, and most of its exhausted as barium sulfate, which is insoluble in water and apparently harmless to humans (Hodges, 1973). However, the possibility of unrecognized dangers should not be overlooked. The traces of barium may have found their way into the ground by mixing with infiltrating rainwater.

The presence of barium in samples collected at the locations cited could also be due to leakage of fuel from the motor pump to the well and mixing into the water.

Theory and Development of the Digital Computer Model

The finite difference model developed by Prickett and Lonnquist (1971) was designed to simulate the response of an aquifer to an imposed stream. Several analytical models, such as the electric analog or Hele-shaw devices, are able to experimentally duplicate aquifer flow situations, but in many aspects the mathematical manipulations of a computer model are more flexible and, therefore, more adequately meet the needs of the modeler. Many mathematical models have been developed for use in aquifer evaluations. The
approximations of the partial differential equation of flow using finite difference equations, as originally developed for application to heat flow problems and oil reservoir evaluations, (Peaceman and Rachford, 1955; Douglas and Peaceman, 1955; Quon and others, 1965 and 1966; Fagin and Stewart, 1966) have been used most efficiently in problems restricted to the evaluation of two or three dimensional flow. The model by Prickett and Lonnquist followed previous attempts at modeling the response of ground water levels to pumping first attempted by the California Department of Water Resources (Tyson and Weber, 1964) and later by others, such as Fiering (1964), Eshett and Longenbaugh (1965), Remson, et al. (1965), Bittinger and others (1967), Pinder and Bredenhoeft (1968), Prickett and Lonnquist (1968a,b) and Taylor and Luthin (1969).

The digital computer techniques developed by Prickett and Lonnquist and used by the Illinois State Water Survey include generalized computer programs that will simulate one-, two-, or three-dimensional nonsteady-state flow problems in heterogeneous aquifers under water table, nonleaky, and leaky artesian conditions. The programs cover time varying pumpage from wells, natural or artificial recharge rates, the relationships between surface and ground water, evapotranspiration, and the conversion of storage coefficients from artesian to water table conditions.

The partial differential equation of non steady-state ground water flow in a confined nonhomogeneous isotropic aquifer in two dimensions may be written as (Prickett and Lonnquist, 1971):
\[ \frac{d}{dx} \left( T \frac{dh}{dx} \right) + \frac{d}{dy} \left( T \frac{h}{dy} \right) = S \frac{dh}{dt} + Q \]  

(\text{V-1})

in which

\( T \) = aquifer transmissivity

\( h \) = head

\( t \) = time

\( S \) = aquifer storage coefficient (dimensionless)

\( Q \) = net groundwater withdrawal rate per unit surface area of the aquifer

\( x, y \) = rectangular coordinates

There is no general solution to the flow equation, however, a numerical solution can be obtained with the finite difference approach. The finite difference approximations, which also provide the theoretical foundation for the electric analog approach, act as the basis for the digital model.

In order to solve the equation for a heterogeneous aquifer with irregular boundaries, one approach is to subdivide the region into rectangular blocks in which the aquifer properties are assumed to be uniform. The continuous derivatives are replaced by finite difference approximations for each nodal point. The result is \( N \) equations and \( N \) unknowns (head values at the nodes), where \( N \) is the number of nodes in the aquifer.

Utilizing a node-centered, finite difference grid in which variable grid spacing is permitted (fig. 39), the flow equation may be as (Prickett and Lonnquist, 1971):
Figure 39. Example of a finite difference grid system (After Prickett and Lonnquist, 1973).
\[ T_{i-1,j,2}(h_{i-1,j} - h_{i,j})/\Delta x^2 + T_{i,j,2}(h_{i+1,j} - h_{i,j})/\Delta x^2 \]
\[ + T_{i,j,1}(h_{imk+1} - h_{i,j})/\Delta y^2 + T_{i,j,1}(h_{i,j-1} - h_{i,j})/\Delta y^2 \]
\[ = S_{i,j}(h_{imk} - h_{0,i,j})\Delta t + Q_{i,j}/\Delta x\Delta y \]

where:

- \( T_{i,j,1} \) = aquifer transmissivity between nodes \( i,j \) and \( i,j+1 \)
- \( T_{i,j,2} \) = aquifer transmissivity between nodes \( i,j \) and \( i+1,j \)
- \( h_{i,j} \) = calculated heads at nodes \( i,j \) at the end of a time increment measured from an arbitrary reference level
- \( h_{0,i,j} \) = calculated heads at nodes \( i,j \) at the end of the previous time increment measured from the same reference level defining \( h_{i,j} \)
- \( \Delta t \) = time increment elapsed since last calculation of heads
- \( S_{i,j} \) = aquifer storage coefficient at node \( i,j \)
- \( Q_{i,j} \) = net withdrawal rate if positive, or net accretion rate if negative at node \( i,j \)

In the finite difference approach, the differentials and are approximated by the finite lengths \( x \) and \( y \).

A modified form of the iterative alternating direction implicit method (ADI) is used to solve the set of \( N \) simultaneous equations. The ADI technique, first described by Peaceman and Rachford (1955), solves two sets of matrix equations with each iteration. The equations for head values in columns are computed implicitly and those along rows are obtained from the previous column computations and are defined explicitly. After all equations have been solved
column by column and row by row, an "iteration" is completed. The calculated heads are then used as initial conditions for the next time increment, after the iterations have been repeated a sufficient number of times to achieve convergence. This technique is considered stable regardless of the size of the time increment used. A more detailed discussion of the mathematical development can be found in Prickett and Lonnquist (1971).

Application of the Digital Model to the Buried-Valley Aquifer at Piketon, Ohio

The digital model was designed for the buried-valley aquifer in the vicinity of Piketon to serve as an example for the use of such models in the evaluation of the availability and potential of groundwater development. It was also used for determining the configuration of the water table under different management conditions. The aquifer has been studied extensively over the past two decades and the accumulated data permit a close evaluation of the applicability of the model by comparing its results with those obtained in the field. The U.S. Geological Survey conducted numerous aquifer tests in the area and many test wells and production wells have been extensively developed. Analysis of these test results have been well documented (Norris and Fidler, 1965, 1966a, 1966b, 1967, 1969; Norris, 1970; Norris and Sedam, 1977). These tests also contain records of the extensive historical data in the development of the aquifer and daily withdrawal rates. The growing increase in the ground water demands and further increase in the withdrawal rates for more industrial
development in the area also favored the choice of this area for digital modeling.

The modeled area is located along the Scioto River Valley and lies in the vicinity of Piketon in the unglaciated part of the aquifer (fig. 40). It is rectangular, 1.8 miles wide and 4.17 miles long, and has an area of 7.5 square miles. The city of Piketon, with a population of 1,400, is in the center of the modeled area. The Scioto River flows along the eastern side of the model boundary in the northern third of the model, then loops across to the western side in the vicinity of Piketon and continues southward then southeast toward the lower boundary. Other drainage channels that enter the model area from the western edge include Peepee and Sunfish Creeks, which are tributaries to the Scioto River. Beaver Creek enters the area from the east and flows southwards into the Scioto River.

Industrial development in the buried valley aquifer began in 1964 with the ERDA Uranium Enrichment Plant pumping 9 mgd for cooling purposes from the aquifer. A feasibility study concluded that further development of the aquifer in the area could be undertaken and withdrawals greater than 25 mgd could be obtained from a series of wells along the Scioto River sustained by induced infiltration from streamflow (Norris and Fidler, 1969)(fig. 41). By March, 1976, the total withdrawal rate was 13 mgd and it is predicted to be 31.3 by the year 1980 (Table 4).
Figure 40. Location of modeled area of the Buried-Valley Aquifer.
Figure 41. Locations of wells pumping from the Buried-Valley aquifer in the Piketon area. The wells are identified by reference to the grid coordinates (i,j).
Aquifer Framework

Inspection of the bedrock contour map (fig. 42) and cross-sections (figs. 43 and 44) reveals a relatively deep channel approximately 2,500 feet wide incised into the bedrock and underlying the existing Scioto River valley in the western part of the area. This deep channel was apparently formed as part of the Deep Stage drainage. As a result of glaciation, the channel was filled with outwash that averages 70-85 feet in thickness in the deeper part of the channel. The bedrock valley becomes shallower eastward away from the incised channel and the thickness of unconsolidated material decreases by 15-20 feet (fig. 43). The saturated thickness of the aquifer is about 20 feet less than the total thickness of the outwash.

The material in the aquifer is mostly coarse sand and medium-grained gravel, which is described by the drillers as clean and loosely compacted. In some places the coarse sand and gravel are interbedded with finer sand layers that are tighter, however, although they function to a slight degree as confining layers, they do not greatly affect the flow of water in the aquifer and the aquifer responds hydraulically as one unit during aquifer tests (Norris and Fidler, 1969). The glacial outwash is overlain by a layer of brown soil and recent alluvium, consisting mainly of sand, silt, and clay, that varies from 5 to 10 feet thick.

Model Framework

In order to design the digital model of the aquifer, the area of
Figure 42. Generalized bedrock contour map showing the bottom of the aquifer.
(After Walker et al., 1965)

(Figure 43. Cross sections showing the buried valley near Piketon.)

(From Ohio Water Plan Inventory, Underground Water Resources, Scioto River Basin, File Index M-15)
Figure 44. Map of Piketon area showing locations of cross sections.
the aquifer was discretized by using a finite difference variable grid of 25 columns by 49 rows. The grid is superimposed on the area of the aquifer as shown in Figure 45. The finite difference variable grid was chosen to provide necessary details in areas of interest that surround the pumping centers. The grid interval in the pumping centers is 250 feet per node and increases to 500 feet towards the west, north and south. The total area of the model, 7.5 square miles, is discretized into 1,225 nodes. Each node is referenced by a two dimensional array having coordinates (i) for column location and (j) for row location within the finite difference grid. The actual aquifer has an area of 4.5 square miles. Its eastern and western limits are defined by impermeable or barrier boundaries where the aquifer thickness diminishes toward the sides of the valley. The northern and southern boundaries are defined as recharge boundaries along which no drawdowns or changes in water levels occur, and the aquifer is assumed to have an infinite extension beyond those two boundaries. The barrier boundaries are incorporated in the model by assuming transmissibility and permeability values of zero for all the nodes that are outside the aquifer boundaries. Recharge boundaries are defined by magnifying the storage factor values on the recharge boundary nodes by at least 100 times that used in the aquifer nodes.

Data Input to Model

In preparing data for input into the model, a series of maps that describe the aquifer properties and parameters had to be
Figure 45. Map of Piketon area showing finite difference grid.
prepared either from existing maps or by constructing new ones from available data, and extrapolating where possible in areas where data were lacking. Among the maps prepared were bedrock surface map (fig. 42), water-table elevation map (fig. 46), elevations of land surface (read from topographic map), and transmissivity and permeability distributions (fig. 47). All these properties and parameters were discretized by superimposing the finite difference variable grid over these maps, and each node of the grid was assigned a value for each of the following parameters:

1) the \((i,j)\) node coordinates that define the location of the node in the grid,
2) transmissivity values in the column and row direction,
3) the storage factor for both water table and artesian conditions,
4) initial head or water table elevation,
5) recharge rate,
6) land surface elevation,
7) elevation below which evapotranspiration ceases,
8) elevation of top of aquifer,
9) permeability both in the column direction and row direction, and
10) elevation of bottom of the aquifer.

The coefficient of transmissivity \((T_1,T_2)\) was entered into the model as an average value obtained from aquifer tests conducted at the Atomic Energy Commission's well field sites. Norris and Fidler (1969)
Figure 46. Part of the water table surface, based on depth to water in wells measured July 12, 1963 (after Norris and Fidler, 1969).
The coefficient of storage, determined from aquifer tests, by Norris and Fidler (1969) ranges from 0.18 to 0.22, typical of a water table situation. The average value of 0.20 was entered as part of the model data. The storage factor (SF) was calculated for each node using the following equation (Pricket and Lonnquist, 1971):

\[
SF(i,j) = 7.48 S \Delta x \Delta y
\]  

(V-3)

where

\[
SF(i,j) = \text{storage factor for node located at coordinates } i,j, \text{ in gallons per foot (gal/ft)}
\]

\[S = \text{the aquifer storage coefficient, a fraction}\]
Figure 47. Map showing the distribution and variations in Input Data for the Piketon Model.
7.48 = number of gallons in a cubic foot of water, in gallons per cubic foot (gal/ft³)

Δx, Δy = finite difference grid intervals, in feet.

The storage factor, SF, is entered in the model program at SF1 for artesian conditions, and SF2 for water table conditions.

The elevations of the initial head (H, also known as water table elevations) were discretized from the water table contour map based on depth to water in wells measured prior to 1963, and additional unpublished data in files of the Ohio Department of Natural Resources, Division of Water (fig. 46).

Elevations of land surface (RH) were discretized from U.S. Geological Survey 7½ minute topographic mpas; elevations at each node were extrapolated, from the contour maps, to the nearest foot.

The average depth below which the effect of evapotranspiration ceases in the area varies significantly, depending on many factors, such as depth to which roots penetrate into the soil, type of cover, slope of the land surface, soil moisture deficiency, the intensity and duration of sunlight, number of days between crop irrigations, amount of precipitation and its distribution, depth to the water
table, and wind velocity. All these factors, at least to some extent, control the depth to the water table. The most significant of these factors and the one that is most effective in determining this average depth is the depth to which roots penetrate. The combined maximum depth to which roots penetrated in the general area of the aquifer and as extrapolated from comparable areas in Ohio was estimated to be 24 to 48 inches in agricultural crop lands and 40 inches to 20 feet in forest land (Jim Bonta, North Appalachian Experimental Watershed Research Center - Agricultural Research Service, personal communication). The average depth of roots ranges from 18 to 24 inches in crop areas and less than 10 feet in wooded areas (Prof. Dale A. Ray, Agronomy Dept., O.S.U., personal communication). On the basis of these estimates, and because vegetation in the area consists mainly of corn and soybeans in the middle of the valley and woods on the sides, a combined average value of 12 inches was chosen as the depth below which evapotranspiration ceases. This depth value was subtracted from the land surface elevation for each node location and entered as elevation below which evapotranspiration ceases, except at the river nodes where this elevation is replaced with the elevation of bottom of the stream bed.

The top of the aquifer (CH) was assumed to be the same as the land surface elevation (RH) due to the absence of any confining beds overlying the aquifer. The aquifer is open to the surface and the unsaturated part acts as extra storage area during recharge.
periods. Figure 48 is a computer generated isopach map of the saturated part of the aquifer obtained by calculating the difference between the elevation of the water table (H) and elevation of bottom of the aquifer (BOT).

The coefficient of permeability (PERM) was entered into the model calculations as a uniform average value of 2,500 gpd/ft², except where reported variations occurred. The values reported were entered and extrapolated to the adjoining areas influenced by these values. Before entering the data for permeability, adjustments for using a variable grid had to be made due to the variability in the vector spacing between the node locations. This was accomplished as follows:

\[ \text{PERM}(i,j,1) = P(\text{ave.}) \times \frac{\Delta x}{\Delta y} \] for permeability calculated in an increasing row order

\[ \text{PERM}(i,j,2) = P(\text{ave.}) \times \frac{\Delta y}{\Delta x} \] for permeability calculated in an increasing column order

\[ \Delta x = \text{distance in feet between nodes } (i,j) \text{ and } (i+1,j) \]

\[ \Delta y = \text{distance in feet between nodes } (i,j) \text{ and } (i,j+1) \]

Figure 47 shows the distribution of permeability input values in different areas of the aquifer.

The elevations of bottom of the aquifer were obtained from structure contour maps of top of bedrock in the valley (Walker et al., 1965; Norris and Fidler, 1969) and from other well log records. Figure 42 is a contour map of the bedrock surface elevations generated by computer.
Figure 48. Isopach map showing the saturated part of the aquifer in the Piketon area.
In order to estimate the average recharge rate to the aquifer, a preliminary survey of the recharge rates computed from the hydrograph separation techniques, mentioned in an earlier chapter, was made. The average recharge rate as computed for the Scioto River at Chillicothe for 1967 is 0.36 mgd/sq. mi. (fig. 7) (equivalent to 7.5 inches of rainfall). The average rate of recharge at Higby is 0.46 mgd/sq.mi. (equivalent to 9.62 inches of rainfall.) Tuller (1967) estimated the rate of recharge for the average year from hydrograph separation techniques to be 0.31 mgd/sq.mi. (equivalent to 6.7 inches of rainfall) at Chillicothe and 0.34 mgd/sq. mi. (equivalent to 7.05 inches of rainfall) at Higby. An average value of 0.37 mgd/sq.mi. (equivalent to 7.7 inches to rainfall) was assumed on the basis of the above rates. However, at an early stage in the calibration of the model, it was found that the assumed recharge rate is too low and had to be increased in order to obtain a situation similar to that in the field. The average value for the recharge rate entered into the model was increased to 0.66 mgd/sq.mi. (equivalent to 13.8 inches of precipitation per year) before the simulated steady state water table began to resemble field data.

Norris (personal communication) estimated the average rate of recharge to the ground water table to be equivalent to a minimum of 12 inches of rainfall per year (0.57 mgd/sq.mi). This is a much closer approximation to the average rate used in the model simulation than rates obtained from hydrograph separation techniques, and supports the validity of the recharge rate that was used in the model.
The recharge rate was converted for each node to a recharge factor \((R)\) according to the equation (Prickett and Lonnquist, 1971):

\[
R(i,j) = \left( \text{ave. recharge rate gpd/ft}^2 \right) \Delta x \times \Delta y 
\]

\(V-4\)

and was entered in the model as a negative withdrawal rate \((-Q)\) to avoid confusion with the stream bed recharge factor (induced infiltration) which was entered as \((R)\).

Induced infiltration through the stream bed is calculated using the equation (Prickett and Lonnquist, 1971):

\[
R(i,j) = \frac{P'}{m'} \times A_s 
\]

\(V-5\)

where

- \(R\) = Recharge rate (gpd)
- \(P'\) = permeability of the stream bed gpd/ft\(^2\)
- \(m'\) = thickness of the stream bed in feet
- \(A_s\) = Area of the stream bed in ft\(^2\)

According to Norris and Fidler (1969) the average width of the Scioto River in this area is 260 feet. The average thickness of the streambed material is 5 feet. The material consists of coarse sand and gravel thinly strewn with larger stones and cobbles; in places the bed is covered with several inches of mud and organic debris. The vertical permeability of the stream bed as reported by Norris and Fidler (1969) averages 27 gpd/ft\(^2\).

The stream channel was modified into vector volumes at each stream node location. Each vector volume extends the full depth of the stream bed \((m')\) and has a horizontal area of 260L, where \(L\) is the length of the stream segment represented at the specific node location. The infiltration rates were entered uniformly over the stream bed nodes, assuming uniform stream bed characteristics. However during testing of the model it became apparent that the
stream bed infiltration rates were higher than they should be and the increase or decrease in head on the river nodes had to be brought about by suppressing the infiltration rates - indirectly increasing or decreasing the permeability of the stream bed. It was found that the model responds best to changes in this parameter and its use was critical in bringing the initial heads to match those obtained in the field. The average permeability of the stream-bed was found to range from 2.3 gpd/ft\(^2\) to 15.4 gpd/ft\(^2\). The reason for the variation is mainly attributable to differences in the nature and composition of the streambed material.

The following assumptions have to be kept in mind when the infiltration rates through the stream bed are concerned (Prickett and Lonnquist, 1971): 1) the wells fully penetrate the aquifer; 2) the drawdown in the flow field is small compared with the saturated thickness of the aquifer; 3) the pumped well is far enough from the stream that the effects of partial penetration of the river is negligible; and 4) the head in the surface water body remains constant.

Infiltration rates were also computed for nodes under Lake White. Leakage through the lake bed has a noticeable effect on recharge to the aquifer. The accumulation of silt and clay sediments in the lake was at the average annual rate of one tenth of a foot as calculated by Sanderson (1948). If this rate is projected up to 1977 the average thickness of sediments that are presently accumulated in the lake bottom is about 4.7 feet. Initial permeability
of sediment material was assumed to be 1 gpd/ft², however when adjustments of infiltration rates were made during the process of simulating field conditions the permeability values were increased to an average of 2.3 gpd/ft².

Other data entered into the model include parameters that are used in the operational control of the model and to provide data for simulating parts of the model assumed to have homogeneous and isotropic properties with identical initial conditions at all nodes. Such parts of the model include the nodes that fall outside the area of the aquifer, and their nodes are controlled by data entered on a default card.

The parameter card contains: 1) the number of time increments (NSTEPS), which correlates with the number of simulations desired; 2) the time increment in days for each simulation (DELTA), 3) ERROR - this term represents the maximum allowable sum of the absolute values of changes in head (in feet) for all node points of the model during an iteration.

The default value card contains data that sets default values for each note point outside the aquifer system. Since each node belonging to the aquifer has its own card, the nodes that lie outside the aquifer have an equal set of data for each node (default values) representing a homogeneous discharge boundary.

A pump parameter card is included when pumping is introduced to the model. The pump parameter card contains data such as the number of pumping nodes in the model (NP), the number of time increments per pumpage change (NSP), and the number of pumping rates in the
pumping schedule (NRT). Following the pump parameter card is
a pumping schedule card for each pumping node that contains
the (i,j) location of the pump node and the punched value of each
pumping rate in the pumping schedule.

Following the preparation of the previous data and punching
it on computer cards, it is entered as data into a modified version
of the basic aquifer simulation program of Prickett and Lonnquist (1971)
as listed in Appendix B and the model is ready to be tested.

Phases of Modeling

Every hydrologic model is uniquely designed based on the nature
of the particular problem under study. In setting up the models
used in this study, emphasis is placed on the following objectives:

1. Amending or justifying conclusions arrived at from
   previous hydrological investigations in the area;
2. Evaluating the effect of stream recharge on the
   ground-water development potential;
3. Testing the effects of projected increases in
   withdrawals from the aquifer;
4. Recommending best possible locations of well fields
   for future planning.

For the digital model to be capable to describe the situation that
exists in the buried-valley aquifer it must first be validated.
A valid hydrologic model is one in which the simulated results
closely correspond to those calculated by field methods. In order
to do that three basic phases were involved:
1. Testing the digital model setup and the program operational sequence;
2. Calibration of the model;
3. Verification of the model.

The last two phases are the most important.

The first phase consisted of running the computer program without incorporating any recharge or discharge parameters for an exaggerated period of time \((1 \times 10^{20} \text{ days})\). The purpose of this run (or group of runs) was first to debug the program, and to correct errors in the data or in the program's operational sequence, secondly to test the validity of the model response by inspecting the water levels at each node or at representative nodes scattered as monitoring nodes over different locations in the model. This was accomplished by expanding the computer program to include a print-out graph of water level declines at each of the monitoring node locations. The different initial water levels \((H_0)\) should decline or rise at each node to an average level that represents the average of all levels. Theoretically the water levels should approach the horizontal level after an exaggerated period of time for each simulation trial.

**Model Calibration**

The second phase was to calibrate the model and to achieve a steady state situation that compares model output with that obtained from field measurements. To achieve this, parameters of recharge, and leakage from or to the stream bed and the lake were introduced. Heads recorded on each node card were chosen as initial heads, and the program was run in an attempt to duplicate the initially assigned
water levels that were obtained in the field prior to the introduction of pumpage, hence achieving a calibrated steady state situation.

Figure 46 is a water-table contour map of the Piketon area based on measurements taken on July 12, 1963 (Norris and Fidler, 1969). This map represents the steady state situation because little pumping has been done except at the Piketon municipal well, whose effect was only local. The map also serves as the earliest recorded data representing the natural conditions. The model had to undergo many changes in many of its parameters such as permeability, recharge rate, and stream bed infiltration rates before the simulated levels began to approach those obtained by field methods. A few trial runs were made, each having an exaggerated period of time \(1 \times 10^{20}\) days, and the water level elevations calculated at the end of each run for each node location in the model were outputed on a set of cards. These cards were later used as input data for yet another computer plotting program (Appendix C) that contains instructions to the computer to plot a water table contour map from the inputed elevations using the VERSATEC Plotting emulator. Figure 49 shows a typical output generated from this process.

The water table map obtained at the end of each run was then compared with the field map; the differences were noted and the proper changes in aquifer parameters were made before another run was made. The process was repeated until the simulated levels were as close as possible to those obtained by field methods (figs. 46, 49). The runs that were made and changes ensuing from them resemble the fine tuning of an instrument. Achievement of a steady situation was
Figure 49. Simulated water-table contour map under natural condition.
further checked by running the model for a few short time increments and observing the time-water level graphs of the representative nodes. There was no noticeable change in the levels and the graphs fell on horizontal lines. Hence the steady state calibration was completed.

By the time the steady state situation was achieved the original input average values of permeability and transmissivity were reduced by 20 percent in some areas of the model. The stream bed permeability ranged from 3.5 to 17.6 gpd/ft$^2$.

Approximately 65 computer runs including many adjustments or combinations of adjustments of aquifer parameters were made before the best fit of the steady state conditions was obtained. This involved about 45 minutes of the IBM 370/168 computer central processing unit time.

After the steady state situation was achieved, the resulting head values on each node of the model were then stored on a disk to be used as initial head values in the following simulations.

Model Verification

In order to test the validity of the model, it was necessary to determine how it stands when compared with independent field observations and tests. To do that, a nine day aquifer test was simulated to determine the aquifer parameters and to compare them with results obtained from field tests carried out by Norris and Fidler (1969). Node (9,16) was established as a pumping well that pumped an average of 1,000 gpm and the model was run for nine steps of one day each. The drawdowns at nodes (8,16), (7,16), and (6,16)
were observed for the third time step (after 3 days of pumping), and the drawdowns were plotted on semi-logarithmic paper against the distance of each node from the pumping node as seen in Figure 50 and Table 7. The transmissivity and storage coefficient of the aquifer were then calculated from the graph by using the Cooper and Jacob equations (in Walton, 1970):

\[
T = \frac{528 Q}{As} \quad (V-6)
\]

\[
S = \frac{Tt}{4790 r^2} \quad (V-7)
\]

where

- \(T\) = coefficient of transmissivity in gpd/ft
- \(Q\) = pumping rate in gpm
- \(As\) = the amount of drawdown over a log cycle in feet
- \(S\) = storage coefficient, unitless
- \(t\) = amount of elapsed time since pumping started, in minutes
- \(r_0\) = the distance at which the drawdown on the projected line is zero, in feet

When the equations were solved, the coefficient of transmissivity as obtained from the solution is 215,500 gpd/ft and the storage coefficient is 0.23. Norris and Fidler (1969), in a similar field test using different methods, obtained an average transmissivity of 207,000 gpd/ft and an average storage coefficient of 0.20 as determined from the time-drawdown curves of different observation wells.

The closeness in the comparison of the results thus gives confidence in the validity of the model and its accuracy in simulating the true situation as they exist in nature.

Having established such needed confidence and validity, modeling
<table>
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<th>Node</th>
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<th>drawdown in feet</th>
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<td>0.23</td>
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</table>
Figure 50. Distance-drawdown graph along parallel line after 3 days of pumping at a rate of 1000 gpm.
the burial-valley aquifer can now be taken one step further. That is to predict the aquifer potential and to measure the influence of future stresses on the behavior of the aquifer.

**Projections of Ground Water Supply in the Piketon Area**

The Ohio Department of Natural Resources is in the process of publishing a comprehensive study of the Central Ohio Water Plan. The study concentrates in details on the analysis of historic water supplies in the area and on projecting the demands on water supplies for the next four decades. The projections are based on historic data and on projections of future growth in population, industry, and agriculture.

The Central Ohio Water Plan does not show a remarkable growth in the demand for public water for Piketon. Projections for public water demands ranged from 0.25 mgd in 1980 to 0.31 mgd for 2020 (Table 3). Industrial demand projections were given as total projections for entire counties; projected demands for Pike County ranged from 6.29 mgd in 1980 to 7.39 mgd by 2020 (Table 5). These projections fall short of the existing industrial use, because the ERDA plant alone used 13 mgd in 1977. The projected needs of the plant for the next two decades reach 25 mgd with the continued expansion of the plant. The projection of demands for public water, as reported by the study, are based on demands that are served by a public water supply system and excludes all demands that would be met by privately owned wells.
The projections given for Piketon in this study are based on historical pumping rates until 1976 and include the projected demands of the ERDA plant, and other projected industrial demands as reported in the Central Ohio Water Plan Study (fig. 51).

Ground Water Levels at the End of 1975, 1976

Historical records of pumpage for the period 1963 to 1976 were obtained and the yearly pumpage rates were entered in the model for each pumping node (fig. 51). Beginning in 1964, the ERDA plant pumped an average of 9 mgd from four wells located on the river meander west of Piketon (Nodes (10,17), (9,16), (8,16), (7,16). Other private and public supply wells raised the total withdrawal rate to 11.3 mgd. The distribution of wells in the aquifer is shown in Figure 41. The locations of the wells were moved to the nearest grid node in order to fit a node location. Total withdrawal rate increased slowly from 1964 to the end of 1975 when it reached an average of 12.3 mgd. This increase in production resulted mainly from the drilling of new private and municipal wells. The biggest increase in withdrawal from the aquifer began in March, 1976, when the ERDA plant increased its total pumpage from the aquifer to an average of 12.3 mgd. The model was set to simulate the ground water levels at the end of 1975. Figure 52 shows the water table contours as of that time.

The composite cone of depression that resulted from extended pumpage began to develop when pumpage started and it continued to deepen and spread with each pumping increase. Figure 52 shows the extent of the cone at the end of 1975 and the increase in the water
Figure 51. Ground Water Demand Projections for Piketon, Ohio.
Figure 52. Water-table contour map showing water levels at the end of 1975.
level gradient in the area of recharge - in this case the Scioto River. Water levels are lowest at the pumping centers, and the cone is deepest at pumping nodes (8,16) and (9,16) where the water level elevations at node (8,16) dropped from 531.4 feet in 1964 to 513.0 feet in 1975; a total drawdown of 18.4 feet. The gradient of the cone is steepest (0.033 ft/ft) in the immediate vicinity of the pumping nodes, mainly due to pumping and recharge from the river bed. However, the cone flattens slowly towards the southwest where the gradient becomes 0.010 ft/ft. The effect of recharge by seepage through the bed of Lake White is clearly shown by the patterns of contour lines southeasterly from the lake area. The cone of depression extends farthest towards the southwest in areas of minimum recharge and away from the stream.

Table 8 shows well characteristics and aquifer characteristics at the end of 1975, as derived from this study, and compares them to results obtained by Norris and Fidler (1969). Their results were obtained by pumping each of the wells individually, with the other wells remaining idle, and hence the differences. Also, their tests were run for only an average of 1 hour per test. There is a close correlation between the results from the two sets of tests as seen from the specific capacity values obtained for each well.

Other much smaller cones began to develop in other areas of the model, such as those around nodes (3,11), (13,20), (15,31) and (18,38). The spread of the cones around each node is localized except around node (13,20) where the smaller cone is developed within the area of influence of the major cone. Pumping from the node commenced

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<td>1200 11.24 106.8</td>
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in 1971 at a rate of 625 gpm. The effects of the smaller cone are minor. The total drawdown at node (13,20) is 6.95 feet, and the specific capacity is 90 gpm/ft of drawdown.

Groundwater Levels at the End of 1976

Beginning with 1976, the ERDA plant increased its production from 9 to 13 mgd, pumped from 15 wells spread around the river bend. In programming the pumping schedule, it was necessary to distribute the total amount of withdrawal from the area among 10 pumping nodes, each having a rate of 1.3 mgd (fig. 41). In doing this, the pumping rate was actually lowered on the previous four nodes that were pumping 2.3 mgd each. Table 9 shows some of the aquifer characteristics derived at each of the ERDA pumping nodes. Although pumping rates and amount of drawdown have increased at each of the 1976 pumping nodes, their specific capacities are lower in 1976 than in 1975, perhaps due to the increase in the radium of the composite draw-down cone and thickness of the already dewatered section.

Figure 53 shows the configuration of the water table at the end of 1976. Inspection of the water table contours shows that the major cone of influence around the concentrated withdrawal area has deepened; the center has shifted from node (8,16) to node (7,20), where the drawdown is 20 feet; and the area has increased markedly. The 530 contour line, closed on the 1975 map, expanded farther and recessed towards the center of the valley, indicating a drop in the water levels from the center of the cone. The spread of the cone is largely away from the river and towards the southwest. The decrease
### TABLE 10: Characteristics of Wells and Aquifer at End of 1976

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Figure 53. Water-table contour map showing water levels at the end of 1976.
in the distance between the contours beneath the river bed in the lower section of the cone is partially due to the effects of recharge and also to the decrease in the permeability of the material in this area.

The smaller cone around node (13,20) has spread farther outward and is expected to merge within the larger cone. Conditions pumping in the rest of the pumping areas resulted in little change in the water table configuration.

The pumping schedule for 1976 was uniformly continued until 1980 in order to simulate water level elevations at the end of 1980. No increase in the existing pumping rates and no new pumping areas in the model area were assumed. Inspection of the predicted water table contours for the end of 1980 (fig. 54) shows that the cone has deepened still more and the lowest drawdown occurs at node (7,20) where it is about 23.2 feet. The area influenced by the cone increased towards the south and the 528 contour has advanced more in that direction. The smaller cone around node (13,20) was absorbed by the larger cone and its former presence is shown only by the curvature of the 526 contour. Other localized cones show a very minor increase in diameter, as shown by the cone around node (15,31); its effects are still localized.

**Projection of Water Levels From 1981-1990**

The projected total withdrawal rate from the buried valley aquifer in the model area at the end of 1980 is 16.33 mgd. Early in 1976, the ERDA plant started to explore the feasibility of increasing
Figure 54. Water-table contour map showing water levels at the end of 1980.
their production from the aquifer from 9 to 25 mgd, an increase of 16 mgd. Part of this increase was to be met by increasing production from the existing well field to the present rate of 13 mgd from 10 wells. The U.S. Geological Survey made a study (Norris, personal communication) of the aquifer characteristics and determined the best site for a new well field along the Scioto River. On the basis of their study, they recommended that the new well field be located along the Scioto River near the confluence of Sunfish Creek and the Scioto River (Norris, personal communication). The proposed increase in production at the ERDA plant of 16 mgd was later reduced to only 4 mgd, which was taken from wells in the existing field as shown in the 1976-1980 pumping schedule. However, ERDA has not entirely abandoned the plan and the possibility remains that the additional 12 mgd will be needed by 1981. The projected increases in public water demands and in other industrial water demands are also incorporated in the total projected withdrawal rate as of the beginning of 1981 as seen in Figure 51.

Eight nodes were chosen as pumping nodes for the proposed ERDA south well field (nodes (15,44), (15,45), (16,45), (17,45), (17,46), (18,46), (19,46), and (20,46)), to supply the projected total demands of 31.5 mgd (fig. 41). The eight wells were set to pump 12 mgd (1.6 mgd per well), and two additional nodes with locations (8,18, and 6,22) in the existing well field were added to the pumping schedule with a pumping rate of 1.6 mgd each. This brings the total withdrawal rate from the model at the beginning of 1981 to 31.5 mgd. The pumping schedule was continued from 1981 to 1990 and simulation of the water-
Table elevations at the end of this period is shown in Figure 55. Table 10 shows the characteristics of the wells and the aquifer at the end of the pumping period. Inspection of the cones of depression (fig. 55) shows a remarkable increase in the area of the cone in the central part of the model, due to the effects of increased pumpage. The cone has also deepened; the greatest drawdown, 34.7 feet, is at node (7,20). The adding of pumping node (8,18) resulted in the steeper gradient of the cone, as shown by the clustering of the contours, and in the development of a minor cone within the major cone.

The relations of drawdown to yield and specific capacities of the wells were determined as outlined in Johnson (1972, p. 107). The amounts of drawdown at each well are still above the critical limit and they represent an average of only 38.5 percent of the maximum drawdown, 62 percent of maximum yield and 80 percent of the maximum specific capacity.

In contrast, the cone of depression developed in the new well field in the southern part of the model area has a much greater effect on the configuration of the water table elevations. The aquifer is narrower here than in the northern part of the area and the cone intersects the discharge boundaries represented by the sides of the valley very early, and further expansion of the cone is along the axis of the aquifer. The distance between the contours is more uniform and the gradient averages 0.010 ft/ft as measured from the center towards the northeast. It is much steeper (0.024 ft/ft) towards the south and southwest due to the Scioto River recharge boundary. The
Figure 55. Water-table contour map showing water levels at the end of 1990.
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<tr>
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<th>Pumping rate (gpm)</th>
<th>St. W. level before pumping</th>
<th>St. W. level after pumping</th>
<th>drawdown (ft)</th>
<th>sp. capacity (gpm/ft)</th>
<th>elevation of bottom of aquifer (ft)</th>
<th>saturated thickness (ft)</th>
<th>% of max. drawdown</th>
<th>% of max. sp. cap.</th>
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Table 10. Characteristics of wells and aquifer at the end of 1990.
maximum drawdown, 26.19 feet, occurs at node (17,45) and represents 44 percent of the maximum drawdown and 70 percent of the maximum yield. 44 percent of maximum drawdown represents 77 percent of the maximum specific capacity, and hence the wells are approaching the minimum allowable specific capacities.

The increase in gradient and areal spread of the composite cone in the south resulted in an increase in the surface area of recharge. The cone almost intersected the discharge boundary to the east (the eastern edge of the valley wall) at an early stage in the pumping schedule.

The average infiltration rate from the stream bed was calculated from the stream recharge rates that were inputed to the model. The average infiltration rate is approximately 0.91 gpd/ft$^2$ or 0.04 mgd/acre. Norris (1977, personal communication) calculated an infiltration rate of 0.05 mgd/acre from a recent aquifer test conducted in the same area.

The spacing of wells affects the behavior of the cone of depression in the southern part of the model area. Because the grid in this area is of a larger dimension (500 x 500 feet), it was not possible to have more pumping nodes in the vicinity of the river. The withdrawal rates had to be divided between a minimum number of wells that would increase the drawdowns at each well and affect the nature of the cone.

Other changes in the water levels include the expansion of the cones around nodes (15,31) and (11,3).
1991-2000 Projections of Ground-Water Demand and Their Effect on the Water Table

Projections of the ground-water demands for this period did not increase significantly over the previous period, however, an additional 0.6 mgd were added to the south well field to cover the projected increase in the demands for this period, making the projected demand in 1991-2000 a total 32.3 mgd.

Figure 56 shows the configuration of the water table, the cones of depression, and the water-table elevations as of the end of year 2000. Table 11 summarizes the characteristics of the wells and the aquifer within the area of concentrated withdrawal. By comparison with the 1990 ground-water table (fig. 55), it is evident that the north cone of depression has already stabilized and reached its maximum area of influence. There is no noticeable change in the amount of drawdowns at any individual well in the area influenced by this cone and the water level elevations have not changed. However, the cone of depression in the southern part of the model area has intersected the discharge boundary to the east and has spread to the north and south along the axis of the valley. The depth of the cone in its central part as shown by node (17,45) has increased to 30.2 feet below the initial static water level. The increase in the size and depth of the cone is due to the projected increase in the water demands. In addition the confinement of the cone by the valley walls caused the cone to expand to the north and south.

The spread of the cone is more pronounced southward as can be seen from the displacement of the contours (example, contour interval
Figure 56. Water-table contour map showing water levels at the end of 2000.
<table>
<thead>
<tr>
<th>Node location</th>
<th>Pumping rate (gpm)</th>
<th>St. W. level. before pumping</th>
<th>St. W. level after pumping</th>
<th>drawdown (ft)</th>
<th>sp. capacity (gpm/ft)</th>
<th>elevation of bottom of aquifer</th>
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<th>% of max. drawdown</th>
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Table 11. Characteristics of wells and aquifer at the end of 2000.
This displacement causes an increased amount of induced infiltration from the streambed and results in a decrease in the gradient of the water table below the stream bed. Consequently, the cross-sectional area of the induced infiltration increases. Much induced infiltration is expected to occur and replenish the aquifer in the area influenced by the cone.

Another important reason for the displacement of the contours is the expansion and contraction of the cone of depression due to its proximity to the southern and eastern boundaries of the model. The central cone appears to have reached a steady state situation, whereas the cone that developed in the southern section seems to oscillate from its center as will be demonstrated later. The cone oscillated within 2 feet over a period of twenty years (1990-2010, 2000-2020). This indicates that the water levels as calculated by using the flow equations are oscillating around the real solution; this is due to the instability inherent in the use of the finite difference approximation.

2001-2020 Projections

The projected ground water demands change very little after the year 2000. An increase of 0.87 mgd was scheduled at the Piketon municipal water supply beginning with the year 2001 and no additional increase to the total demand was scheduled until the year 2020. The model was scheduled to simulate the periods 2001-2010 and 2011-2020 with the same pumping schedule.
Figure 57 shows the configuration of the water table at the end of year 2010. The increase in withdrawal from node (14,20) resulted in a minor increase in drawdown on the rest of the pumping nodes (Table 12). The area of influence of the cone has increased as the size of the cone increased (compare Figures 56 and 57).

In comparing the amounts of drawdowns that occurred in the southern cone, as shown in Tables 11 and 12, there is a decrease in drawdown at the end of 2010. This increase represents the instability of the cone and its oscillation around the real solution.

The central cone stabilized again by the end of 2020, as seen in Figure 58, and the characteristics of the wells and aquifer remain unchanged (Table 13). The only significant change in the water table configuration at the end of 2020 is the opening up of the small cone developed around node (15,31). This represents a decrease in the water table elevations in this area and the lowering of the water table due to pumping, is spreading to the outer areas and in and from the pumping centers. The southern cone, however, oscillates around the position it occupied at the end of 2000 (fig. 56).

Conclusion

The buried valley aquifer in the Piketon area is composed of a heterogeneous mixture of coarse sand and gravel with minor silt and clay layers - variations in composition from place to place in the aquifer have a direct effect on the aquifer potential and on its hydrologic characteristics. The aquifer is unconfined and open to direct recharge from precipitation. There is also a direct connection
Figure 57. Water-table contour map showing water levels at the end of 2010.
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<th>Node Location</th>
<th>Pumping rate (gpm)</th>
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Table 12. Characteristics of wells and aquifer at the end of 2010.
Figure 58. Water-table contour map showing water levels at the end of 2020.
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<th>Pumping rate (Gpm)</th>
<th>St. W. level before pumping (ft)</th>
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<th>drawdown (ft)</th>
<th>sp. capacity (gpm/ft²)</th>
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Table 13. Characteristics of wells and aquifer at the end of 2020.
between the aquifer and the Scioto River, and leakage from the aquifer into the stream is occurring where the stream intersects the water table. Induced infiltration through the stream bed into the ground water occurs where reversal of the water-table gradient occurs as in areas influenced by pumpage. The model results show that the induced infiltration rates range from 0.5 to 3.1 gpd/ft$^2$ of streambed, and depend on the permeability of the stream bed.

The aquifer has a very high potential and ground-water storage far exceeds the projected water supply demands in the area. The modeled area of the aquifer is 5.4 square miles and its average thickness is approximately 55 feet. Assuming a porosity of 30 percent and a specific yield of 25 percent, the total amount of ground water in storage is approximately 15.4 billion gallons of which 12.8 billion gallons are available for withdrawal. Average recharge rate to the aquifer, as determined from digital model simulation is approximately 660,000 gpd/sq. mi.

Modeling the aquifer has shown that the projected withdrawal plans are realistic and that the aquifer can easily sustain the projected demands without excessive drawdowns.

Maximum drawdowns of 35.2 feet are predicted for the year 2010 and corresponds to maximum withdrawal of 17 mgd from the area of the central cone west of Piketon. This drawdown represents only 56 percent of the maximum drawdown and shows that greater withdrawals are feasible. The model projections have also shown that the steady state balance between withdrawals and recharge is reached within one year.
in the central pumping area.

The model has performed successfully as a tool to simulate realistic future demands. The model was calibrated and verified against field tests, and could be successfully used for testing any new development schemes that may be planned in the area.
VI. HYDROGEOLOGY OF THE SILURIAN-DEVONIAN CARBONATE AQUIFER

The Silurian-Devonian aquifer in the Scioto River Basin includes all the carbonate rock units that lie between the Silurian Osgood Shale and the overlying Devonian Ohio Shale (fig. 11). Composed of hydraulically interconnected beds of limestone and dolomite, the aquifer consists, in ascending order, of the Lockport Dolomite, the Bass Islands Group, and the Columbus and Delaware Limestones. These rocks crop out or lie beneath glacial deposits in an area of approximately 3000 square miles in the western two-thirds of the Scioto River Basin (fig. 31). Towards the east they are overlain by the Ohio and Olentangy Shales of Upper Devonian Age. The Lockport Dolomite crop out in a narrow belt in the southwestern corner of the basin and sub crops beneath glacial till along the eastern flank of the Cincinnati Arch throughout most of Fayette, Clinton, and Highland Counties. The Bass Islands Group is the most extensive and sub crops beneath the till in the western and northwestern part of the basin in Highland, Fayette, Madison, Champaign, Union, Logan, Marion, and Hardin Counties. The Columbus and Delaware Limestone sub crop beneath till in a narrow wedge, approximately 8 miles wide, that extends from southwestern Crawford County to northwestern Franklin County. Where exposed, the carbonate
sequence ranges in thickness from a few feet along the eastern flank of the Cincinnati arch where most of the section has been removed by erosion, to at least 650 feet in the northeastern part of the basin (fig. 59). The variations in thickness are due mainly to the stratigraphic thickening of the carbonate section towards the Appalachian structural basin and to post depositional erosion of parts of the section following uplift of the Cincinnati Arch.

The Osgood Shale and some Upper Ordovician shales, at the base of the aquifer, provide a seal that separates the aquifer from Lower Ordovician carbonates and prohibits the upward migration of water from them. The Ohio and Olengany shales, which overlie much of the aquifer in the eastern half of the basin, act as a aquitard and restrict vertical movement of water to or from the aquifer.

As shown by structure contour maps (figs. 60 and 61), the carbonate beds dip easterly, average from the crest of the Cincinnati Arch, over most of the basin. In the extreme northwestern corner of the basin in Logan and Hardin Counties, the beds are on the crest of the Cincinnati Arch and have a very gentle dip away from its crest to the north and northeast. The regional dip of the carbonate section in the basin averages about 25 feet per mile as measured along a line extending from east-Central Champaign County into eastern Franklin County (Norris and Fidler, 1973).

Figure 62 shows the configuration of the bedrock surface. This surface is a relatively flat upland surface deeply incised by broad steep-sided valleys more than 300 feet deep. The bedrock surface is a part of a well developed peneplain (Lexington Peneplain
Figure 59. Isopach map showing digital thickness of the Silurian-Devonian carbonate aquifer in the Scioto River Basin (contour interval is 100 feet).
Figure 60. Digitized map showing structure contours on bottom of the Silurian-Devonian carbonate aquifer (contour interval is 100 feet).
Figure 61. Digitized map showing structure contours on the top of the Silurian-Devonian carbonate aquifer (contour interval is 100 feet).
Figure 62. Configuration of the bedrock surface in the Scioto River Basin (after Allong, 1971).
of Fenneman, 1938), which was uplifted in several stages, the last of which was the late Tertiary time and resulted in the deep trenching of valleys such as the Teays Valley and its principal tributaries. The altitude of the bedrock surface ranges from about 750 to 950 feet in most of the glaciated part of the basin, and reaches 1100 feet in Champaign County in the west and Monroe County in the northeast. In Madison and Champaign Counties the buried Teays valley is at least 400 feet lower than the general surface of the Lexington Peneplain (Morris, 1958). The hills bordering the Teays Valley in the unglaciated southern part of the basin rise 200 to 300 feet above the valley floor and the maximum relief on the bedrock in this area exceeds 600 feet.

Hydrologic Properties of the Silurian-Devonian Aquifer

Aquifer Permeability

The Silurian-Devonian aquifer consists of limestone and dolomite whose permeability has been developed chiefly as a result of solution by water moving through joints, bedding planes, and intergranular openings. Most of the permeability was developed after the uplift of the Cincinnati Arch and in conjunction with the erosional processes by which beds younger than the Silurian-Devonian carbonates were progressively removed from the rising arch. The Cincinnati Arch is known to have emerged above the sea several times before the end of the Paleozoic Era and each uplift and emergence caused fracturing of the rocks. The large and extensive openings within the aquifer resulted from the circulation of ground water through these fractures. Norris
and Fidler (1971) indicated that the beds that underwent most solution and those now having the highest permeability lie on the flanks of the arch in northwest Ohio (fig. ).

The most prominent zone of high permeability occurs in the lower part of the Bass Islands Group. This zone, known as the "Newburg Zone", is about 15 feet thick and lies from 3 to 5 feet above the top of the Lockport (fig. 12). Recognized throughout much of eastern Ohio where the carbonate rocks are deeply buried, it is characterized by large cavities and solution channels. It is a major source of water in many deep wells. Norris and Fidler (1973) described the historical development of the permeability and related the development of the Newburg Zone to the position of the water table prior to glaciation. They state:

"The zone of maximum permeability in southwest Ohio is determined to a large extent by the range in the position of the water table over geologic time. In the eons following the final emergence of the southwest Ohio land mass there no doubt occurred a certain amount of upward and downward shifting in the position of the water table, perhaps over a range of several tens of feet, as suggested by the extensive development of permeability vertically within the aquifer. A major readjustment of the water table occurred with the onset of glaciation. Before the Pleistocene the streams in western Ohio were deeply entrenched in the limestone. The master stream of the preglacial drainage system, the Teays River -- its valley now abandoned and deeply filled with glacial drift in much of western Ohio -- declined in altitude from 568 feet in south-central Madison County to 538 feet in southern Champaign County (Norris and Spicer, 1958, p. 218). By comparison, the altitude of the bed of the Great Miami River at Dayton is approximately 720 feet, and the altitude of the bed of the Scioto River at Columbus is about 680 feet. Thus, the water table in southwest Ohio just before glaciation was significantly lower than it is today, perhaps as much as 200 feet lower in some areas. Development of permeability was related to the position of this former
base level, which in much of the study area is below the base of the carbonate aquifer. In the time required for establishment of this low base-level control, the water table moved through virtually the full thickness of the carbonate aquifer, resulting in the development and enlargement of solutional openings at all levels.

More detailed discussions on the genesis and development of permeability in the Newburg Zone are found in Stout (1935), Norris (1956), Norris and Fidler (1971 and 1973).

**Specific Capacity**

Aquifer test data and geophysical logs of test holes in the carbonate aquifer reveal the existence of a regionally extensive zone of relatively high permeability; wells drilled into this zone have relatively high yields. The zone appears to underlie an irregularly shaped area that includes virtually all of Union and Hardin Counties and significant parts of Marion, Logan, Delaware, Champaign, Madison, Franklin, and Pickaway Counties (fig. 63), and that extend northward in northwestern Ohio in two strips that lie on either side of the crest of the Cincinnati Arch. The specific capacity of wells drilled in this high yield area ranges from 5.4 to 106 gpm per foot of drawdown (gpm/ft) and averages 30 gpm per foot (Norris and Fidler, 1973).

The specific capacity of wells drilled in areas outside the high yield zone are much lower, range from 0.5 to 5.0 gpm per foot, and average 2.6.

**Transmissivity**

Short-term aquifer tests conducted by the Ohio Division of Water for several test wells drilled in the carbonate aquifer gave wide ranges in transmissivity, mainly due to the heterogeneous
Figure 63. County outline map of Ohio showing location of area of high-yielding wells (shaded) in the carbonate-rock aquifer. Heavy broken lines bound present study area; dots mark axis of Cincinnati arch (after Norris and Fidler, 1973).
nature of the secondary openings, both vertically and horizontally. Transmissivity values ranged from 6,800 to 210,000 gpd/ft and averaged 27,700 in the high yield area. In the rest of the aquifer some wells have been reported dry, or they yielded too little water even to conduct aquifer tests.

The Ohio Division of Water also conducted aquifer tests in Logan and Hardin Counties in which they reported transmissivity values of 10,000 and 12,000 gpd/ft, respectively. Walton (1962), reported a transmissivity of 126,000 gpd/ft in a test conducted in the carbonate aquifer at Ada in Hardin County. Walker and others (1970) reported transmissivity values from aquifer tests conducted in Sinica, Allen, and Sandusky Counties in northwest Ohio. Their reported values were 9,600, 19,250, and 3,700 gpd/ft, respectively.

Rowland (1969) computed the average transmissivity of the Silurian-Devonian aquifer in the Upper Auglaize River Basin from two cones of influence that developed due to pumping around Findlay and Lima. His reported values were 10,500 gpd/ft for Lima and 15,500 gpd/ft for Findlay. He also calculated the transmissivity on the basis of well radius by correcting the specific capacities for partial penetration and concluded that for design purposes an average transmissivity of 50,000 gpd/ft is a good value on which to predict regional drawdown in the aquifer.

The wide variations in permeability and transmissivity in the carbonates are mainly due to the random distribution of fracture systems and solution channels. Large solution channels and subterranean streams can develop along zones of weakness and result
in turbulent flow systems and high storage capacities, whereas the rest of the rock remains relatively impervious and characterized by little or no flow and storage capacity.

**Recharge to the Aquifer**

Much of the recharge to the carbonate aquifer is derived from vertical leakage through the overlying glacial deposits where the potentiometric surface of the aquifer is lower than the water table. The distribution of recharge is variable throughout the aquifer due to changes in the thickness and composition of the glacial drift. Recharge is highest where the drift is thin or where it contains sand lenses and lowest where the drift is thick.

Another significant and effective form of recharge occurs where the aquifer is associated with buried valleys. Most of the recharge to the Newburg occurs in areas where the zone crops out along the walls of these buried valleys. Several of these buried valleys (fig. 62) have been reported by Norris and Spicer (1958), Schmidt (1958), and Walker, et al. (1965). Leakage from the valley fill, commonly sand and gravel, into the Newburg zone occurs through the extensively developed solution caverns and fracture systems and contributes to the higher yields in this zone.

Direct recharge from precipitation occurs only in a small area of outcrop of the Lockport outside the glacial limit in the southwestern corner of the basin. Its contribution to the overall recharge is negligible and its significance is restricted to wells pumped locally for domestic supplies.
Due to the lack of extensive aquifer tests and restricted development of the aquifer, little is known about the rates of recharge to the carbonate aquifer in the Scioto River Basin. In evaluating recharge to the carbonate aquifer, much reliance was given to estimates determined in other carbonate areas with similar hydrologic characteristics. Rowland (1969) determined the rate of recharge to Silurian-Devonian carbonate aquifer in the Upper Auglaize River Basin from the analysis of 13 cones of influence. He calculated recharge rates ranging from 6,800 to 75,300 gpd/sq. mi. with an average of 21,200 gpd/sq. mi. Furthermore, he also estimated recharge rates by stream hydrograph separation techniques; the rates varied from 6,020 to 56,700 gpd/sq. mi. and averaged 17,600 gpd/sq. mi. Walton (1965) estimated the recharge to a drift-covered dolomite aquifer in Illinois in which the rates varied from 64,000 to 225,000 gpd/sq. mi. and averaged 201,000 gpd/sq. mi.

Bloyd (1974) calculated the ground-water recharge for several sub-basins in the Ohio Region using the 60-percent flow-duration data. He assumed that the 60-percent flow to be indicators of natural ground-water recharge. By this method Bloyd calculated the average rate of recharge in the Scioto River Basin to be approximately 73,000 gpd/sq. mi.

Estimates of recharge rates for the carbonate aquifer in the Scioto River Basin were made from analyses of stream hydrographs. These estimates were based on the records of flow for one year (1967) representing the near average flow in the basin. Ground-water runoff was separated from total runoff for each gaging station on streams within
that part of the basin that is underlain by the subcrop of the Silurian-Devonian aquifer (fig. 64). The recharge rate estimated by this method varied from 123,000 to 310,000 gpd/sq. mi. and averaged 247,500 gpd/sq. mi. There is a close correlation between this average recharge rate and the average rate obtained by Walton. Differences in the above estimates are due to the variability of the drift composition and presence or absence of buried outwash valleys and sand and gravel lenses in the till; all of which have direct effects on the vertical permeability of till and the recharge rate to the Silurian-Devonian aquifer. Also the above values should not be confused as being the actual amount of recharge that reaches the carbonate aquifer, since a considerable part of the ground-water runoff is derived from the till itself and from sand and gravel lenses buried in the till. These estimates can at best be considered as the average ground-water runoff contributed from the drainage area above the stream gaging station. The part of recharge that actually reaches the carbonates is somewhere below the average rate given above. The recharge factor applied in the aquifer model averaged below that determined by the hydrograph separation technique, as will be shown in a later section.

Water in the Silurian-Devonian aquifer occurs under both confined and unconfined conditions. The confined aquifer is more common due to the low permeability of the overlying glacial drift and the progressive deepening of the zones of high permeability away from the flank of the Cincinnati Arch. Water in the confined areas of the aquifer will rise
Figure 64. Recharge rates for the Silurian-Devonian aquifer estimated from hydrograph separation. Numbers are in 1000's of gpd/sq. mi. Heavy line represents the eastern limit of the Silurian-Devonian Carbonate Subcrop.
to an imaginary surface, defined as the potentiometric surface, when the aquifer is penetrated by a well. Depth to water in feet below land surface in wells drilled into the carbonate aquifer typically varies from 10-50 feet. In topographically low areas water levels in relatively deep wells that penetrate the Newburg rise above the water table. In some areas the head is above the land surface and locally wells flow in Zane township in Logan County and Stokes and Pike townships in Madison County.

Figure 65 shows the configuration of the potentiometric surface for the carbonate aquifer. The contour map represents average water-level conditions throughout a variable period of time, best representing the water level conditions for the present decade (mid sixties to mid seventies).

Water enters the regional flow system in the higher parts of the basin along the topographic divides in the west, northwest, and northeast and flows toward the principal discharge areas, which include the Scioto and Olentangy Rivers (fig. 65). The Scioto River flows over the carbonate outcrop in most of its channel north of the Pickaway - Franklin County line. The Olentangy River also flows over bedrock in Delaware County. Natural discharge from the carbonate aquifer occurs along the stream channels in these areas as well as into the overlying sand and gravel aquifers that are in the buried valleys overlying the carbonate bedrock.

The potentiometric surface slopes east-southeast towards the center of the basin, from the high points in the west, with an average gradient of 12 feet per mile. The direction of flow correlates
Figure 65. Map of the Scioto River Basin showing approximate configuration of the potentiometric surface in the Silurian-Devonian carbonate aquifer. (partially after Norris and Fidler, 1973).
with the regional dip of the carbonate rocks. Flow in the eastern part of the basin however, is from northeast to southwest. Flow also occurs from high points in the northwest and out of the basin towards the north and northeast.

**Potential Yields and Utilization**

Limestone and dolomites are a major source of ground water in much of central and western Ohio. The Silurian-Devonian carbonate aquifer is a principal source of water for domestic, municipal, and industrial use in the western two-thirds of the Scioto River Basin. Potential yields of 100 to 500 gpm are available from wells drilled in most of the subcrop area of the aquifer (fig. 32) and yields of 500 to 600 gpm are available in the high yield area (fig. 63). Yields of 25 to 100 gpm can be obtained from wells drilled in the weathered layers of the Lockport Dolomite in the southwestern corner of the basin, while 5 to 25 gpm are generally available from wells in the upper few feet of these carbonates in the higher upland area along the western boundaries of the basin.

The estimated average daily public use of ground water from the carbonate aquifer in 1974 was about 9.6 mgd (Table 14). The average private industrial use was about 13 mgd. The increase in population, expanding industries, and the rising costs of treatment of surface water will undoubtedly lead to a more extensive use of the carbonate aquifer as a main source of water supply.

Present development of the aquifer is centered in areas of major population and industrial concentrations. The largest total
TABLE 14. GROUND WATER AVERAGE DEMAND PROJECTION FOR PUBLIC SUPPLY FROM THE CARBONATE AQUIFER. (After O.D.N.R., In preparation)

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withdrawal occurs in Columbus (Table 16) where an average of about 10 mgd are pumped to satisfy the needs of private industries. Other major pumping centers are in Marion where an average of 4 mgd are used for the public supply system and in Marysville and Kenton, where use is 2.5 and 2.2 mgd, respectively.

Tables 14 and 15 show the statistical projections of public and industrial ground-water use from the carbonate aquifer. The projections are based on data in a recent study, the Central Ohio Water Plan (in preparation), that is being conducted by the Ohio Division of Water. Future demands are directly proportional to increases in population, economic growth, and per capita use of water. The per capita use in 1974 from public supplies in the basin ranged from 31 to 388 and averaged 128 gpd per capita for all public water supply systems in the area (Ohio Division of Water, in preparation). Projections of future water use from public water supply systems from 1974 to 2020 were based on an increasing rate of 1.3 gallons per day per capita per projected year. The public water supply projections in Table 16 include both supplies that are used by public and municipal water systems and industries. Projections provided in the Central Ohio Water Plan Study indicate that public supplies will provide approximately 14 percent of the industrial water needs in 1980, increase to 17 percent in the year 2000, to 20 percent in the year 2020. This indicates also that more industries will depend on the public supply systems and more water will have to be provided from the carbonate aquifer.
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<th>1974</th>
<th>1980</th>
<th>2000</th>
<th>2020</th>
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</tr>
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</tr>
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</tr>
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</tr>
<tr>
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<td>12.97</td>
<td>15.90</td>
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<td>24.65</td>
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</table>

(1) Modified from data in Table 40, O.D.N.R., Div. of Water, 1977 based daily average demands.
TABLE 15. PROJECTION OF TOTAL AVERAGE SUPPLY OF GROUND WATER FROM THE CARBONATE AQUIFER IN THE SCIOTO RIVER (after Ohio Div. of Water, in preparation).

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<td>Alger</td>
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<td>2.522</td>
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<td>2.458</td>
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<tr>
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<td>Hardin</td>
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<td>0.125</td>
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<td>0.137</td>
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<tr>
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<td>Champaign</td>
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<td>0.271</td>
<td>0.26</td>
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</tr>
<tr>
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<td>Union</td>
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<td>.08</td>
<td>0.099</td>
<td>0.117</td>
<td>0.135</td>
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<tr>
<td>Mt. Sterling</td>
<td>Madison</td>
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<td>0.092</td>
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<td>0.112</td>
<td>0.123</td>
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<td>0.077</td>
<td>0.105</td>
<td>0.131</td>
<td>0.157</td>
</tr>
<tr>
<td>Plain City</td>
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<td>0.574</td>
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<td>0.798</td>
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<tr>
<td>Prospect</td>
<td>Marion</td>
<td>-</td>
<td>0.138</td>
<td>0.15</td>
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<td>0.168</td>
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<tr>
<td>Richwood</td>
<td>Union</td>
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<td>0.357</td>
<td>0.432</td>
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<td>Sabina</td>
<td>Clinton</td>
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<td>0.312</td>
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<td>0.13</td>
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<td>0.13</td>
</tr>
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<td>0.061</td>
<td>0.067</td>
<td>0.071</td>
<td>0.075</td>
</tr>
<tr>
<td>Washington C.H.</td>
<td>Fayette</td>
<td>0.417</td>
<td>0.448</td>
<td>0.545</td>
<td>0.642</td>
<td>0.678</td>
<td>0.713</td>
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<tr>
<td>West Jefferson</td>
<td>Madison</td>
<td>0.45</td>
<td>0.592</td>
<td>0.867</td>
<td>1.141</td>
<td>1.374</td>
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<td>West Mansfield</td>
<td>Logan</td>
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<td>0.070</td>
<td>0.095</td>
<td>0.119</td>
<td>0.137</td>
<td>0.155</td>
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<tr>
<td><strong>Total</strong></td>
<td></td>
<td>22.489</td>
<td>26.838</td>
<td>32.645</td>
<td>38.452</td>
<td>41.198</td>
<td>43.944</td>
</tr>
</tbody>
</table>
Industrial water supply projections are influenced by the costs of providing the supply and by the pollution abatement requirements that are set to meet the stream water quality standards. The projections are made on the basis of the per capita water use for the number of employees in each industry. Two basic assumptions are emphasized in the methods by which the industrial supply projections were obtained. The first assumption is that each gallon of water currently withdrawn for manufacturing is being used twice before it is discharged as a plant effluent. The second assumption is that the industrial productivity (efficiency in labor, machinery, and manufacturing processes) will increase 24 percent in the first decade and remain constant for the rest of the projected period. Table 15 shows that industries in Columbus produce more water from the carbonate aquifer than do industries in any other town.

Projections of the total demand for water by public and industrial water systems are shown in Table 16. These total average daily demands, withdrawn in different parts of the carbonate aquifer, were entered as pumping nodes in the aquifer model. The influence of these demands on the water levels in the aquifer are significant in Columbus, Marion, Marysville, Kenton and Delaware. The total demands are expected to increase 19 percent by 1980, 72 percent by 2000 and almost double by 2020.

Chemical Composition of Ground Water in the Aquifer

The chemical composition of water from the carbonate aquifer is strongly influenced by the soluble minerals of the rocks, the ground-
water flow system, and the length of time the water remains in contact with the rock. With increasing time and distance traveled the concentrations of dissolved minerals will increase and the dissolved solids concentration will increase also leading perhaps to deterioration of the water quality in the aquifer.

Norris and Fidler (1973), related the chemical quality of water in the carbonate aquifer in southwest Ohio to the regional flow system. In studying the chemical composition of water from the carbonate aquifer, they were able to identify three basic water types that are related to the regional flow system. Wells in the recharge areas along the higher morainal terraines yield a calcium bicarbonate water. As it moves downward and eastward toward discharge areas, the water becomes progressively more mineralized, changing to a calcium sulfate type in the discharge areas. The change is attributed basically to the presence of gypsum and anhydrite, which occur as localized impurities throughout the carbonate webs in western Ohio. Gypsum dissolves more readily in water than dolomite and calcite and its solubility is not dependent on the presence of carbon dioxide. Hence the sulfate concentrations increase as the water moves from areas of recharge to areas of discharge. The reverse is true for the bicarbonate ion. Its proportion as a fraction of the total anion concentration in ground water decreases due to the addition of sulfate and chloride. Norris and Fidler (1973) demonstrated this fact by mapping the ratio of the bicarbonate ion concentration to the total anions in water samples from wells in the carbonate aquifer in different
places in the regional flow system. The change in the ratio of bicarbonate ions to the total anions in representative samples from deeper wells in the carbonate aquifer in the Scioto River Basin are shown in Figure 66. Higher bicarbonate percentages, generally greater than 80 percent, occur in the recharge areas along the western and northeastern boundaries of the basin. The ratio decreases eastward (southwestward in Morrow, Crawford, and Marion Counties) towards the center of the basin in the lowland areas where discharge normally occurs along the valleys of major streams, such as the Scioto and Olentangy Rivers in Delaware and Franklin Counties. The bicarbonate ratio in these areas is below 40 percent.

The potentiometric surface map of the carbonate aquifer (fig. 65) and the bicarbonate ratio map (fig. 66) both show the general direction of regional flow. Higher bicarbonate ratio occurs where the potentiometric surface is also relatively high, and lower bicarbonate ratio occurs where the potentiometric surface is relatively low. Areas of relatively low bicarbonate percent (generally less than 40 percent) occur in a trough-shaped area that extends from the Scioto River Valley in Northern Pickaway, Franklin and Delaware Counties northwestward through Union and Hardin Counties. This area is characterized by relatively low potentiometric surface levels and is considered to be a major discharge area in which flowing wells are known to occur, especially in Delaware, Union, and Hardin Counties.
Figure 66. Map of Scioto River Basin showing ratio in percent of bicarbonate ion concentration (meq/l) to total anions in water from the Silurian-Devonian carbonate aquifer (after Norris and Fiedler, 1973).
The major chemical constituents of selected water samples from 12 wells tapping the regional flow system of the carbonate aquifer in the basin are shown in Table 17. The analyses were grouped in three sets, each consisting of four wells representing the chemical composition of water in each of the recharge, intermediate and discharge areas of the regional flow system with in the aquifer, in order to further explain the relation between the water quality and the regional flow system. The use of this method was initially reported by Norris and Fidler (1973) and is limited here to that part of the carbonate aquifer within the Scioto River Basin. The locations of wells from which the samples were taken are shown in Figure 66. Evidence of change in the chemical composition of the water as it moves from areas of recharge through intermediate areas to areas of discharge is shown by these three sets of analyses. With the exception of the bicarbonate ion, the concentration of all other major ions generally increase from recharge areas to discharge areas. This confirms further the relation between the chemical quality of the water and the regional flow system.

The average concentration of calcium in wells tapping the recharge areas is about 93 mg/l, increasing to 129 mg/l in the intermediate area to 357 mg/l in the discharge area. Dissolved solids increase from an average of 454 mg/l in the recharge area to 696 mg/l in the intermediate area and to 1,832 mg/l in the discharge area. For sulfate, the average values for the same areas are 78, 250, and 1,095 mg/l, respectively, and for chloride they are 4, 10, and 34 mg/l. Conversely
TABLE 17: Chemical Analysis of Water from Selected Wells Drilled into the Silurian-Devonian Carbonate Aquifer and Sand and Gravel Aquifers in the Scioto River Basin Showing the Relation Between Chemical Characteristics and the Regional Hydrologic Environment (partly after Norris and Fidler, 1973).

<table>
<thead>
<tr>
<th>Well Number</th>
<th>Regional Area</th>
<th>Regional Depth (ft.)</th>
<th>Depth to Bedrock (ft.)</th>
<th>Yield (gpm)</th>
<th>Silica (S_2O_3) (mg/l)</th>
<th>Fe (mg/l)</th>
<th>Mn (mg/l)</th>
<th>Ca (mg/l)</th>
<th>Mg (mg/l)</th>
<th>Na (mg/l)</th>
<th>K (mg/l)</th>
<th>Total Cations (mg/l)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Recharge Area</td>
<td>250</td>
<td>138</td>
<td>600</td>
<td>15</td>
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<td>83</td>
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<td>4.3</td>
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<td>2</td>
<td>Recharge Area</td>
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<td>275</td>
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<td>1.1</td>
<td>0.05</td>
<td>100</td>
<td>4.99</td>
<td>35</td>
<td>2.87</td>
<td>18</td>
</tr>
<tr>
<td>3</td>
<td>Intermediate Area</td>
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<td>70</td>
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<td>0.53</td>
<td>0.22</td>
<td>140</td>
<td>6.99</td>
<td>44</td>
<td>3.62</td>
<td>56</td>
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<td>500</td>
<td>19</td>
<td>2.6</td>
<td>0.05</td>
<td>140</td>
<td>6.99</td>
<td>56</td>
<td>4.61</td>
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<td>6.99</td>
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<td>65</td>
<td>450</td>
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<td>1.2</td>
<td>0.09</td>
<td>200</td>
<td>9.30</td>
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<td>--</td>
<td>90</td>
<td>4.89</td>
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<td>14</td>
<td>--</td>
<td>2.3</td>
<td>--</td>
<td>--</td>
<td>98</td>
<td>4.89</td>
<td>36</td>
<td>2.96</td>
<td>3.4</td>
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</table>
TABLE 17: Chemical Analysis of Water from Selected Wells Drilled into the Silurian-Devonian Carbonate Aquifer and Sand and Gravel Aquifers in the Scioto River Basin showing the Relation Between Chemical Characteristics and the Regional Hydrologic Environment (partly after Morris and Fidler, 1973). (continued)

<table>
<thead>
<tr>
<th>Well Number</th>
<th>D.S. mg/l</th>
<th>Hardness as CaCO₃ mg/l</th>
<th>HCO₃⁻ mg/l</th>
<th>SO₄²⁻ mg/l</th>
<th>C mg/l</th>
<th>F mg/l</th>
<th>NO₃⁻ mg/l</th>
<th>Total Anions mg/l</th>
<th>Bicarbonate Total Anions Ratio</th>
<th>Sp. Conduct. Micro-Anions ohm/m</th>
<th>pH</th>
</tr>
</thead>
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<td>1c</td>
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<td>300</td>
<td>404</td>
<td>6.62</td>
<td>27</td>
<td>0.56</td>
<td>2.0</td>
<td>0.06</td>
<td>1.5</td>
<td>0.09</td>
<td>0.1</td>
</tr>
<tr>
<td>2c</td>
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<td>6.98</td>
<td>91</td>
<td>1.90</td>
<td>4.0</td>
<td>0.11</td>
<td>1.7</td>
<td>0.10</td>
<td>0.00</td>
</tr>
<tr>
<td>3</td>
<td>572</td>
<td>460</td>
<td>404</td>
<td>6.62</td>
<td>170</td>
<td>3.54</td>
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<td>0.00</td>
</tr>
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<td>2</td>
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(a) Carbonate aquifer is overlain by shale. (b) From Allong, 1971. (c) Flowing wells.
the total bicarbonate proportion in relation to the total anions decrease along a flow line. It averages 422 mg/l in recharge areas, 390 mg/l in intermediate areas, and 343 mg/l in the discharge area.

The general increase in the ionic concentrations of the preceding constituents reflects clearly the progressive mineralization of water as it moves through the regional flow system. The slight decrease in the bicarbonate ion is caused by the change from a calcium bicarbonate water in the recharge area to a calcium sulfate water in the discharge area.

Water in the carbonate aquifer is characterized by great differences in the amount of dissolved constituents from one place to another. The water is generally hard (350-2200 mg/l) but it can be used for most purposes after softening and removal of iron. The reason for this great variation is possibly the presence of gypsum or leakage of highly mineralized brine water from deeper aquifers and aquicludes. Dissolved solids range from 370-2850 and increase eastward where the carbonates are overlain by thick shale and sandstone. Water from wells in recharge areas have lower concentrations of dissolved solids due to the recharge from precipitation.

Dissolved iron in ground water is especially troublesome. Its excessive presence in water will stain laundry, fabrics and fixtures. As a result of bacterial activity iron also forms
encrustations in pipes. Iron, present in most of the samples from wells tapping the carbonate aquifer, generally exceeds the recommended limit of 0.3 mg/l. The average concentration in samples from 25 wells scattered throughout the Scioto Basin averaged about 1.9 mg/l. In some cases the concentration exceeds 5 mg/l and in one well in Delaware, where the aquifer is overlain by shale and sandstones, the concentration of iron is 29 mg/l. Iron is mostly derived from clays and other impurities in the glacial drift above the aquifer. It is also derived through leakage of water from shales overlying and in some places underlying the carbonate aquifer.

Fluoride is reported in all of the samples, and in most of them its concentration exceeds the recommended U.S. Public Health limit of 1.5 mg/l. Fluoride in concentrations higher than 1.5 mg/l may cause dental fluorosis in children and hence constitutes grounds for rejecting the water for drinking purposes. Fluoride in water derived from carbonate rocks may be related to the occurrence of the mineral fluorite, (CaF₂), which may occur in veins and as cavity fillings, or as minor impurity in most carbonate rocks of marine origin.

The evidence of upward leakage of water from the carbonate aquifer into the overlying glacial till and buried sand and gravel aquifers can be indirectly shown from the chemical analysis of water samples. In attempting to explain the evidence of this leakage the following assumptions were made: 1) barring any leakage from other aquifers or pollution from surface water, the chemical composition of water in the sand and gravel aquifers should not vary significantly from one
place to another within the basin, and 2) the quality of water in the sand and gravel aquifers in the recharge area of the carbonate aquifer should not be affected by the quality of the water in the carbonate aquifer itself due to the movement of water downward into the carbonate in this area.

Samples 13 to 16 in Table 17 show the chemical composition of water from wells in sand and gravel aquifers in the recharge and discharge areas (sample 15 and 16) of the carbonate aquifer. The locations of the wells are shown in Figure 66. Samples 13 and 14 were collected from wells in the recharge area of the carbonate aquifer and samples 15 and 16 from wells in its discharge area. Comparison of the major chemical constituents in these two sets shows a very great difference in the values. Since the wells of both sets penetrate only the sand and gravel above the carbonate aquifer and occur in areas where no other geologic formations overlie it, one can only assume that the difference in quality of water must be influenced by the regional flow system in the carbonate aquifer.

Concentrations of major chemical constituents of water samples from sand and gravel in the recharge area differ markedly from those in a similar aquifer in the discharge area. Sulfate in water from the sand and gravel in the discharge area is almost 10 times higher than in the recharge area, furthermore, hardness and dissolved solids almost double. On the other hand, a comparison of the concentration of chemical constituents in water derived from sand and gravel and the carbonate aquifer in the same area shows a close similarity in
most of the constituents. This is due to the fact that recharge water into the carbonate aquifer originates in the overlying sand and gravel and their respective qualities, therefore, should correspond.

The values of the major parameters such as dissolved solids, hardness, sulfate, and calcium, in samples taken from wells in sand and gravel in the discharge area are also similar to those reported from the carbonate aquifer. This is indicative of the influence of the quality of water in the carbonate aquifer in the discharge area on the quality of water in the overlying sand and gravel aquifer. Since the direction of flow in the regional discharge area is upward, water from the carbonate aquifer flows into and mixes with water in the overlying material.

Application of the Digital Model to the Silurian-Devonian Carbonate Aquifer

Another modified version of the digital computer program (Appendix B) developed by Prickett and Lonnquist (1971) was used in the model designed for the Silurian-Devonian carbonate aquifer. The main purpose of the model was to evaluate the ground-water availability in the carbonate aquifer and its potential use in future public and industrial development. The model was also used to predict the behavior and configuration of the potentiometric surface in response to different management conditions.
The carbonate aquifer has not been extensively developed in the Scioto River Basin and presently its use is restricted largely to areas with no alternative water supplies. However, the increase in industrialization and population growth, coupled with the migration of industries away from large population centers will undoubtedly lead to future large scale ground-water developments of the carbonate aquifer. The potential productivity of the aquifer, especially in the high yield area, will favor expansion of those municipalities near it, and the impact of their growth will undoubtedly influence the productivity and the water levels of the carbonate aquifer within the next 40 years.

The carbonate model consists of a rectangular area that covers about 80 percent (approximately 5,832 sq. mi.) of the Scioto River Basin (fig. 67). It extends from approximately the western limits of the carbonate subcrop in the basin eastward to include parts of the Muskingum, Hocking, Sandusky River basins. The aquifer properties were discretized by superimposing a square mesh finite difference grid over maps of the aquifer properites, such as the potentiometric surface, elevations of the top and bottom of the aquifer, transmissivity, permeability, and recharge rates.

The dimensions of the finite difference grid are defined by 28 columns, 55 rows and a total of 1,540 node locations, with a distance of 2 miles between adjacent nodes. The procedure used in the buried valley aquifer model for referencing the nodes was also followed (fig. 39).
Figure 67. Map showing the location and finite difference grid used in the carbonate aquifer.
The edges of the grid were chosen to coincide with hydrologic boundaries as seen on the potentiometric map (fig. 65). Since hydrologic boundaries are boundaries across which no flow occurs, they are represented as barrier boundaries in the model by assigning modified transmissivities and storage factors according to the vector volume concept. The southern limit of the grid was chosen to coincide with the limits of the aquifer outside of which it becomes tight (as determined from well logs and aquifer tests) and little or no flow occurs across that boundary.

Data Input to the Model

A series of maps that describe the aquifer properties and parameters involved in the model were prepared either from existing maps or by constructing new maps from available data and extrapolating where possible in areas that lacked data. The aquifer properties were discretized by superimposing a uniform finite difference grid over these maps and the value of each of the following parameters was recorded for each node in the grid:
1) The node coordinates \((i,j)\) that define the location of the node in the grid.
2) Transmissivity values \(T(i,j,1)\) in the increasing row direction and \(T(i,j,2)\) in the increasing column direction. These values are equal since we are using a uniform grid spacing. Different transmissivity values were calculated from aquifer conducted in southwest Ohio by the Ohio Division of Water. Due to the heterogeneous nature of the secondary openings in the aquifer, the transmissivity
values were calculated from aquifer tests varied greatly from one test area to another. In attempting to discretize the transmissivity, the modeled area was divided into blocks according to the range of transmissivity values within each block (fig. 68). The values within each block were then added and an average transmissivity was taken to represent the transmissivity for each of the nodes included with the particular block.

3) The coefficient of storage was determined from aquifer tests and the values ranged from 0.0001 to 0.004. An estimated average value of 0.003 was chosen to represent the storage factor for artesian conditions in the model, and 0.03 for water table conditions in the recharge area and in areas where dewatering of the aquifer occurs or is anticipated to occur as a result of pumping. The average values were further compared with values representing storage coefficients for similar aquifers under the same conditions in other areas. Rowland (1969) estimated an average storage coefficient of the Silurian-Devonian carbonate aquifer in the Upper Auglaize River Basin to be 0.0024. Walker et al. (1970) calculated the range in storage coefficient to be 0.001 to 0.0002, from tests conducted on the carbonate aquifer in Northwestern Ohio. Comparison of these values with the one averaged for the model aquifer shows a close correlation.

The coefficient of storage was converted to a storage factor (SF) using equation (V-3) in order to enter it into model calculations. The storage factor (SF1) representing leaky artesian conditions was $2.5 \times 10^{+6}$, and that representing the water-table conditions (SF2) was entered as $2.5 \times 10^{+7}$. 
Figure 68. Transmissivity blocks as entered into the carbonate model calculations.
4) The recharge factor (R) was calculated from the average rate of recharge as obtained from the hydrograph separation techniques and assumed to represent the average natural ground water recharge to the aquifer. As indicated earlier in the chapter, the average rate of recharge obtained is about 247,500 gpd/sq. mi. This value was transformed to a recharge factor using equation (V-4) and entered in the calculations of the model during the early stages of its development. When tested, the model failed to duplicate elevations of the potentiometric surface that existed from field data and the resulting simulated elevations were very distorted, even when the average recharge factor value was brought down by several magnitudes. The recharge factor (R) was zoned into blocks covering the grid area of the model. Figure 69 shows the final distribution of the recharge factors as entered into the model at the end of the verification stage.

The recharge rate to the carbonate aquifer is influenced by the vertical permeability and the thickness of till overlying the aquifer in the subcrop area. Values calculated for vertical permeability and the leakage coefficient (P'/m') from rates of recharge as shown in Table 18 and entered into the model in the Scioto River Basin, are correlative with values calculated from the steady state leaky artesian equation (Rowland, 1969, p. 43) in the Upper Auglaize River Basin, and those calculated from an aquifer test at Big Island near Marion (Norris, 1977, personal communication).
Figure 69. Distribution of recharge factors as entered into the model calculations.
TABLE 18: Comparison of Recharge Rates and Till Permeability in the Scioto River Basin and Upper Auglaize River Basin.

<table>
<thead>
<tr>
<th>Investigator</th>
<th>Location</th>
<th>Leakage Coefficient $p'/m'$ gpd/ft. $^3$</th>
<th>Permeability of Stream bed ($p'$) gpd/ft.$^2$</th>
<th>Recharge Rate gpd/sq. mi.</th>
<th>Remarks</th>
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<td>Rowland$^a$</td>
<td>Lima</td>
<td>0.000087</td>
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<td>0.0001</td>
<td>0.002</td>
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<td>Norris$^b$</td>
<td>Big Island</td>
<td>0.000002-0.000007</td>
<td>0.00075-0.025</td>
<td>597-1991</td>
<td>till thickness = 35 ft.</td>
</tr>
<tr>
<td></td>
<td>Marion</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ukayli</td>
<td>Scioto River</td>
<td>0.00014-0.00072</td>
<td>0.0012-0.036$^c$</td>
<td>4000-20,000</td>
<td>till thickness (assumed average) = 50 ft.</td>
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<td></td>
<td>Basin</td>
<td></td>
<td>ave=0.019</td>
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</table>

$^a$ Rowland (1969, p. 44).

$^b$ Norris, personal communication (1971).

$^c$ Values obtained from recharge rates as accepted in the digital calculations during verification.
The recharge rate in the Scioto River Basin as calculated from the recharge factors in Figure 69 varies from 4000 to 20,000 gpd/sq. mi.. Table 18 shows a comparison of the average recharge rate calculated from values entered in the model and those calculated by Rowland and Norris. Recharge rates representing different areas as shown in Table 18 vary greatly from those obtained by hydrograph separation techniques as described earlier in this chapter. They are more consistent and correlative, however, than the values calculated from the hydrograph separation techniques, and thus are considered more representative of the average rate of recharge to the carbonate aquifer than the previous values.

5) Elevations of the initial head (HO) (fig. 70) were discretized from the potentiometric surface contour map (fig. 65). The initial head values of nodes that occur between two intervening contours were evaluated to the nearest 5 feet.

6) The land or stream surface elevations (RH) were discretized from U.S. Geological Survey topographic maps (1:250,000 scale). Wherever possible, the elevations were evaluated to the nearest 5 feet.

7) Elevations of the top of the aquifer (CH) were discretized from the structure contour map of the carbonate aquifer (fig. 61). This map was compiled partly from the structure contour map of the Subsurface Silurian-Devonian "Big Lime" of Ohio (Owens, 1970) and partly from the bedrock surface map (fig. 62).

8) Elevations of bottom of the aquifer (BOT) were discretized from a structure contour map on the Lockport Dolomite (fig. 60). The map was compiled partly from Norris (1973) and partly from
Figure 70. Digitized contour map of the initial head as entered in the carbonate model.
Janssens (1968).

The bottom of the aquifer is taken as the top of the Lockport dolomite since most of the productive part of the aquifer exists in a zone (Newburg Zone) whose bottom is just a few feet above the Lockport Dolomite. However, part of the Lockport Dolomite that subcrops beneath the glacial till is included as part of the aquifer thickness, and is accounted for in the model.

Permeability is calculated within the model program. The program was revised to calculate the permeability at each node using the equation:

\[ P = \frac{T}{m} \]  \hspace{1cm} (V1-1)

where

\[ P = \text{permeability in gpd/ft}^2 \]
\[ T = \text{transmissivity in gpd/ft} \]
\[ m = \text{saturated thickness of the aquifer in feet} \]

The importance of this step is to make the transmissivity dependent on the saturated thickness of the aquifer, especially after the conversion from artesian conditions to water-table conditions.

Other data entered into the model are similar to those included in the buried valley aquifer model, as explained in the previous chapter. The default value card, however, is excluded from the program because data for each node in the model are included separately.

All data described previously were prepared and punched on computer cards for each node in the grid and entered as data into the aquifer simulation program (Appendix B).
In order to test the validity of the model and its ability to simulate the conditions under which the aquifer exists and its response to future changes, the model was first set to run without any inflow or outflow to the system. This was accomplished by eliminating natural recharge to the aquifer and the effects of pumping, and to carry the simulation for an exaggerated period of time (200,000 days). The purposes of these tests were to debug the model and to correct any errors in the data or in the program's operational sequence. If the program's operational sequence is correct, the initial water level elevations at each node location would drop or rise to a certain level that would represent the average level at the end of the simulation period.

After the model was tested, the natural recharge and historical pumpage were introduced into the program and the second phase was undertaken. This phase involved calibrating and achieving a hypothetical steady state situation in which the potentiometric elevations produced by simulation are comparable to those obtained by field measurements represented in Figure 65. The model had to undergo changes in many of the parameters such as recharge rate, transmissivity, and streambed leakage factors, before the simulated potentiometric levels started to match field measurements. A few trial runs, each having an exaggerated simulation period (time step) of 200,000 days were made, and the water level elevations calculated at the end of each run were punched on a set of computer cards that were used as input data for the plotting program (Appendix B) to produce a water level contour map representing the potentiometric
surface of the model at the end of the simulation period. Each map was then compared with the field map (fig. 65), differences were noted and the proper adjustments in the aquifer parameters were made before each consequent run. The process was repeated several times until the resulting levels produced by simulation correlated as close as possible with those obtained by field methods. Thus the hypothetical steady state is achieved; the model is calibrated and ready to be used in simulating water levels using historical and projected pumping rates.

Approximately 52 computer runs, involving many adjustments or combinations of adjustments of aquifer parameters, were made before the best fit of the steady state conditions was obtained. This involved about 14 minutes of the IBM 370/168 Central Processing Unit time.

Historical pumping in major areas of industrial developments in the area of the aquifer model were then reviewed, and past and future projected pumping rates were tabulated (Table 19), and punched on computer cards. The water levels in Figure 65 were discretized according to the finite difference grid, and the discretized water levels were used as initial heads in the first simulation period (fig. 70). The model was then allowed to simulate the water levels at the end of each pumping period according to Table 19 and the water levels produced were outputed as printouts and cards that are used in another program (Appendix B) to plot the water level contours. The water levels at the end of each simulation period were also stored on a high level index disk for use as initial
Table 19. Projection of total average supply of ground water from the carbonate aquifer in the modeled area.

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<td>22</td>
<td>30</td>
<td>1.05</td>
<td>1.37</td>
<td>1.54</td>
<td>1.92</td>
<td>2.16</td>
<td>2.31</td>
<td>2.31</td>
<td>COLUMBUS</td>
<td>FRANKLIN</td>
</tr>
<tr>
<td>22</td>
<td>31</td>
<td>1.16</td>
<td>1.37</td>
<td>1.54</td>
<td>1.92</td>
<td>2.16</td>
<td>2.31</td>
<td>2.31</td>
<td>COLUMBUS</td>
<td>FRANKLIN</td>
</tr>
</tbody>
</table>

TOTAL

|   | 22.46E+02 | 25.64E+02 | 26.46E+02 | 30.45E+02 | 31.24E+02 | 34.96E+02 | 41.26E+02 | 43.94E+02 | 205 |
heads in subsequent simulation periods.

The configuration of the potentiometric surface based on the total average daily demands at the end of the years 1979, 1989, 1999, 2009, 2019, and 2020 (figs. 71-76). The cones of depression around major pumping centers began to develop immediately after heavy pumping was initiated. The diameter of this cone is about 4 miles at the end of the first pumping period (1979) and it is about 8 miles at the end of 2029 (fig. 77). Drawdown at the center of the cone is about 89 feet at the end of the first pumping period and increases to 257 feet by the end of 2029.

The increase in the diameter of the cone and the amount of drawdown are attributed to the low to moderate transmissivity of the aquifer in the Columbus area and the lower rate of recharge. Drawdowns could be less than those obtained in this model if pumping were distributed over more pumping nodes, but geographical limitations on the locations of such nodes on the present grid dimensions limit the number of nodes that can be used without greatly distorting the actual positions of the wells under the present model conditions, average depth of wells in Columbus area will have to exceed 400 feet (or 270 feet below the initial potentiometric surface elevation) to avoid dewatering the aquifer below the critical pumping levels.

The influence of other major pumping centers on the regional trend of the potentiometric surface becomes more significant towards the end of 2009 (fig. 74) when the cones of depression around the pumping centers have deepened and widened due to the increase in
Figure 71. Potentiometric surface contour map showing water levels in the Silurian-Devonian carbonate aquifer at the end of 1979.
Figure 72. Potentiometric surface contour map showing water levels in the Silurian-Devonian carbonate aquifer at the end of 1989.
Figure 73. Potentiometric surface contour map showing water levels in the Silurian-Devonian carbonate aquifer at the end of 1999.
Figure 74. Potentiometric surface contour map showing water levels in the Silurian-Devonian carbonate aquifer at the end of 2009
Figure 75. Potentiometric surface contour map showing water levels in the Silurian-Devonian carbonate aquifer at the end of 2019.
Figure 76. Potentiometric surface contour map showing water levels in the Silurian-Devonian carbonate aquifer at the end of 2029.
Figure 77. Area of influence of the cone of depression at Columbus as a result of increased pumping rates from 1971 to 2030.
pumping rates from one simulation period to the other, as reflected from Table 19. Maximum drawdowns at Delaware, Marion and Kenton reach 58, 62, and 46 feet respectively. Water levels gradually fall below the top of the aquifer in most of the major pumping centers and conversion from artesian to water-table conditions is established in the vicinity of each cone of depression by as late as the year 2020 (fig. 78).

To verify the validity of the model in simulating the water levels correctly and to test its reliability in predicting, two aquifer test analysis were made from the simulated results, one at Columbus, because of significant pumping and spread of the cone of depression, and the other at Marion, where the aquifer is unconfined. Drawdowns at the end of 1979 (Columbus), and 2020 (Marion) pumping periods were calculated at distances of 2, 4, 6, and 8 miles from the pumping nodes. The amount of drawdown was plotted against the square of the distance from the pumped well to the measuring point on logarithmic paper, and the standard non-leaky artesian Theis type curve was superimposed over the graph (figs. 79,80). The transmissivity and storage coefficient of the aquifer were then calculated and compared with field data. Transmissivity at Columbus at the end of 1979 was 6,750 gpd/ft and the storage coefficient is 0.004 (fig. 79). Short term tests near Columbus conducted by the Ohio Division of Water provided transmissivity values that ranged from 2,900 to 11,000 gpd/ft (Ohio Div. of Water, Preliminary Reports, Central Ohio Project, Fred
Figure 78. Diagram illustrating the time of conversion from Artesian to water table conditions in the vicinity of major pumping centers listed in chronological order.
Figure 79. Distance-drawdown graph of the (15,31) at Columbus after 8 years of pumping (1971-1979).

<table>
<thead>
<tr>
<th>Node from P.W</th>
<th>Distance from P.W</th>
<th>Drawdown</th>
</tr>
</thead>
<tbody>
<tr>
<td>(14,31)</td>
<td>2 mi.</td>
<td>38'</td>
</tr>
<tr>
<td>(13,31)</td>
<td>4 mi.</td>
<td>18'</td>
</tr>
<tr>
<td>(12,31)</td>
<td>6 mi.</td>
<td>10'</td>
</tr>
<tr>
<td>(11,31)</td>
<td>8 mi.</td>
<td>3'</td>
</tr>
</tbody>
</table>

\[
T = \frac{114.6 \times Q \times w(u)}{S} = \frac{114.6 \times 765 \times 1.6745}{13} \text{ gdp/ft}
\]

\[
S = \frac{T u t}{2693 r^2} = \frac{6745 \times 0.1 \times 4.21 \times 10^6}{2693 	imes 2.5 \times 10^8} = 0.004
\]

MATCH POINT Δ

\[
u = 0.1
\]

\[
w(u) = 1
\]

\[
s = 13 \text{ ft}
\]

\[
t = 4.21 \times 10^6 \text{ min}
\]

\[
Q = 765 \text{ gpm}
\]

\[
r^2 = 2.5 \times 10^8
\]
Claire and Assoc., Columbus, 1971). Storage coefficients were not available but the reports assumed an average value of 0.0004.

The transmissivity at Marion was calculated from the model data 76,000 gpd/ft and the storage coefficient is 0.024 (fig. 80). Marion lies within the high yielding area where the transmissivity values, calculated from short term aquifer tests in an area between Prospect and Marion, ranged from 37,000 to 84,000 gpd/ft. The value calculated from the simulations lies well within this range.

The results of the above two tests agree with results obtained by field methods and are considered as another step in checking the validity of the model and its ability to correctly simulate future stresses on the aquifer.

Model results indicate that the aquifer is capable of supplying the projected needs of most industries and municipalities within the subcrop area with an adequate supply of water beyond the year 2030. Industrial developments that will require large quantities of water will have to be situated in the high yield area, where the supply is greater and larger withdrawal rates can be obtained without imposing considerable stresses on the aquifer yield.

Variations in the transmissivity of the aquifer ranged from 7,000 in the low-yield areas to 40,000 gpd/ft in the high-yield areas. The recharge rate varies from 4,000 to 20,000 gpd/sq. mi.. It is highest in the recharge area along the western border of the area. The average specific capacity of wells in the aquifer ranges
Figure 80. Distance-drawdown graph for pumping node (13,8) at Marion, Ohio after 58 years of pumping (1971-2029).

<table>
<thead>
<tr>
<th>Node</th>
<th>Distance from pump node</th>
<th>Drawdown</th>
</tr>
</thead>
<tbody>
<tr>
<td>(14,8)</td>
<td>2</td>
<td>22</td>
</tr>
<tr>
<td>(15,8)</td>
<td>4</td>
<td>12</td>
</tr>
<tr>
<td>(16,8)</td>
<td>6</td>
<td>7</td>
</tr>
<tr>
<td>(17,8)</td>
<td>8</td>
<td>6</td>
</tr>
</tbody>
</table>

\[
T = \frac{114.6 \times 4800 \times 1}{7.2} = 76,400 \text{ gpd/ft}
\]

\[
S = \frac{76,400 \times 0.1 \times 3.05 \times 10^7}{2.693 \times 3.6 \times 10^8} = 0.024
\]

MATCH POINT A

\[u = 0.1\]

\[w(u) = 1.0\]

\[s = 7.2 \text{ ft}\]

\[r^2 = 3.6 \times 10^8 \text{ ft}^2\]

\[t = 3.05 \times 07\]

\[Q = 4800 \text{ gpm}\]
from 2.6 gpm per foot of drawdown in areas outside the high-yield area to 30 in the high-yield area.
VII. GENERAL CONCLUSIONS

Ground water in the Scioto River basin is available from both consolidated and unconsolidated aquifers. Of these types, the buried-valley aquifer and the Silurian-Devonian carbonate aquifer are the most widespread and yield the greatest quantities of water. Both aquifers have great storage capacity. Variations between the two aquifer systems, however, are distinct.

The buried valley aquifer is relatively homogeneous, the permeability is high and yields are sufficiently large to attract industries and municipal developments, as in the case in Piketon, Chillicothe, Circleville, and Columbus. Total ground water available from storage per square mile area of the buried valley aquifer equals 2.4 billion gallons. The average rate of recharge to the aquifer is estimated to be 660,000 gpd/sq. mi.. Induced infiltration rate through the streambed varies from 0.5 to 3.1 gpd/sq. foot of streambed.

Since the quality of the ground water in the buried valley aquifer is fairly uniform, and far superior to that in streams, attention of large industries and future public supply developments will focus on this aquifer as the main source of needed water supply.

The Silurian-Devonian carbonate aquifer, although less dependable than the buried valley aquifer, yields appreciable amounts of ground
water that will fulfill the rising needs of public and industrial demands for at least the coming five decades before any major dewatering can take place. Extensive development of the carbonate aquifer will probably be restricted to the high-yield areas because they provide the largest and most dependable supplies.

Permeability of the aquifer is dependant on the size, frequency, and distribution of joints and solution channels and therefore the yield varies from one area to another. The presence of an especially high permeability zone, the "Newburg Zone", is related to major fluctuations of the water level during past geologic times. This resulted in the enlargement of solution openings and an increase in the vertical permeability, making this part of the aquifer the highest yielding zone within the carbonate sequence.

There is a direct relationship between the chemical quality of ground water in the carbonate aquifer and the regional flow system. Gradual changes occur in the quality of ground water as it moves from areas of recharge to areas of discharge. The increase in concentration of some selected constituents can be predicted as water moves along a flow line. Water in the carbonate aquifer is characterized by wide ranges in most of its chemical constituents. It is generally hard, but in most cases can be used for most purposes after softening and iron removal.

Upward leakage of water from the carbonate aquifer into glacial deposits, especially the buried valley aquifer, occurs in discharge areas where the carbonate aquifer is confined.
Modeling results indicate that the carbonate aquifer is capable of supplying greater quantities of water than required by industrial and municipal developments within the aquifer subcrop area.

The most significant result of this study is the development of the digital computer models to evaluate and plan ground-water development schemes in high yield areas such as the buried-valley and carbonate aquifer. The models also provide, at an insignificant cost to the developers, the ability to predict whether ground-water supplies within a given area would be enough to meet the predicted public and industrial demands.
APPENDIX A

This program utilizes various flow separation techniques to determine the relative contributions of ground water runoff and surface runoff to stream flow.

Subroutines are each summoned by a different call statement.

SUBROUTINE HDTRKP(X,D,IE)

SEE FORTRAN SSP FOR COMPLETE WRITEUP...

SUBROUTINE VFLVDAtSTA,DRA,DIS,N,NCNT)

THIS SUBROUTINE CALCULATES A FLOW DURATION CURVE FOR THE YEAR AND PLATTS IT OUT ON THE VERSATEC PLOTTER.

This section creates the basic graph outline.

END
CALL SYMBOL(-1.0,1.0,0.14,'DISCHARGE IN CFS/MI 2',90.0,21)
DO 11 I = 1,46
YPLOT=(ALOG10(AY(I)))+2.25)+6.73
IF(AY(I).LT.0.01) GO TO 31
IF(AY(I).LT.0.10) GO TO 52
IF(AY(I).LT.1.00) GO TO 53
CALL NUMBER(-0.5,YPLOT,0.07,AY(I),0.0,-1)
GO TO 30
51 CALL NUMBER(-0.5,YPLOT,0.07,AY(I),0.0,3)
GO TO 30
52 CALL NUMBER(-0.5,YPLOT,0.07,AY(I),0.0,0,2)
GO TO 30
53 CALL NUMBER(-0.5,YPLOT,0.07,AY(I),0.0,1)
30 CALL PLOT(0.3,YPLOT,3)
CALL PLOT(9.6,YPLOT,2)
111 CONTINUE
DO 112 I = 1,23
CALL NUMBER(AK(I),X(0),IER)
XPLOT=X(0)+0.07
IF(AK(I).LT.0.10) GO TO 90
CALL NUMBER(YPLOT,0.07,AY(I),0.0,-1)
GO TO 90
80 CALL NUMBER(YPLOT,0.07,-0.4,0.07,(100.*AX(I)),0.0,1)
90 CALL PLOT(YPLOT,11.3,2)
112 CONTINUE
55 MISS=0
HOFLOW=0
PHAI=0
KFO=0
C DATA SCAN FOR POSSIBLE MISSING DATA OR NO FLOW (<0.01)
DO 60 I = 1,N
IF(DISC(I).LT.0.0) GO TO 61
IF(DISC(I).LT.0.09) NOFLOW=NOFLOW+1
MISS=MISS+1
61 MISS=MISS+1
60 CONTINUE
IF(I.F.0) MISS+1
I=1,N-1
DO I = 1,N
IF(LEB(J).LE.DBS(J+1)) GO TO 1
2 CONTINUE
1 CONTINUE
CONTINUE
DAYS=FLOAT(KNN)
DO B = 1,NNN
APER(I)=(KNN-B+1)*100/KNN
DISDEN(I)=DISC(I)/DRA
3 CONTINUE
C WRITES OUT FLOW DURATION TABLE
WRITE (6,13) STA,DRA
DO 11 I = 1,NNN,0
IF(ERN.EQ.365,AUD.I.EQ.361) GO TO 12
11=1,7
WRITE (6,21) (APER(N),DISDEN(N),N=I,11)
11 CONTINUE
GO TO 29
12 WRITE (6,22) (APER(N),DISDEN(N),N=361,365)
DELTA1=DISDEN(37)-DISDEN(36)
DELTA2=DISDEN(329)-DISDEN(323)
DELTA3=DISDEN(274)-DISDEN(273)
DELTA4=DISDEN(274)-DISDEN(273)
C00=DISDEN(36)+DELTA1/*2)
C10=DISDEN(329)+DELTA2*/2)
C70=DISDEN(274)+DELTA3*/2)
C20=DISDEN(274)+DELTA4*/2)
IF(C00.LE.9.0) GO TO 24
IF(C70.LE.9.0) GO TO 24
GO TO 29
24 WRITE(6,26)
   IF (DAS.LT.0.001) DAS=0.001
   YY=(ALOG10(DAS)-2.25)+6.73
25 IF (I.EQ.1) 10=8
   XPER=AVER(1)/100.
   IF (XPER.LT.1.0) XPER=.999
   IF (XPER.LE.0.601) XPER=.901
   CALL NDTRI(XPER, X.D, IER)
   IF (IER.NE.0) WRITE(6,14)
   XF=0.5*X+.81
   IF (I.EQ.1) CALL NUMBER((XX+0.02), YY, 0.07, RECNT, 0.0, -1)
   CALL PLOT(XX, YY, IC)
   IF (XPER.GT.0.99) CALL SYMBOL(XX, YY, 0.035, 2, 0.0, -1)
10 CONTINUE
   IF (RCHT.GT.0.01) CALL NUMBER((XX-0.07), YY, 0.07, RECNT, 0.0, -1)
   CALL SYMBOL(XX, 0.0, 13, 0.0, -1)
   IF (IFLOW.GT.0) GO TO 100
   GO TO 103
C INDICATES POSITION OF LAST NO FLOW WITH AN ASTERISK IF NOFLOW>0)
100 DO 102 I=1,MAX
   IF (DAS.I.LT.0.01) GO TO 102
   IF (DAS.LT.0.01.AND.DISC(I+1).GT.0.01) GO TO 1001
   GO TO 102
101 IF (XPER.GT.1.0) XPER/.100.
   IF (XPER.LE.0.601) XPER/.901
   CALL NDTRI(XPER, X.D, IER)
   XF=0.5*X+1.5
   CALL SYMBOL(XX, -0.10, 0.14, 11, 0.0, -1)
102 CONTINUE
C WRITES SUMMARY INFORMATION AT TOP OF PAGE FOR EACH CURVE
103 IF (RCHT.GT.0.01) CALL NUMBER((XX+0.8), YY, 0.07, RECNT, 0.0, -1)
   CALL SYMBOL(XX, 12.25, 0.07, DAYS, 0.0, -1)
   CALL SYMBOL(XX, 11.73, 0.07, 'DAYS PLOTTED', 0.0, 11)
   IF (MISS.LE.0.0) GO TO 71
   IF (IFLOW.GT.0) GO TO 70
   CALL SYMBOL(XX, 11.8, 0.07, 'DAYS NO FLOW(*)', 0.0, 19)
   CALL SYMBOL(XX, 11.30, 0.07, 'DAYS MISSING DATA', 0.0, 18)
70 CALL NUMBER((XX+0.8), 12.3, 0.07, 'DAYS NO FLOW(*)', 0.0, 16)
71 IF (MISS.LE.0.0) CALL NUMBER((XX+0.8), 12.0, 0.07, 'DAYS MISSING DATA', 0.0, 13)
   CALL SYMBOL((XX+0.8), 12.25, 0.07, 'DAYS NO FLOW(*)', 0.0, 16)
72 IF (IFLOW.GT.0) CALL NUMBER((XX+0.8), 12.0, 0.07, 'DAYS NO FLOW(*)', 0.0, 16)
77 RETURN
13 FORMAT (I1, 'FLOW-DURATION CURVE FOR', 1X, 'SQ.MI')
15 FORMAT ('O', 'A R R N I N G ERROR IN NDTRI')
21 FORMAT (4X, 10(F6.2, 1X, F6.2))
```
226 FORMAT (' ', A6X, 5(F6.2, 1X, F6.3, 2X))
230 FORMAT (' ', A7X, 'THE RATIO (Q10/Q90)**1/2 = ', F8.2, 10X, '(Q25/Q75)**1/2 = ', F8.2, 10X, 'C123/Q75')
260 FORMAT (' ', 'Q90 IS LESS THAN OR EQUAL TO ZERO, RATIO IS MEANINGLESS.')
270 FORMAT (' ', 'Q75 IS LESS THAN OR EQUAL TO ZERO, RATIO IS MEANINGLESS.')
END
```
DIMENSION DIS(365), CDIS(365), TECH(6), STA(16)
REAL IDIS, LMIN, LMAX
NNISS=0
NNN=0
2 READ(5,9,END=4) YR, STA, DRA, N, (DIS(I), I=1,N)
3 NN=NN+1
4 DO 1 I=1,N
5 IF(DIS(I) .LT. 0.0) NMISS=NMISS+1
1 CONTINUE
6 RINTR=DRA**0.2
7 RINTR=RINTR*2
8 IF(RINTR.LE.6.0) INTR=3
9 IF(RINTR.LE.6.0 .AND. RINTR.GT.4.0) INTR=5
10 IF(RINTR.LE.10.0 .AND. RINTR.GT.6.0) INTR=7
11 IF(RINTR.GT.10.0) INTR=11

THE SUBROUTINES ARE EACH SUMMONED BY A DIFFERENT CALL STATEMENT:
CALL LOCMIN (DIS,N,CDIS,INTR,TECH)
CALL VERSAP(STA, DRA,N,DIS,CDIS,TECH)

NNISS=0
GO TO 2

FORMAT('0',60(' '*))
FORMAT('1',60(' '*))
FORMAT('0',16A4,1X,F7.1,10X,'19','12')
FORMAT('0',40X, 'MISSING DATA: ',13,'DAYS')
FORMAT('12',16A4,1X,F10.2/110/10.0)
END

SUBROUTINE FXINTR (DIS,N,CDIS,INT,TECH)
DIMENSION DIS(N), CDIS(N), TITLE(20), TECH(6), SECH(6)
DATA SECH(2), SECH(3), SECH(4), SECH(5), SECH(6) / 'DAY','FIXE','D IN
*','TERV','AL'/
DATA 'THREE,FIVE,SEVEN,NINE,ELEVEN' 3 ' 5 ' 7 ' 9 ' 11
99 DO 99 J=2,6
89 TECH(J)=SECH(J)
90 IF(INT.EQ. 3) TECH(1)=THREE
91 IF(INT.EQ. 5) TECH(1)=FIVE
92 IF(INT.EQ. 7) TECH(1)=SEVEN
93 IF(INT.EQ. 9) TECH(1)=NINE
94 IF(INT.EQ.11) TECH(1)=ELEVEN
WRITE (6,7) INT
K=INT
DO 3 I=1,K
1 IF(DIS(I).LT.0.0) GO TO 3
2 PMIN=100000.
3 L1=((I-1)*INT)+1
4 L2=INT
5 DO 1 J=L1,L2
6 IF(DIS(J).LT.PMIN) PMIN=DIS(J)
1 CONTINUE
2 DO 2 J=L1,L2
3 GDIS(J)=PMIN
2 CONTINUE
3 CONTINUE
4 NI=(K*INT)+1
5 IF(K*INT.EQ.N) GO TO 6
6 PMIN=100000.
7 DO 4 J=1,N
8 IF(DIS(J).LT.0.0) GO TO 4
9 IF(DIS(J).LT.PMIN) PMIN=DIS(J)
4 CONTINUE
5 DO 5 J=1,N
6 GDIS(J)=PMIN
5 CONTINUE
6 RETURN
7 WRITE ('0','FIXED INTERVAL, INTERVAL=',13,'DAYS')
SUBROUTINE SLINTR(DIS, DG, INTRVL, TECH)
DIMENSION DIS(10),titulo(20), TEC(6), SECH(6)
DATA SECH(2), SECH(3), SECH(4), SECH(5), SECH(6) / 'DAY', 'SLID', 'IMG
* ', 'INT', 'INRVL', */
DATA THREE, FIVE, SEVEN, NINE, ELEVEN/ '3', '5', '7', '9', '11
* */
DO 99 J=2, 6
99 TECH(J)=SECH(J)
INT=INTRVL
IF (INT.EQ. 3) TECH(1)=THREE
IF (INT.EQ. 5) TECH(1)=FIVE
IF (INT.EQ. 7) TECH(1)=SEVEN
IF (INT.EQ. 9) TECH(1)=NINE
IF (INT.EQ. 11) TECH(1)=ELEVEN
WRITE (6,9) INTRVL, INT*(INTRVL-1)/2
DO 8 I=1, K
IF (DIS(I).LT.6.0) GO TO 8
IF (I-(INT+1)) 4, 1, 1
1 IF((365-I)-(INT+1)) 6, 2, 2
2 PMIN=100000.
K1=1+INT
K2=I+INT
DO 3 J=K1, K2
IF (DIS(J).LT.PMIN) PHIN=DIS(J)
CONTINUE
3 CONTINUE
4 PMIN=PMIN
K1=I+INT
K2=I+INT
DO 5 J=1, K2
IF (DIS(J).LT.PMIN) PHIN=DIS(J)
CONTINUE
5 CONTINUE
6 PMIN=PMIN
K1=I+INT
K2=I+INT
DO 7 J=1, 365
IF (DIS(J).LT.PMIN) PHIN=DIS(J)
CONTINUE
7 CONTINUE
8 PMIN=PMIN
RETURN
9 FORMAT ('0', 'SLIDING INTERVAL, INTERVAL=', I3, 'DAYS')
END
SUBROUTINE LOCIN(DIS, N, CDIS, INTRVL, TECH
DIMENSION DIS(N), CDIS(N), IPOINT(400), TEC(6), SECH(6)
DATA SECH(2), SECH(3), SECH(4), SECH(5), SECH(6) / 'DAY', 'LOCA', 'L HI',
I* 'NIMA' , */
DATA THREE, FIVE, SEVEN, NINE, ELEVEN/ '3', '5', '7', '9', '11
*/
DO 1 J=2, 6
1 TECH(J)=SECH(J)
INT=INTRVL
IF (INT.EQ. 3) TECH(1)=THREE
IF (INT.EQ. 5) TECH(1)=FIVE
IF (INT.EQ. 7) TECH(1)=SEVEN
IF (INT.EQ. 9) TECH(1)=NINE
IF (INT.EQ. 11) TECH(1)=ELEVEN
WRITE (6,24) INTRVL, NUMPT=0
1 IF (INTRVL.EQ. 9) GO TO 2
IF (INTRVL.EQ. 5) GO TO 5
IF (INTRVL.EQ. 7) GO TO 8
IF (INTRVL.EQ. 9) GO TO 11
IF (INTRVL.EQ. 11) GO TO 14
2 L=365-I
4 I=2, L
IF (DIS(J).LE.DSS(I+1).AND.DSS(I).LE.DSS(I-1)) GO TO 3
GO TO 4
3 NUMPT=NUMPT+1
IPOINT(NUMPT)=I
CONTINUE
GO TO 17
5 \[ L = 365 - 2 \]
6 \[ \text{DO 7 } I = 3, L \]
7 \[ \text{IF(DSS(1).LE.DSS(I+1).AND.DSS(1).LE.DSS(I-1).AND.DSS(1).LE.DSS(I+2)} \]
8 \[ \text{AND.DSS(1).LE.DSS(I-2)) GO TO 6} \]
9 \[ \text{GO TO 7} \]
10 \[ \text{NUMPT=NUMPT+1} \]
11 \[ \text{IPOINT(I) = I} \]
12 \[ \text{CONTINUE} \]
13 \[ \text{GO TO 17} \]
14 \[ \text{L = 365 - 3} \]
15 \[ \text{DO 10 } I = 4, L \]
16 \[ \text{IF(DSS(1).LE.DSS(I+1).AND.DSS(1).LE.DSS(I+2).AND.DSS(1).LE.DSS(I+3)} \]
17 \[ \text{AND.DSS(1).LE.DSS(I-1).AND.DSS(1).LE.DSS(I-2).AND.DSS(1).LE.DSS(I)} \]
18 \[ \text{AND.DSS(I).LE.DSS(I-3).GO TO 9} \]
19 \[ \text{GO TO 10} \]
20 \[ \text{NUMPT=NUMPT+1} \]
21 \[ \text{IPOINT(I) = I} \]
22 \[ \text{CONTINUE} \]
23 \[ \text{GO TO 17} \]
24 \[ \text{L = 365 - 4} \]
25 \[ \text{DO 13 } I = 5, L \]
26 \[ \text{IF(DSS(1).LE.DSS(I+1).AND.DSS(1).LE.DSS(I+2).AND.DSS(1).LE.DSS(I+3)} \]
27 \[ \text{AND.DSS(1).LE.DSS(I-1).AND.DSS(1).LE.DSS(I-2).AND.DSS(1).LE.DSS(I)} \]
28 \[ \text{AND.DSS(I).LE.DSS(I-3).AND.DSS(I).LE.DSS(1-4).AND.DSS(1).LE.DSS(1+4)) GO TO 12} \]
29 \[ \text{GO TO 13} \]
30 \[ \text{NUMPT=NUMPT+1} \]
31 \[ \text{IPOINT(I) = I} \]
32 \[ \text{CONTINUE} \]
33 \[ \text{GO TO 17} \]
34 \[ \text{L = 365 - 5} \]
35 \[ \text{DO 16 } I = 6, L \]
36 \[ \text{IF(DSS(1).LE.DSS(I+1).AND.DSS(1).LE.DSS(I+2).AND.DSS(1).LE.DSS(I+3)} \]
37 \[ \text{AND.DSS(1).LE.DSS(I-1).AND.DSS(1).LE.DSS(I-2).AND.DSS(1).LE.DSS(I)} \]
38 \[ \text{AND.DSS(I).LE.DSS(I-3).AND.DSS(I).LE.DSS(1-4).AND.DSS(1).LE.DSS(1+4)) GO TO 15} \]
39 \[ \text{GO TO 16} \]
40 \[ \text{NUMPT=NUMPT+1} \]
41 \[ \text{IPOINT(I) = I} \]
42 \[ \text{CONTINUE} \]
43 \[ \text{GO TO 17} \]
44 \[ \text{J = NUMPT - 1} \]
45 \[ \text{L = IPOINT(NUMPT)} \]
46 \[ \text{DO 18 } I = 1, J \]
47 \[ \text{CDIS(I) = DSS(I)} \]
48 \[ \text{CONTINUE} \]
49 \[ \text{DO 19 } J = L, 365 \]
50 \[ \text{CDIS(I) = DSS(I-L)} \]
51 \[ \text{CONTINUE} \]
52 \[ \text{DO 21 } I = 1, K \]
53 \[ \text{IP1 = IPOINT(I)} \]
54 \[ \text{IP2 = IPOINT(I+1)} \]
55 \[ \text{CDIS(IP1) = DSS(IP1)} \]
56 \[ \text{CDIS(IP2) = DSS(IP2)} \]
57 \[ \text{ISTART = IP1} \]
58 \[ \text{IEND = IP2} \]
59 \[ \text{DO 20 } J = ISTART, IEND \]
60 \[ \text{X = J - IP1} \]
61 \[ \text{Y = IP2 - IP1} \]
62 \[ \text{IF(CDIS(IP1).EQ.0.0) CDIS(IP1) = 0.01} \]
63 \[ \text{IF(CDIS(IP2).EQ.0.0) CDIS(IP2) = 0.01} \]
64 \[ \text{CDIS(J) = 10.**((X/Y)**(ALOG10(CDIS(IP2))-ALOG10(CDIS(IP1))))+ALOG10(CDIS(IP1)))} \]
65 \[ \text{CONTINUE} \]
66 \[ \text{DO 22 } IJK = 1, 365 \]
67 \[ \text{IF(CDIS(IJK).GT.DSS(IJK)) CDIS(IJK) = DSS(IJK)} \]
68 \[ \text{CONTINUE} \]
69 \[ \text{RETURN} \]
70 \[ \text{RETURN} \]
71 \[ \text{FORMAT ('O','LOCAL MINIMA. INTERVAL=',13,'DAYS')} \]
72 \[ \text{END} \]

SUBROUTINE VERSAP (STA, DRA, N, DIS, CDIS, TECII)
C ***** THIS SUBROUTINE PRODUCES OUTPUT ON THE VERSATEC PLOTTER *****
DIMENSION CDIS(365), DIS(365), TITLE(16), TECII(6), BCDC3), STA(16)
DIMENSION TOTAL(4), GROUND(6), GROUND(5), RECHAR(4), CF(5), GPD(3)
REAL IDIS,LHN,LNAX
DATA BCD, 'LOG', 'H13', 'S'
DATA TOTAL, 'TOTAL', 'L DI', 'SCHA', 'RGE'
DATA GROUND, 'GROU', 'ND W', 'ATER', 'RNM', 'OFF'
DATA GROUND, 'GROU', 'ND W', 'ATER', 'AS'
DATA RECHAR, 'RECH', 'ARGE', 'RAT', 'E'
DATA CF, 'CF O', 'R', 'IN', 'CHES'
DATA GPD, 'GPD', '/S', '11', '
C ****** CALCULATE SUMMARY INFORMATION ******
NMISS = 0
DAYS = 0.0
RMAX = 0.0
RMIN = 100000000.
TOTAL = 0.0
DO 10 I = 1,N
IF(DIS(I).LT.0.0) NMISS = NMISS + 1
IF(DIS(I).LT.0.0) GO TO 10
DAYS = DAYS + 1.0
IF(DIS(I).GT.RMAX) RMAX = DIS(I)
IF(DIS(I).LT.RMIN) RMIN = DIS(I)
TOTALDI = TOTALDI + DIS(I)
TOTALCW = TOTALCW + CDIS(I)
10 CONTINUE
TOQUAN = 66400. * TOTALDI
TOCH = 66400. * TOTALCW
TOTALD = 0.03719 * (TOTALDI / DRA)
TOTALQ = 0.03719 * (TOTALCW / DRA)
RDIH = TOQUAN / DRA
TDGH = TOCH / DRA
PERCEN = (TOQUAN / TOCH) * 100.
RECH = TOCH / 7.40 / DAYS
RECE = FLOAT(RECH / 1000.)
RECH = RECE*1000.0
RECE = RECE / 1000.0
XQUAN = ALOG10(TOQUAN)
XQUCH = XQUAN
YQUAN = 2.23 + (ALOG10(CDIS(I)) / 1.6)
CALL PLOT (X, Y, IC)
11 CONTINUE
CALL PLOT (0.0, 0.0, 3)
THIS SECTION PLOTS THE GROUNDWATER-
SURFACE WATER CURVE
DO 2 I = 1,N
IC = 2
IF(1.EQ.1) IC = 3
X = 1.025 + 1.0*DIS(I)
IF(DIS(I).EQ.0.00) DIS(I) = 0.01
IF(DIS(I).LT.0.01) IC = 3
Y = 2.25 + (ALOG10(DIS(I))/1.6)
CALL PLOT (X, Y, IC)
2 CONTINUE
CALL AXSN (10, 150, 1, 'DISCHARGE', -9, -5, 90, -2, 1.6, 0.625, 0, 1)
CALL AXSN (10, 3, 1.345, BCD, -12, 5, 90, -3, 1.6, 0.625, 0, 1)
CALL AXSN (10, 000, 6, 000, 'DAYS', -9, 125, 180, 0, 40, 0.75, 2, -1)
IF(ARMN.GT.1.0) GO TO 3
IF(ARMN.LT.10000) GO TO 4
IF(ARMN.LT.100000 .AND. ARMN.GT.0.1) GO TO 5
GO TO 6
3 CALL SYMBOL (2.0, 1.25, 0.14, RECHAR, 0.0, 0.16)
CALL NUMBER (4.5, 1.23, 0.14, RECHG, 0.0, -1)
CALL SYMBOL (6.5, 1.23, 0.14, GF0, 0.0, 12)
CALL SYMBOL (2.0, 1.50, 0.14, GROUND, 0.0, 20)
CALL NUMBER (8.1, 1.50, 0.14, PERCEN, 0.0, 1)
CALL SYMBOL (2.0, 1.75, 0.14, GROUND, 0.0, 24)
CALL NUMBER (8.1, 1.75, 0.14, TOQUCW, 0.0, 3)
CALL SYMBOL (6.05, 1.75, 0.14, CF, 0.0, 69, 0.0, -1)
CALL NUMBER(6.1, 1.75, 0.14, TOQUGWI, 0.0, 0.2)
CALL SYMBOL (6.7, 2.00, 0.14, CF, 0.0, 20)
CALL NUMBER (7.6, 2.00, 0.14, TOTAL, 0.0, 0.2)
GO TO 6
4 CALL SYMBOL (2.0, 4.75, 0.14, RECHAR, 0.0, 0.16)
CALL NUMBER (4.5, 4.75, 0.14, RECHG, 0.0, -1)
CALL SYMBOL (6.5, 4.75, 0.14, GF0, 0.0, 12)
CALL SYMBOL (2.0, 5.00, 0.14, GROUND, 0.0, 20)
CALL NUMBER (8.1, 5.00, 0.14, PERCEN, 0.0, 1)
CALL SYMBOL (2.0, 5.25, 0.14, GROUND, 0.0, 24)
CALL NUMBER (8.1, 5.25, 0.14, TOQUCW, 0.0, 3)
CALL SYMBOL (6.05, 5.25, 0.14, CF, 0.0, 69, 0.0, -1)
CALL NUMBER(6.1, 5.25, 0.14, TOQUGWI, 0.0, 0.2)
CALL SYMBOL (6.7, 5.50, 0.14, CF, 0.0, 20)
CALL NUMBER (7.6, 5.50, 0.14, TOTAL, 0.0, 0.2)
GO TO 6
5 CALL SYMBOL (2.0, 1.25, 0.14, RECHAR, 0.0, 0.16)
CALL NUMBER (4.5, 1.23, 0.14, RECHG, 0.0, -1)
CALL SYMBOL (6.5, 1.23, 0.14, GF0, 0.0, 12)
CALL SYMBOL (2.0, 1.50, 0.14, GROUND, 0.0, 20)
CALL NUMBER (8.1, 1.50, 0.14, PERCEN, 0.0, 1)
CALL SYMBOL (2.0, 1.75, 0.14, GROUND, 0.0, 24)
CALL NUMBER (8.1, 1.75, 0.14, TOQUCW, 0.0, 3)
CALL SYMBOL (6.05, 1.75, 0.14, CF, 0.0, 69, 0.0, -1)
CALL NUMBER(6.1, 1.75, 0.14, TOQUGWI, 0.0, 0.2)
CALL SYMBOL (6.7, 1.75, 0.14, CF, 0.0, 20)
CALL NUMBER (7.6, 2.00, 0.14, TOTAL, 0.0, 0.2)
BEGIN
CALL SYMBOL (1.000, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (1.107, 6.050, 0.07, 'OCT', 0.0, 0.3)
CALL SYMBOL (1.775, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (2.525, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (3.500, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (5.617, 6.050, 0.07, 'FEB', 0.0, 0.3)
CALL SYMBOL (4.075, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (4.775, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (5.550, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (5.867, 6.030, 0.07, 'APR', 0.0, 0.3)
CALL SYMBOL (6.590, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (7.075, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (7.625, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (8.142, 6.050, 0.07, 'JUL', 0.0, 0.3)
CALL SYMBOL (9.600, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (9.375, 6.000, 0.07, 90, 0.0, -1)
CALL SYMBOL (9.692,6.000,0.07,'SEP',0.0,3)
CALL SYMBOL (10.123,6.000,0.07,90,0.0,-1)
CALL SYMBOL (1.6.5,0.14,TECH,0.,24)
CALL SYMBOL (1.7,0.14,TITLE,0.0,64)
CALL SYMBOL (9.25,6.5,0.14,TECH,0.0,24)
CALL SYMBOL (10.25,6.5,0.14,'SMI',0.0,6)
IF(NMISS.GT.0) GO TO 99
CALL PLOT
RETURN

99 CALL NUMBER(6.5,6.5,0.14,NMISS,0.0,-1)
CALL SYMBOL(7.0,6.5,0.14,'MISSING DAYS',0.0,13)
CALL PLOT
RETURN
END
APPENDIX B

PROGRAM 1

********* PIKE TON *********
COMPUTER PROGRAM LISTING FOR MODEL AQUIFER
MODIFIED AFTER
ILLINOIS STATE WATER SURVEY
BASIC AQUIFER SIMULATION PROGRAM WITH
LEAKY ARTESIAN, INDUCED INFILTRATION,
EVAPOTRANSPIRATION, ARTESIAN TO WATER-
TABLE STORAGE COEFFICIENT CONVERSION,
WATER TABLE CONDITIONS, AND VARIABLE NET
WITHDRAWAL RATES

BY PRICKETT AND LONNQUIST

DEFINITION OF VARIABLES

H0(I,J)-----HEADS AT START OF TIME
INCREMENT (I,J)

H1(I,J)-----HEADS AT END OF TIME
INCREMENT (FT)

SF1(I,J)----STORAGE FACTOR FOR
ARTESIAN CONDITIONS (GAL)

SF2(I,J)----STORAGE FACTOR FOR WATER-
TABLE CONDITIONS (GAL/FT)

Q(I,J)-----CONSTANT WITHDRAWAL
RATES (GPD)

PERM(I,J,1)---HYDRAULIC CONDUCTIVITY OF
AQUIFER BETWEEN I,J AND I,J+1
(GAL/DAY/SQ FT)

PERM(I,J,2)---HYDRAULIC CONDUCTIVITY OF
AQUIFER BETWEEN I,J AND I+1,J
(GAL/DAY/SQ FT)

T(I,J,1)---AQUIFER TRANSMISSIVITY
BETWEEN I,J AND I,J+1

T(I,J,2)---AQUIFER TRANSMISSIVITY
BETWEEN I,J AND I+1,J

DOT(I,J)---ELEVATION OF BOTTOM OF
AQUIFER (FT)

R(I,J)-----RECHARGE FACTOR
(GAL/DAY/FT)

RH(I,J)----ELEVATION OF LAND OR
STREAM SURFACE (FT)

RDB(I,J)---ELEVATION OF BOTTOM OF
STREAMBED OR ELEVATION
BELOW WHICH EVAPOTRANS-
PERSION CEASES (FT)

CH(I,J)---ELEVATION OF TOP OF
AQUIFER (FT)

B----------PEACEMAN-RACHFORD
B ARRAY

C----------PEACEMAN-RACHFORD
C ARRAY

AA, BB, CC, DD---COEFFICIENTS IN WATER
BALANCE EQUATIONS

NR----------NO. OF ROWS IN MODEL

NC----------NO. OF COLUMNS IN MODEL

NSTEPS-----NO. OF TIME INCREMENTS

DELTA------TIME INCREMENTS (DAYS)

HIII, S1, QQ, TT---EXAMPLE MODEL

DEFAULT VALUES

I----------MODEL COLUMN NUMBER
J----------MODEL ROW NUMBER
IP----------I COORDINATE OF PUMP

JP----------J COORDINATE OF PUMP

P(K, KC)-----PUMPING RATE AT WELL
K AT RATE KC (GPD)
DIMENSION H(50,50), H0C50,50), 1SFK50,50), 0(50,50), T(50,50,2).
2D(50), C(50, 0(50,50), DL(50,50).
3, IP(100), JP(100), P(100,100).
4, NRT(50,50,2), B(50,50), SF(2(59,50).
5, R(50,50), R(50,50), N(50,50), R(50,50).
6, NODE(50,2), HEAD(50,100), LINE(100), ALINE(100).
7, TIME(50), IDME(50,50).

DIMENSION DIFC23,49).

TURN OFF UNDERFLOW TRAP

CALL ERRSET(208,296,-1,1).

DEFINE INPUT AND OUTPUT DEVICE NUMBERS

INTEGER OUT
IN=5
OUT=6.

READ PARAMETER CARD AND
DEFAULT VALUE CARD
READ(IN,10) NSTEPS, DELTA, ERROR,
INC, NR, TT, S1, IH, QD, RR, RLD, S2, CCH, PP, BOTT
FORMAT(16,2F6.0/216,F6.0)

READ NODE DATA FOR TIME-
WATER LEVEL CURVES

READ(IN,14) NN, AL, ((NODE(I,J), J=1,2), I=1,NN)
FORMAT(16,F6.0,10(14,13)/11(14,13)/17(14,13))

READ PUMP PARAMETER CARD

READ(IN,15) NP, NSP, NRT
FORMAT(16)

WRITE(OUT,304)
FORMAT(112('X'))
WRITE(OUT,303) NSTEPS, DELTA, ERROR, INC, NR, TT, S1, HQ, QD,
RR, RLD, S2, CCH, PP, BOTT
FORMAT(16,F6.0,10(14,13)/11(14,13)/17(14,13))
WRITE(OUT,303) NP, NSP, NRT
FORMAT(16,F6.0,10(14,13)/11(14,13)/17(14,13))
WRITE(OUT,304)

READ PUMPING SCHEDULES

DO 16 I=1, NP
READ(IN,17) IP(I), JP(I),
1(P(I,K), K=1,NRT)
FORMAT(1356,6F6.0/6F6.0)
DO 18 I=1, NP
WRITE(6,19) IP(I), JP(I), (P(I,K), K=1,NRT)
FILL ARRAYS WITH DEFAULT VALUES

DO 20 I=1,NC
DO 20 J=1,NR
T(I,J,1)=TT
T(I,J,2)=TT
PERM(I,J,1)=PP
PERM(I,J,2)=PP
SF1(I,J)=S1
SF2(I,J)=S2
HI(I,J)=HH
HO(I,J)=HH
HI(I,J)=HR
RH(I,J)=RR
RD(I,J)=RR
CH(I,J)=CH
HOT(I,J)=BOTT

Q(I,J)=QQ

READ NODE CARDS

READ(IN,40,END=50)I,J,T(I,J,1),T(I,J,2)
SF1(I,J),HI(I,J),HI(I,J),HO(I,J),HH(I,J)
SF2(I,J),HH(I,J),CH(I,J),PERM(I,J,1),
PERM(I,J,2),BOTT(I,J)

FORMAT(213,2F6.0,2F4.0,9F6.0)
GO TO 30

START OF SIMULATION

TIME=0
READ(IN,16,END=90)I,J,T(I,J,1),T(I,J,2)
SF1(I,J),CH(I,J),HI(I,J),HI(I,J),HO(I,J),HH(I,J)
SF2(I,J),HH(I,J),CH(I,J)
HOT(I,J)=BOTT(I,J)
FORMAT(10X,6(F5.2))
DEL*DELTA
KG=1
DO 320 ISTEP=1,NSTEPS

ENTER PUMPAGE SCHEDULES

Z=(ISTEP-1.0)/NSP+1.0
IF(Z-KC-.00 .46,49
DO 47 K=1,NP
I=IP(K)
J=JP(K)
Q(I,J)=P(K,KC)
DELTA=DEL
KG=KG+1

TRANSMISSIVITY CALCULATED AT
BEGINNING OF SIMULATION

DO 500 I=1,NC
DO 500 J=1,NR
IF(PERM(I,J,1).EQ.0.0.OR.J.EQ.NR)GO TO 505
T(I,J,1)=PERM(I,J,1)*SQRT(CH(I,J)-CH(I+J,J))+BOTT(I,J)
CONTINUE
IF(PERM(I,J,2).EQ.0.0.OR.I.EQ.NC)GO TO 500
T(I,J,2)=PERM(I,J,2)*SQRT(CH(I,J)-CH(I,J))+BOTT(I,J)
CONTINUE

PREDICT READS FOR NEXT
TIME INCREMENT
DO 51 I*1, NC
DO 51 J*1, NR
D=H(I, J)-HO(I, J)
HO(I, J)=H(I, J)
51 IF(H(I, J).LE.BOT(I, J)) R(I, J)=BOT(I, J)+0.01

C REFINE HEADS

C UPDATE TIME AND INITIALIZE ERROR
AND ITERATION COUNTER

C TIME=TIME+DELTA
ITER=0
52 ITER=ITER+1
IF(MOD(ITER, 10).EQ.0) WRITE(6, 1001) ITER, E
1001 FORMAT(6X, 'ITERATION #', 16, E15.5)
E=0.0

C COLUMN CALCULATIONS

DO 160 J*1, NR
C CALCULATE PEACEMAN-RACHFORD

AA=0.0
IF(H(I, J).LT.DOT(I, J)) R(I, J)=DOT(I, J)
IF(H(I, J).LT.RD(I, J)) GO TO 53

RD=1.0
GO TO 54
53 RE=H(I, J)-RD(I, J)*R(I, J)
RD=0.0
54 IF(H(I, J).LE.CH(I, J)) GO TO 55

S=SF1(I, J)
GO TO 56
55 S=SF2(I, J)
56 DD=S/DELTA+RC(I, J)*RB
DD=HOC(I, J)-CH(I, J)*CSF1(I, J)-SF2(I, J))/DELTA
CC=0.0
IF(J-1).GT.70, 60, 60
AA=T(I, J-1, 1)
BB=BB+T(I, J-1, 1)
70 IF(J-IR).GT.0, 90, 80
80 CONTINUE
83 CC=T(I, J, 1)
BB=BB+T(I, J, 1)
90 IF(J-1).GT.100, 110, 100
100 BB=BB+T(I, 1, J, 2)
DD=DD+H(I, 1, J)*T(I-1, J, 2)
110 IF(I-NC).GT.120, 130, 120
120 CONTINUE
123 DD=BB+T(I, 1, J, 2)
DD=DD+H(I+1, J)*T(I, J, 2)
130 W=BB-AA*B(J-1)
(B(J)=CC/W
140 C(J)=(DD-AA*C(J-1))/W

C RE-ESTIMATE HEADS

E=E+ABS(H(I, NR)-C(NR))
H(I, NR)=C(NR)
N=NR-1
150 NA=C(N)-B(N)*H(I, N+1)
E=E+ABS(NA-H(I, N))
H(I, N)=NA
N=N-1
IF(N.GT.0) GO TO 150

DO 160 J*1, NR
IF(H(I, J).GT.BOT(I, J)) GO TO 160
E=E+BOT(I, J)+0.01-H(I, J)
237

160 CONTINUE

C ROW CALCULATIONS

C

DO 270 J=1, NR
DO 250 I=1, NC
AA=0.0
IF(H(I,J).LT.BOT(I,J))H(I,J)=BOT(I,J)
IF(H(I,J).LT.RD(I,J))GO TO 163
RE=RD(I,J)*R(I,J)
RB=1.0
GO TO 164
163 RE=RRH(I,J)-RD(I,J)*R(I,J)
RB=0.0
164 IF(H(I,J).LE.CH(I,J))GO TO 165
S=SF1(I,J)
GO TO 166
165 S=SF2(I,J)
166 BB=S/DELTA+R(I,J)*R(I,J)
DD=H(I,J)-CH(I,J)*S/DELTA-Q(I,J)+HE
IF(CH(I,J).LT.CH(I,J))DD-DD+
ICH(I,J)-CH(I,J))*CSF1(I,J)-SF2(I,J)/DELTA
CC=0.0
IF(J-1).LT.170,100,170
170 BB=BB+T(I-1,J)
DD=DD+H(I,J)*T(I-1,J,1,1)
180 IF(J-NN).LT.190,200,190
CONTINUE
190 BB=BB+H(I,J)*T(I,J,1,1)
DD=DD+H(I,J)*T(I,J,1,1)
BB=BB+T(I,J,1,1)
200 IF(I-NR).LT.210,220,210
AA=AA+B(I,J,2)
DD=DD+-H(I,J)*T(I,J,2)
BB=BB+H(I,J)*T(I,J,2)
210 CONTINUE
220 BB=BB+T(I,J,2)
CC=-T(I,J,2)
230 CONTINUE
235 BB=BB+T(I,J,2)
CC=-T(I,J,2)
240 BB=BB-CC/2
250 G(I)=(DD-AA*G(I-1))/W
C RE-ESTIMATE HEADS

C

E=E+ABS(H(NC,J)-G(NC))
W(NC,J)=G(NC)
N=NC-1
260 HA=G(N)-R(N)=H(N+1,J)
E=E+ABS(H(N,J)-HA)
W(N,J)=HA
N=N-1
IF(N.GT.0)GO TO 260
DO 270 N=1,NC
IF(H(N,J).GT.BOT(N,J))GO TO 270
E=E+BOT(N,J)+0.01-H(N,J)
W(N,J)=BOT(N,J)+0.01
270 CONTINUE
IF(ITER.GT.9)GO TO 999
IF(E.GT.ERROR) GO TO 52
999 CONTINUE
IF(ISTEP.LT.NSTEPS)GO TO 280
C WRITE(17)H
C PRINT RESULTS

C

DO 281 J=1,49
281 WRITE(7,292)(H(I,J),1=1,25)
292 FORMAT(10F12.2)
WRITE(7,11)
DO 1 J=1,49
1 WRITE(7,292)(BOT(I,J),1=1,25)
WRITE(7,11)
11 FORMAT(2X,'**********')
DO 2 J=1,49
2 WRITE(7,292)(DIF(I,J),1=1,25)
282 WRITEOUT,500)TIME,E,ITER
FORMAT(62,TIME=,FG.2///,E20.7,15)

WRITE OUT 25 X 49 NODES ON TWO ADJACENT FACES
WRITE(OUT,290) (J,((K,J),I=1,13),J=1,49)
WRITE(OUT,291) (J,((K,J),I=14,25),J=1,49)

FORMAT(16,5X,12F8.2///)
CONTINUE

TIMET(ISTEP)=TIME
STORE RESULTS

DO 320 K=1,NN
I=NODE(K,1)
J=NODE(K,2)

HEAD(K,ISTEP)=H(I,J)

WRITE(OUT,301)
FORMAT(ISTEP TIME(DAYS))
DO 1000 ISTEP=1,NSTEPS

WRITE(OUT,302) ISTEP,TIMET(ISTEP)

FORMAT(16,5X,FG.2.

PRINT TIME TABLE

DO 345 K=1,NN
WRITE(OUT,340) (NODE(K,J),J=1,2),(I,I=5,NSTEPS,5)

FORMAT('NODE(',I3',',I3',')///@2010G)
IBOT=1000000
ITOP=-1000000
DO 341 I=1,NSTEPS
LINE(I)=HEAD(K,K)/AL
IF(LINE(I).LT.ITOP) ITOP=LINE(I)

IF(LINE(I).EQ.1/5.0) LINE(I)=VERT

IF(LINE(I).EQ.ITOP) ALINE(I)=PLOT
X=AL*ITOP
WRITE(OUT,344)X,(ALINE(I),I=1,NSTEPS)

FORMAT(F10.3,2K,100A1)
ITOP=ITOP-1
IF(ITOP.GE.IBOT) GO TO 342

CONTINUE

PRINT HEAD SHOWING AREAS WHERE AQUIFER IS DRY AS 'x'

DO 350 I=1,NC
DO 350 J=1,NR
HDRY(I,J)=H(I,J)-BOT(I,J)
IF (HC(I,J),CE.699.95) HDRY(I,J)=10000000.
IF (ABS(HDHY(I,J)).LT.0.015) HDRY(I,J)=10000000.

CONTINUE
WRITE(OUT,300)TIME,E,ITER
WRITE(OUT,290) (J,((K,J),I=1,13),J=1,49)
WRITE(OUT,291) (J,((K,J),I=14,25),J=1,49)
STOP
END
APPENDIX B

PROGRAM B

********** CARBONATE AQUIFER **********

COMPUTER PROGRAM LISTING FOR THE SILURIAN-
DEVONIAN CARBONATE AQUIFER
MODIFIED AFTER
ILLINOIS STATE WATER SURVEY
BASIC AQUIFER SIMULATION PROGRAM WITH
LEAKY ARTESIAN, INDUCED INFILTRATION,
EVAPOTRANSPIRATION, ARTESIAN TO WATER-
TABLE STORAGE COEFFICIENT CONVERSION,
WATER TABLE CONDITIONS, AND VARIABLE NET
WITHDRAWAL RATES

PRICKETT AND LONNQUIST

DEFINITION OF VARIABLES

H0(I,J) ---- HEADS AT START OF TIME
INCREMENT (1, J)
H(I,J) ---- HEADS AT END OF TIME
INCREMENT (FT)
SF1(I,J) ---- STORAGE FACTOR FOR
ARTESIAN CONDITIONS (GAL)
SF2(I,J) ---- STORAGE FACTOR FOR WATER-
TABLE CONDITIONS (GAL/FT)
Q(I,J) ---- CONSTANT WITHDRAWAL
RATES (GPD)
PERM(I,J,1) ---- HYDRAULIC CONDUCTIVITY OF
AQUIFER BETWEEN I,J AND I+1,
J+1 (GAL/DAY/SQ FT)
PERM(I,J,2) ---- HYDRAULIC CONDUCTIVITY OF
AQUIFER BETWEEN I,J AND I+1,
J+1 (GAL/DAY/SQ FT)
T(I,J,1) ---- AQUIFER TRANSMISSIVITY
BETWEEN I,J AND I,J+1
T(I,J,2) ---- AQUIFER TRANSMISSIVITY
BETWEEN I,J AND I,J+1
BOT(I,J) ---- ELEVATION OF BOTTOM OF
AQUIFER (FT)
R(I,J) ---- RECHARGE FACTOR
CAL/DAY/FT)
RH(I,J) ---- ELEVATION OF LAND OR
STREAM SURFACE (FT)
RD(I,J) ---- ELEVATION OF BOTTOM OF
STREAMBED OR ELEVATION
BELOW WHICH EVAPOTRANS-
PIRATION CEASES (FT)
CH(I,J) ---- ELEVATION OF TOP OF
AQUIFER (FT)
B ------- PEACEMAN-RACHFORD
B ARRAY
C ------- PEACEMAN-RACHFORD
C ARRAY
AA, BB, CC, DD --- COEFFICIENTS IN WATER
BALANCE EQUATIONS
NR ------- NO. OF ROWS IN MODEL
NC ------- NO. OF COLUMNS IN MODEL
NSTEPS ------- NO. OF TIME INCREMENTS
DELTA ------- TIME INCREMENTS (DAYS)
I, J, S1, S2, T T ------- EXAMPLE MODEL
DEFAULT VALUES
I ------- MODEL COLUMN NUMBER
J ------- MODEL ROW NUMBER
IP ------- I COORDINATE OF PUMP
JP ------- J COORDINATE OF PUMP
PK, RK ------- PUMPING RATE AT WELL
K AT RATE KC (GPD)
NP ------- NUMBER OF PUMPS
RSP ------- NUMBER OF TIME INCREMENTS
PER PUMPAGE CHANGE
NRT ------- NUMBER OF RATES IN
PUMPING SCHEDULE
```
C
C NN----------NO. OF NODES FOR WI
C NN----------NO. OF NODES FOR WHICH
C TIME-WATER LEVEL CURVES
C ARE DESIRED
C NODE(I,J)----------I,J COORDINATES OF
C NODE WHERE T-WL CURVE
C IS DESIRED
C AL----------SCALE FACTOR FOR T-WL
C CURVES

DIMENSION HC(30,60), HC(30,60),
1SF(30,60), Q(30,60), T(30,60,2),
2D(60), C(60), R(30,60), DL(30,60),
3IP(100), JP(100), P(100,20),
4PERM(30,60,2), BUT(30,60), SF2(30,60),
5UNI(30,60), UN(30,60), CH(30,60),
6NODE(60,2), HEAD(60,100), LINE(100), ALINE(100),
7TIME(60), HDRY(35,70), X(30)
DIMENSION CIDC(2n,08)
DIMENSION DIFC2D, GS)

C
C TURN OFF UNDERFLOW TRAP
C CALL ERRSET(208,256,-1,1)
C
C DEFINE INPUT AND OUTPUT DEVICE NUMBERS
C INTEGER OUT
IN=3
OUT=6

C READ PARAMETER CARD AND
C DEFAULT VALUE CARD
C READ(IN,10)NSTEPS,DELTA,ERROR
10 FORMAT(16,2F6.0)

C READ NODE DATA FOR TIME-
C WATER LEVEL CURVES
C READ(IN,14)NN, AL, ((NODE(I,J), J=1,2), I=1,NN)
14 FORMAT(I5,F3.0;0,IOC 14,13)/11 C 14,13)/4(14.13))

C READ PUMP PARAMETER CARD
C READ(IN,15) NP,NSP,NRT
15 FORMAT(316)

WRITE(OUT,304)
304 FORMAT(2X,112('**'))
WRITE(OUT,303)NSTEPS,DELTA,ERROR
303 FORMAT(8X,'PARAMETER CARD',8X,16,2P10.0)
WRITE(OUT,305)NP,NSP,NRT
305 FORMAT(8X,'PUMP PARAMETER',9X,316)
WRITE(OUT,304)
304 FORMAT(2X,112('**'))
NC=28
NR=53

C READ PUMPING SCHEDULES
C DO 16 I=1, NP
16 READ(IN,17) IP(I), JP(I),
17 FORMAT(2I3,2X,ES.2)
DO 18 I=1, NP
18 WRITE(6,19) IP(I), JP(I),
19 FORMAT(2X,2I3,2X,E14.2)
```
FILL IN T VALUES

DO 101 J = 1, 8
DO 101 I = 1, 19
T(I, J) = 40000.
101
T(I, J) = 40000.
DO 102 J = 9, 10
DO 102 I = 5, 19
T(I, J) = 30000.
102
T(I, J) = 30000.
DO 103 J = 14, 26
DO 103 I = 5, 12
T(I, J) = 30000.
103
T(I, J) = 30000.
DO 104 J = 9, 20
DO 104 I = 1, 4
T(I, J) = 6000.
104
T(I, J) = 6000.
DO 105 J = 27, 39
DO 105 I = 15, 28
T(I, J) = 10000.
105
T(I, J) = 10000.
DO 106 J = 27, 33
DO 106 I = 4, 10
T(I, J) = 7000.
106
T(I, J) = 7000.
DO 107 J = 14, 26
DO 107 I = 13, 19
T(I, J) = 7000.
107
T(I, J) = 7000.
DO 108 J = 1, 21
DO 108 I = 20, 28
T(I, J) = 20000.
108
T(I, J) = 20000.
DO 109 J = 22, 34
DO 109 I = 18, 22
T(I, J) = 6000.
109
T(I, J) = 6000.
DO 110 J = 1, 21
DO 110 I = 20, 28
T(I, J) = 20000.
110
T(I, J) = 20000.
DO 111 J = 1, 14
DO 111 I = 3, 14
T(I, J) = 4000.
111
T(I, J) = 4000.
DO 112 J = 27, 33
DO 112 I = 11, 14
T(I, J) = 7000.
112
T(I, J) = 7000.
DO 113 J = 34, 33
DO 113 I = 11, 14
T(I, J) = 5000.
113
T(I, J) = 5000.
DO 114 J = 40, 55
DO 114 I = 13, 28
T(I, J) = 7000.
114
T(I, J) = 7000.
DO 115 J = 9, 13
DO 115 I = 14, 28
T(I, J) = 10000.
115
T(I, J) = 10000.
DO 116 J = 22, 34
DO 116 I = 18, 22
T(I, J) = 6000.
116
T(I, J) = 6000.
DO 117 J = 8, 35
DO 117 I = 1, 6
T(I, J) = 4000.
117
T(I, J) = 4000.
DO 118 J = 18, 28
DO 118 I = 30, 55
T(I, J) = 5000.
118
T(I, J) = 5000.
DO 20 J = 1, NC
DO 20 I = 1, NR
SF1(I, J) = 2.5E6
SF2(I, J) = 2.5E7
K(I, J) = 0.
ADJUSTING T-VALUES AT BOUNDARY NODES

DO 7 I=1,20,27
DO 7 J=1,H4
7 T(I,J,1)=T(I,J,1)/2
DO 8 J=1,53
I=28
8 T(I,J,2)=0.
DO 9 J=1,58,54
DO 9 I=1,128
9 T(I,J,2)=T(I,J,2)/2
DO 10 I=1,128
10 J=55

READ NODE CARDS
READ(I,H4,I,J,I,J),R(I,J),RH(I,J),RD(I,J),CH(I,J),

START OF SIMULATION
TIME=0
READ(22)IO
DO 388 I=1,NC
DO 388 J=1,HR
IF(Q(I,J).GE.0.)Q(I,J)=-24000.
IF(Q(I,J).GE.100.)Q(I,J)=0.
IF(T(I,J,1).LE.1.)T(I,J,1)=100.
IF(T(I,J,2).LE.1.)T(I,J,2)=100.
PERM(I,J,1)=T(I,J,1)/(CH(I,J)-BOT(I,J))
PERM(I,J,2)=T(I,J,2)/(CH(I,J)-BOT(I,J))
388
K(I,J)=I0(I,J)
DO 1 J=0,47
DO 1 I=1,16
1 QC(I,J)=-40000.
DO 5 J=33,55
DO 5 I=1,14
5 QC(I,J)=-16000.
DO 4 J=1,12
DO 4 I=7,37
4 QC(I,J)=-70000.
DO 3 J=1,12
DO 3 I=1,16
3 QC(I,J)=0.
DO 6 J=48,43
DO 6 I=1,2
6 QC(I,J)=-30000.
DO 119 J=1,12
DO 119 I=1,8
119 QC(I,J)=-00000.

DEL*DELTA
KC=1
DO 320 ISTEP=1,NSTEPS

ENTER PUMPAGE SCHEDULES

Z=(ISTEP-1.0)/KSP+1.0
IF(Z-KC)49,46,49

46 DO 47 K=1,NP
47 QC(I,J)=P(K,KC)
DELTA=DEL.
KC=KC+1
TRANSMISSIVITY CALCULATED AT BEGINNING OF SIMULATION

DO 500 J=1,NC
DO 500 I=1,MR
IF((PERM(I,J).EQ.0.0.OR.J.EQ.MR))GO TO 505

CHECK FOR DIVIDE CHECK IN SQRT CALCULATIONS

CH(I,J)=(AMIN(H(I,J),CH(I,J))·BOT(I,J))·(AMIN(H(I,J+1),CH(I,J+1))·BOT(I,J+1))

IF((PERM(I,J).GT.0.0))GO TO 201
WRITE(6,202) I,J,H(I,J),CH(I,J),BOT(I,J),H(I,J+1),CH(I,J+1),

202 FORMAT(2I3,6E10.3)
201 T(I,J)=PERM(I,J,1)·SQRT((AMIN(H(I,J),CH(I,J))·

BOT(I,J))·(AMIN(H(I,J+1),CH(I,J+1))·BOT(I,J+1)))

CONTINUE

IF((PERM(I,J,2).EQ.0.0.OR.I.EQ.MR))GO TO 500
T(I,J,2)=PERM(I,J,2)·SQRT((AMIN(H(I,J),CH(I,J))·

BOT(I,J))·(AMIN(H(I+1,J),CH(I+1,J))·BOT(I+1,J)))

CONTINUE

PREDICT HEADS FOR NEXT TIME INCREMENT

DO 51 I=1,NC
DO 51 J=1,MR
H(I,J)=HOC(I,J)
HOC(I,J)=H(I,J)
IF(H(I,J).LE.BOT(I,J))H(I,J)=BOT(I,J)+0.01

REFINE HEADS

UPDATE TIME AND INITIALIZE ERROR AND ITERATION COUNTER

TIME=TIME+DELTA
ITER=0
52 ITER=ITER+1
IF(NOT(ITER.10)).EQ.0)WRITE(6,1001) ITER,E
1001 FORMAT(3X,'ITERATION ''.16 .E15.5')

COLUMN CALCULATIONS

DO 140 I=1,NC
DO 140 J=1,MR
CALCULATE PEACEMAN-RACHFORD B AND C ARRAYS

AA=0.0
IF(H(I,J).LT.BOT(I,J))H(I,J)=BOT(I,J)
IF(H(I,J).LT.RD(I,J))GO TO 53
RE=RH(I,J)*RK(I,J)
RB=1.0
GO TO 54
53 RE=(RE·H(I,J))·RD(I,J)·RK(I,J)
RB=0.0
54 IF(H(I,J).LE.CH(I,J))GO TO 55
S=SF1(I,J)
GO TO 56
55 S=SF2(I,J)
56 RD=S·DELTA·RK(I,J)=RB
DD=HO(I,J)·S·DELTA-QT(I,J)+RE
IF((HO(I,J)-CH(I,J))·RD(I,J)-CH(I,J))·LT.0.0)DD=DD+
HH(I,J)-CH(I,J)·(SF1(I,J)-SF2(I,J))/DELTA
CC=0.0
IF(J-1).LT.60,70,60
AA=-T(I,J-1,1)
BB=DB+T(I,J-1,1)
70 IF(J-NR)80,90,80
80 CONTINUE
83 CC=-T(I,J,1)
BB=BB+T(I,J,1)
90 IF(I-1)100,110,100
100 BB=BB+T(I-1,J,2)
BB=BB+H(I-1,J)*T(I-1,J,2)
110 IF(I-NR)120,130,120
120 CONTINUE
123 BB=BB+T(I,J,2)
BB=BB+H(+1,J)*T(1,J,2)
130 H=BB-AA*B(J-1)
H(J)=GC/W
140 G(J)=(DD-AA*G(J-1))/W
C
C RE-ESTIMATE HEADS
C
E=E+ABS(H(1,NR)-G(NR))
H(1,NR)=G(NR)
H=NR-1
150 HA=G(N)-B(N)*H(1,N+1)
E=E+ABS(HA-R(1,N))
H(1,N)=HA
H=N-1
IF(H.GT.0)GO TO 150
DO 160 N=1,NR
IF(H(I,N).GT.BOT(1,N))GO TO 160
E=E+B07CI,N)+0.01-HC I,N)
H(I,N)=BOT(1,N)+0.01
160 CONTINUE
C
C ROW CALCULATIONS
C
DO 270 J=1,NR
DO 250 I=1,NC
AA=0.0
IF(H(I,J).LT.BOT(I,J))H(I,J)=BOT(I,J)
IF(H(I,J).LT.RD(I,J))GO TO 163
H=RD(I,J)*H(I,J)
RM=1.0
GO TO 164
163 RE=RD(I,J)-RD(I,J)*RD(1,J)
RM=0.0
164 IF(H(I,J).LE.CH(I,J))GO TO 165
S=SF1(I,J)
GO TO 166
165 S=SF2(I,J)
166 BB=BB+RD(I,J)*R(I,J)*RB
BB=BB+H(I,J)*S/DELTA-Q(I,J)+RE
IF(H(1,J).LT.CH(I,J))R(I,J)=CH(I,J)
LT.0.0)BB=BB+
1*(H(I,J)-CH(I,J))*(SF1(I,J)-SF2(I,J))/DELTA
CC=0.0
IF(J-1)170,190,170
170 BB=BB+T(I,J-1,1)
BB=BB+H(I,J-1,1)*T(I,J-1,1)
130 IF(J-NR)190,200,190
190 CONTINUE
195 BB=BB+H(I,J+1,1)*T(I,J+1,1)
BB=BB+T(I,J,1)
200 IF(I-1)210,220,210
210 BB=BB+T(I-1,J,2)
AA=TC(I-1,J,2)
220 IF(I-NR)230,240,230
230 CONTINUE
235 BB=BB+T(I,J,2)
CC=TC(I,J,2)
240 W=AA*GB(I-1)
H(I)=GC/W
250 G(I)=(DD-AA*G(I-1))/W
C
C RE-ESTIMATE HEADS
C
E=E+ABS(H(NC,J)-G(NC))
H(NC,J)=G(NC)
N=NC-1
260 HA=G(N)-B(N)*H(N+1,J)
E=E+ABS(H(N,J)-HA)
H(N,J)=HA
I = N - 1
IF (I .GE. 0) GO TO 260
DO 270 N = 1, NC
IF (K(N,J) .GE. BOT(N,J)) GO TO 270
E = E + Bot(N,J) + 0.01 - H(K,J)
H(K,J) = Bot(N,J) + 0.01
270 CONTINUE
IF (IITER .GE. 9) GO TO 999
IF (E .GE. ERROR) GO TO 52
999 CONTINUE
201 J = 1, NR
WRITE (7, 292) (H(I,J), I = 1, NR)
292 FORMAT (10F0.0)
C PRINT RESULTS
C 282 WRITE (OUT, 300) TIME, E, IITER
300 FORMAT (6HTIME*F10.2///, E20.7, 15///)
DO 283 I = 1, NR
283 X(I) = 1
WRITE (6, 284) (X(I), I = 1, 21)
284 FORMAT (10X, 21F5.0//)
C WRITES OUT NC X NR NODES ON TWO ADJACENT PAGES
C WRITE (OUT, 290) (J, (H(I,J), I = 1, 21), J = 1, NR)
DO 285 1 = 22, 28
285 X(I) = 1
WRITE (OUT, 286) (X(I), I = 22, 23)
286 FORMAT (10X, 9X, 7F5.0//)
WRITE (OUT, 291) (J, (H(I,J), I = 22, NC), J = 1, NR)
291 FORMAT (15, 5X, 7F5.0//)
280 CONTINUE
TTIME(ISTEP) = TIME
C STORE RESULTS
C DO 320 K = 1, NN
I = NODE(K, I)
J = NODE(K, 2)
320 HEAD(K, ISTEP) = H(I, J)
C PRINT TIME TABLE
C WRITE (OUT, 301)
301 FORMAT (' ISTEP , TIME(DAYS) ')
DO 1000 ISTEP = 1, NSTEPS
1000 WRITE (OUT, 302) ISTEP, TTIME(ISTEP)
302 FORMAT (16, 6X, F6.2)
C PRINT RESULTS
C DO 345 K = 1, NN
WRITE (OUT, 340) (NODE(K, J), J = 1, 2), (I, I = 5, NSTEPS, 5)
340 FORMAT ('I', 'NODE(', ', ', '13', ',', '13', ')/'///12X05)
ITOP = 1000000
I = 1, NSTEPS
LINE(1) = HEAD(K, I)/AL
IF (LINE(1) .GT. ITOP) ITOP = LINE(1)
341 IF (LINE(1) .LT. BOT) BOT = LINE(1)
DO 343 I = 1, NSTEPS
DATA BLANK, VERT, PLOT/ ' ', '1', 'x' /
ALINE(1) = BLANK
IF (1/I, EQ. 1/BOT) ALINE(1) = VERT
343 IF (LINE(1) .EQ. ITOP) ALINE(1) = PLOT
Y = ALINE(1)/ITOP
WRITE (OUT, 344) Y, (ALINE(1), I = 1, NSTEPS)
344 FORMAT (F10.3, 2X, 100A1)
ITOP = ITOP + 1
IF (ITOP .GE. 1000) GO TO 342
345 CONTINUE
STOP
END
APPENDIX B  Program 3

Computer program to plot the water-table contour map of the Buried-Valley aquifer at Piketon.

C  PROGRAM TO READ DATA STORED ON TAPE AND TO
C  INTERPRET DATA TO MACHINE LANGUAGE.
C  THE DATA ON CARDS IS THEN FED INTO THE PLOTTER
C  TO PLOT THE REQUIRED MAP OR GRAPH.

DIMENSION XDELT(25), YDELT(49), ZT(25), X(49,25), Y(49,25), Z(49,25),
IT(350), T2(350), T3(350), T4(350), T5(350), T6(350), IARY(350)
DIMENSION ZH(2,110)

DATA YDELT/0.0,14*0.0,0*0.25,26*0.80/
DATA XDELT/0.0,10*0.25,14*0.0/
CALL PLOTCL(6000)
IG=2

5 READ(8, 10, END=13) ZM
10 FORMAT(1154X,F6.3,1X,F6.3))
DO 14 I=1, 115
  IF(ZM(I,1).LT.0.0) GO TO 12
C CALL THE PLOTTING PROGRAM FROM THE LIBRARY
C
CALL PLOT(ZM(I,1), ZM(2,1), IC)
IC=2
GO TO 14
12 IC=1
14 CONTINUE
GO TO 5

CALL PLOTTING PROGRAM AGAIN TO BE ABLE
C TO GO TO ORIGIONAL POINT

15 CALL PLOT(0.0,0.0,0.0,1)
   YT=0.0
   DO 30 J=1,49
     YT=YT+YDELT(J)
     X(J)=YT
   30 CONTINUE
   DO 100 J=1,23
     XT=X(J)+XDELT(J)
   100 CONTINUE

CALL AXIS(0.,2.5,' *',1.7.,90.,11.,2.,5)
CALL AXIS(22.,0.,' *',-1.2.,90.,1.,4.,25)
CALL AXIS(9.,9.5,' *',1.13.,0.,23.,2.,5)
CALL AXIS(7.,9.5,' *',1.2.,0.,15.,4.,25)
CALL AXIS(9.,0.,' *',-1.13.,0.,23.,2.,5)
CALL GCTOUR(Y,X,Z,49,25,2.0)
CALL PLOT(0.0,0.0,0.0)
STOP
END
**APPENDIX C**

Table 20. Records of wells drilled in the Buried-Valley aquifer and whose numbers and locations are shown in Plate 1.

<table>
<thead>
<tr>
<th>WELL NUMBER</th>
<th>COUNTY</th>
<th>ELEVATION ABOVE SEA</th>
<th>TOTAL DEPTH</th>
<th>DEPTH TO WATER LEVEL</th>
<th>PUMPING RATE</th>
<th>SPFD/Ft</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>FRAKLIN</td>
<td>720</td>
<td>51</td>
<td>16</td>
<td>500</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>FRAKLIN</td>
<td>720</td>
<td>50</td>
<td>15</td>
<td>500</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>FRAKLIN</td>
<td>720</td>
<td>48</td>
<td>12</td>
<td>150</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>FRAKLIN</td>
<td>720</td>
<td>51</td>
<td>13</td>
<td>100</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>FRAKLIN</td>
<td>720</td>
<td>50</td>
<td>11</td>
<td>200</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>FRAKLIN</td>
<td>720</td>
<td>50</td>
<td>10</td>
<td>150</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>FRAKLIN</td>
<td>720</td>
<td>48</td>
<td>8</td>
<td>100</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>FRAKLIN</td>
<td>720</td>
<td>46</td>
<td>6</td>
<td>130</td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>FRAKLIN</td>
<td>720</td>
<td>45</td>
<td>5</td>
<td>100</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>FRAKLIN</td>
<td>720</td>
<td>40</td>
<td>4</td>
<td>150</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>FRAKLIN</td>
<td>720</td>
<td>40</td>
<td>3</td>
<td>50</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>FRAKLIN</td>
<td>720</td>
<td>40</td>
<td>2</td>
<td>10</td>
<td></td>
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Table 20. Records of wells drilled in the Buried-Valley aquifer and whose numbers and locations are shown in Plate 1. (continued)

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Table 20. Records of wells drilled in the Buried-Valley aquifer and whose numbers and locations are shown in Plate 1. (continued)

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EXPLANATION

Water level contour shows altitude of water level. Contour interval 5 feet. Datum is mean sea level.

Well and well number
Buried valley boundary
WATER LEVEL CONTOUR MAP OF THE BURIED-VALLEY AQUIFER BETWEEN COLUMBUS AND PORTSMOUTH, OHIO.