



113
606
THS



3 1293 10681 7368

LIBRARY
Michigan State
University

This is to certify that the
thesis entitled

THE VARIATION IN HYDRAULIC CONDUCTIVITY
OF HETEROGENEOUS DRIFT DEPOSITS

presented by

CHERYL A. KEHRES

has been accepted towards fulfillment
of the requirements for

M.Sc. degree in Geology

Grahame J. Larson
Major professor

Dr. Grahame J. Larson

Date June 15, 1984



RETURNING MATERIALS:

Place in book drop to
remove this checkout from
your record. FINES will
be charged if book is
returned after the date
stamped below.

~~2002-0000~~

2002-0000

2002-0007

2002-0009

JUL 12 2003

**THE VARIATION IN HYDRAULIC CONDUCTIVITY
OF HETEROGENEOUS DRIFT DEPOSITS**

by

Cheryl A. Kehres

A THESIS

**Submitted to
Michigan State University
in partial fulfillment of the requirements
for the degree of**

MASTER OF SCIENCE

Department of Geology

1984

ABSTRACT

THE VARIATION IN HYDRAULIC CONDUCTIVITY OF HETEROGENEOUS DRIFT DEPOSITS

By

Cheryl A. Kehres

Aquifer test analysis and laboratory permeability studies are the most common methods which have been used to determine the hydraulic conductivities of a leaky artesian system in a glaciated area. These techniques, however, can not adequately determine the variability in hydraulic conductivity of drift deposits that is expected over a broad regional area. In this study a finite-element model is applied to such an area in an attempt to estimate this anticipated range in hydraulic conductivity.

The area chosen for this investigation lies in south-central Michigan and encompasses 16 townships. It is underlain by Pleistocene drift that rests directly on the major bedrock aquifer for the region. A data base of approximately 2000 well logs was used to define drift and aquifer lithology, and areas of probable high recharge. The hydraulic conductivity values of drift determined from this research are 1.8×10^{-8} ft/sec (5.5×10^{-7} cm/sec) for till deposits and 1.8×10^{-7} ft/sec (5.5×10^{-6} cm/sec) for sand and gravel deposits. It can be concluded that application of a finite-element model can assess the variability in hydraulic conductivity of drift deposits over a regional area and should provide a more accurate description of the hydrogeology of regions characterized by such deposits.

To Matt

ACKNOWLEDGEMENTS

There have been many individuals who deserve acknowledgement for their assistance during the course of my research. I would like to extend my appreciation to my thesis committee members, Dr. John Wilband and Dr. James Trow for their review of this manuscript. I also wish to extend my deepest appreciation to Dr. David Wiggert for his patience and guidance in teaching me the application of finite-element analysis. In addition, I wish to extend my sincere appreciation to Dr. Grahame Larson, my committee chairman, who not only taught me how to approach a research problem and provided expertise in the field of hydrogeology, but also gave support to me as a friend.

I would also like to thank Mr. Ron Hohenstein of the Lansing Board of Water and Light, Mr. David Jansen of the East Lansing - Meridian Township Water and Sewer Authority, Mr. Bob Roller of the Tri-County Regional Planning Commission, and individuals of Delta Township, Snell Environmental Group, and Keck Consulting. These individuals were most helpful in extending the technical data which made my research possible.

Also, I wish to thank Dr. Genevieve Segol and Mr. Tim Kelly who gave me the opportunity to gain professional experience while I was working on my Master's degree, as well as Dr. Satya Sehgal and Kim Kesler of GMC Associates for giving me the flexibility to finish my thesis. Finally, I wish to thank Bill Monaghan for all his help as well as Sue, Sandy, Bruce, Steve, and my parents.

TABLE OF CONTENTS

LIST OF FIGURES.....	v
LIST OF TABLES.....	vi
INTRODUCTION.....	1
STUDY AREA.....	2
HYDRAULIC CONDITIONS.....	8
FINITE-ELEMENT MODEL.....	15
Finite-Element Grid.....	16
Assumptions.....	16
CALIBRATION.....	21
DISCUSSION.....	42
Electric-Analog Model.....	42
Dewatering.....	43
CONCLUSIONS.....	45
BIBLIOGRAPHY.....	47
APPENDIX.....	49

LIST OF FIGURES

Figure 1:	Study Area.....	4
Figure 2:	Schematic of Drift Overlying the Saginaw Formation.....	7
Figure 3:	Bedrock Lithology Map.....	10
Figure 4:	Areas of Direct Hydraulic Connection through Drift to Bedrock.....	13
Figure 5:	Finite-Element Grid.....	18
Figure 6:	Finite-Element Grid with Pumping Centers.....	50
Figure 7:	Simulated 1945 Potentiometric Surface.....	24
Figure 8:	Actual 1945 Potentiometric Surface.....	26
Figure 9:	Simulated 1964 Potentiometric Surface.....	28
Figure 10:	Actual 1964 Potentiometric Surface.....	30
Figure 11:	Simulated 1979 Potentiometric Surface.....	33
Figure 12:	Actual 1979 Potentiometric Surface.....	35
Figure 13:	Final Simulated 1945 Potentiometric Surface..	38
Figure 14:	Final Simulated 1964 Potentiometric Surface..	40

LIST OF TABLES

Table 1: Pumping Schedules.....	51
Table 2: Hydraulic Conductivity of Drift.....	46

INTRODUCTION

Often in glaciated areas the hydrologic regime is that of a leaky artesian system with drift acting as a semi-pervious layer. This condition has prompted several investigations in recent years that have focused primarily on the hydraulic properties of drift. They have included laboratory permeability studies (Desaulniers, et al. 1981; Norris, 1962), aquifer testing (Walton, 1962), and digital modeling (Grisak and Cherry, 1975). Collectively, these studies have shown that drift can have a range in hydraulic conductivity on the order of 10^{-13} to 10^{-7} ft/sec (10^{-11} to 10^{-5} cm/sec).

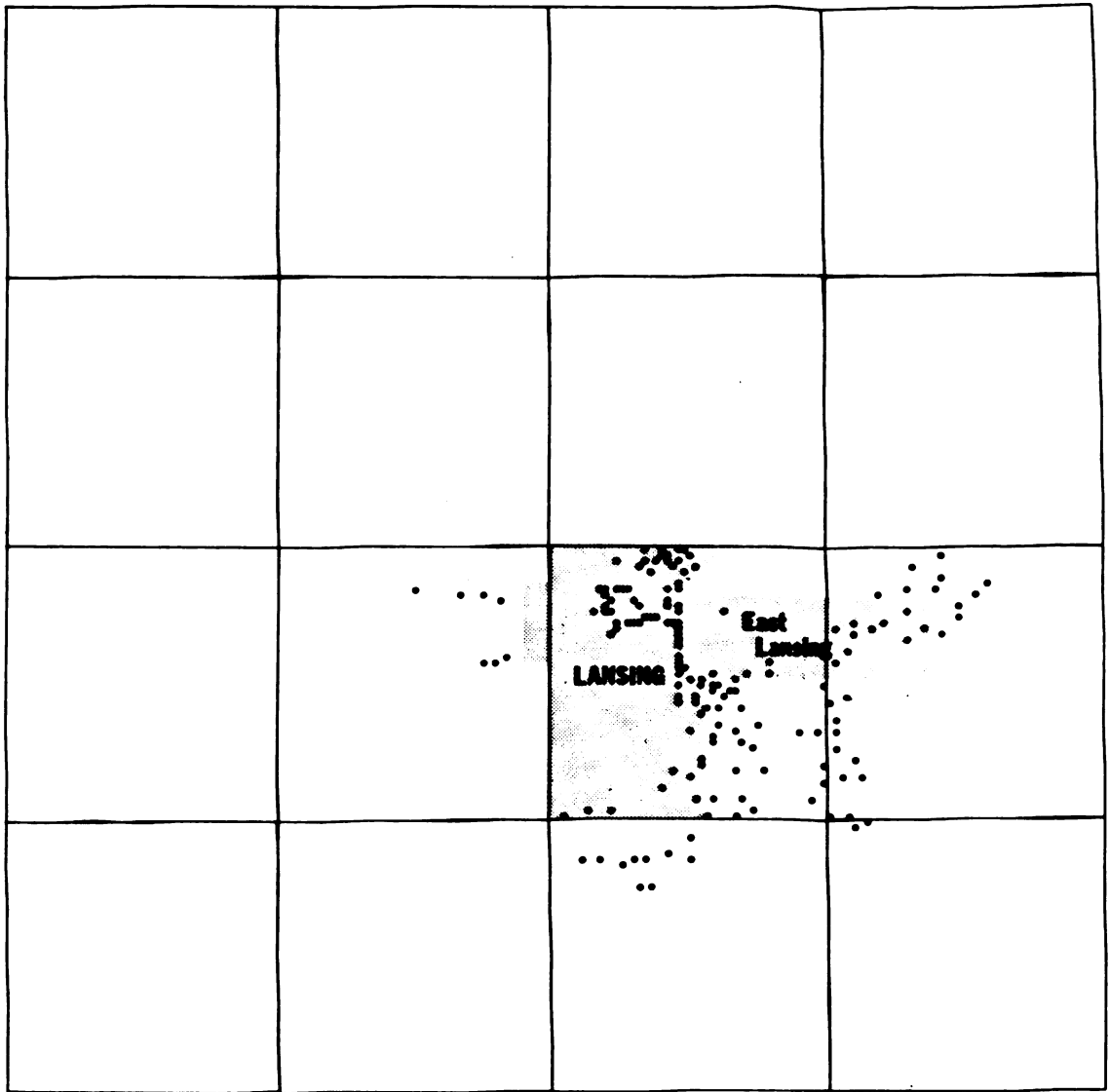
The above investigations have been mainly site specific rather than regional, and have generally treated the drift as a homogeneous unit. However, in a broad region characterized by heterogeneous drift deposits a range in hydraulic conductivity values as opposed to a single value can be expected. Therefore, in this investigation an analysis utilizing a finite-element model is applied to a broad region characterized by varied drift deposits in an attempt to estimate this anticipated range in hydraulic conductivity.

STUDY AREA

The area chosen for this investigation lies in south-central Michigan (Figure 1) and encompasses 16 townships that make up the Lansing - East Lansing metropolitan region. It is underlain by Pleistocene drift that varies from 50 to 150 feet (15 to 46 m) in thickness (Stuart, 1945; Vanlier, et al. 1973). This drift rests directly on the Saginaw Formation which is the major bedrock aquifer for the region. The Saginaw Formation is confined underneath by the Bayport Limestone. The entire area depends almost exclusively upon groundwater for its water supply (39.6 mgd) and contains approximately 1800 private wells and 167 municipal and industrial wells.

The Saginaw Formation is composed chiefly of interbedded sandstone and shale (Kelly, 1936; Vanlier, et al. 1973). It dips slightly towards the northeast and ranges in thickness from approximately 350 feet (107 m) in the southern part of the study area to 400 feet (122 m) in the northern part. The transmissibility of the aquifer has been determined from previous aquifer studies (Stuart, 1945; Firouzian, 1963) and ranges between 6.2×10^{-3} ft²/sec (5.76 cm²/sec) and 0.12 ft²/sec (111.5 cm²/sec). Assuming an average aquifer thickness of 375 ft (114 m) these values result in a range of hydraulic conductivity from 1.2×10^{-5} ft/sec (3.6×10^{-4} cm/sec) to 3.3×10^{-4} ft/sec (1.0×10^{-2} cm/sec).

Figure 1. Study Area. Darkened circles indicate municipal wells.

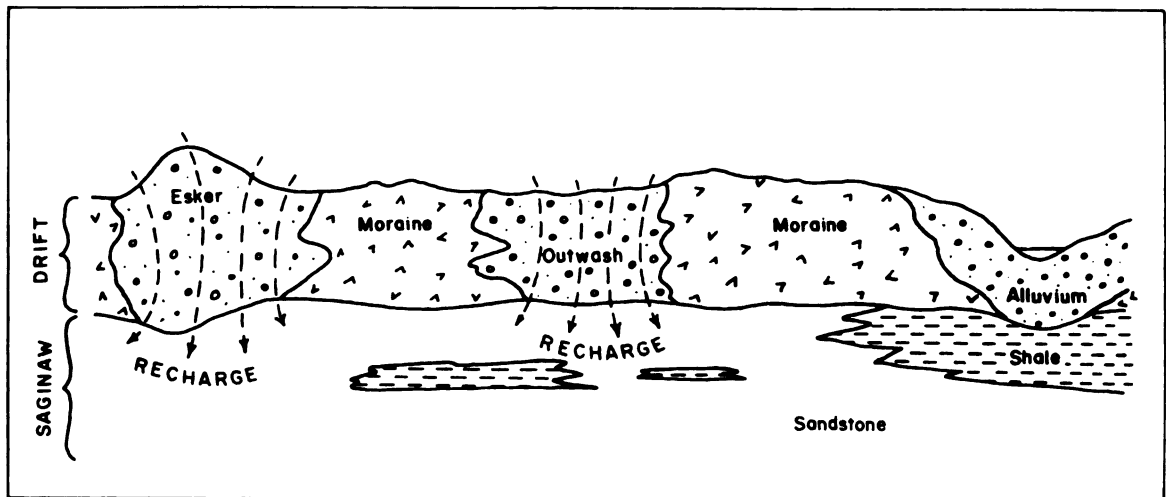


0 5 miles



The drift in the region consists mostly of clay till. However, in some areas there are localized deposits of sand and gravel associated with eskers and rivers. It has been proposed that recharge through the drift to the underlying aquifer occurs mainly through the permeable materials associated with the esker and river deposits; whereas, less recharge occurs through the till or where shales are in direct contact with the drift (Stuart, 1945; Vanlier, et al. 1973; Long, et al. 1981; Monaghan and Ritter, 1982), (Figure 2). It is this geologic setting which creates a leaky artesian groundwater system in the study area. A value of 1.8×10^{-7} ft/sec (5.5×10^{-6} cm/sec) has been suggested for the hydraulic conductivity of the sands and gravel (Lovato and Larson, 1979). However, conductivity values for the till have not been determined.

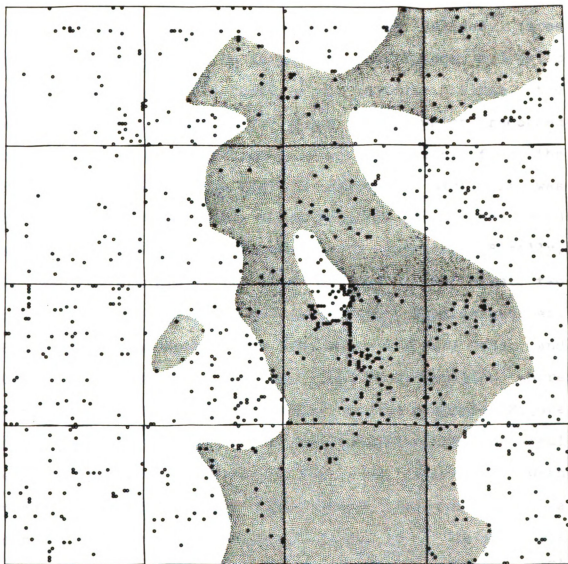
Figure 2. Schematic of Drift Overlying the Saginaw Formation.



HYDRAULIC CONDITIONS

A bedrock lithology map was prepared from a computerized data base of 893 private and municipal wells in the study area. To determine the lateral and vertical change in lithology of the aquifer, the Saginaw Formation was divided into three major lithologic units: sandstone, shale, and sandstone and shale (i.e. sandy shale, shaley sandstone, and thin beds of alternating shale and sandstone). If a well log recorded greater than 33% sandstone, the bedrock in the vicinity of the well was defined as sandstone. Similarly, if the well log recorded greater than 33% shale or greater than 33% shale and sandstone, the bedrock in the vicinity of the well was defined as shale or shale and sandstone, respectively. This averaging technique was applied at depths of 10 feet, 100 feet, 150 feet, and greater than 300 feet into the aquifer. It was found that the aquifer lithology was fairly constant with depth and consisted of a higher percentage of sandstone in the center of the study area than to the east or west (Figure 3). In conjunction with a review of results of numerous pump tests which have been performed in the study area (Keck Consulting; Stuart, 1945; East Lansing - Meridian Water and Sewer Authority), defining the bedrock lithology enabled assignment of hydraulic conductivity values to the aquifer. As a result, the center of study area was initially assigned a

Figure 3. Bedrock Lithology Map. Shaded area represents mostly sandstone; unshaded area represents mostly shale. Darkened circles indicate municipal wells; open circles indicate private wells.

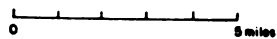
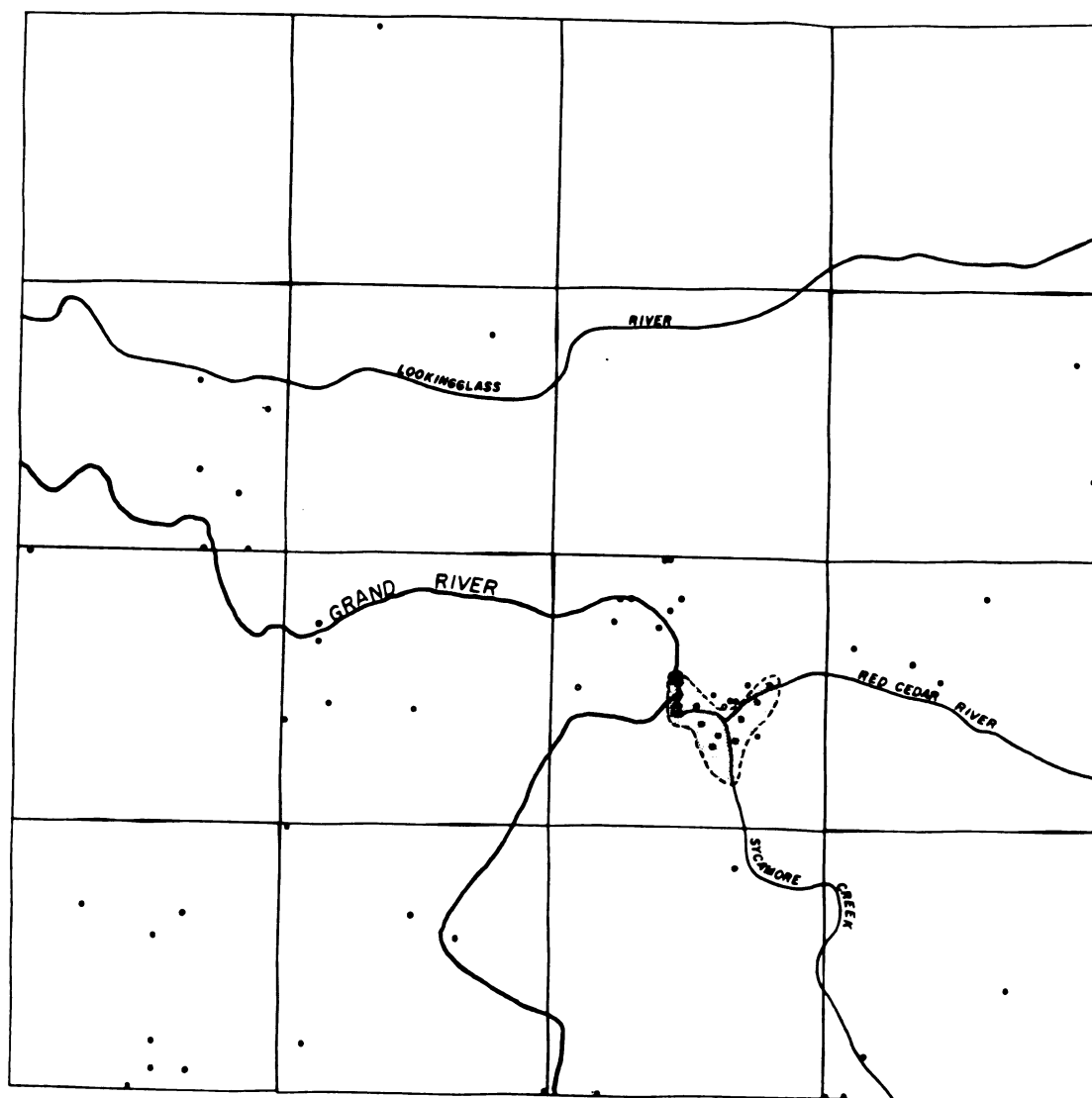


conductivity value of 7.9×10^{-5} ft/sec (2.4×10^{-3} cm/sec), whereas, the area consisting of a higher percent of shale was assigned a conductivity value of 2.3×10^{-5} ft/sec (7.0×10^{-4} cm/sec).

Once the aquifer conductivity was defined, three criteria were applied to define areas of direct hydraulic connection through the drift to the bedrock. First, the well file was searched for those wells which recorded continuous units of permeable sand and gravel from the ground surface to the bedrock aquifer. Next, it was determined if those wells which recorded continuous sand and gravel were in direct contact with sandstone. Finally, it was determined if the areas defined by the first two conditions were near any rivers or other recharge sources. From this assessment, the area where the Red Cedar and Grand Rivers converge was identified as a location where substantial recharge to the aquifer seemed probable (Figure 4). Also identified were a few deposits of smaller areal extent that were in direct contact with sandstone. With these areas of suspected high recharge differentiated, a conductivity value of 3.3×10^{-9} ft/sec (10^{-7} cm/sec) was assigned to the till; a value of 3.3×10^{-8} ft/sec (10^{-6} cm/sec) was assigned to the permeable sands and gravel of the suspected high recharge areas.

In conjunction with the hydraulic values of the drift and bedrock aquifer, pumping data were assembled for the period 1945-79. Because of the configuration of the

Figure 4. Areas of Direct Hydraulic Connection through Drift to Bedrock. Circles represent wells that recorded continuous sand and gravel to bedrock. Shaded area represents identified high recharge area.



municipal well field, the wells were then clustered into pumping centers. With this information a finite element grid was designed.

FINITE-ELEMENT MODEL

The finite-element model (GEOFLOW) used in this study is a two dimensional computer program which utilizes the Galerkin-based finite-element method to transform the mathematical equations governing groundwater flow to a system of first-order partial differential equations as a means of evaluating the potentiometric surface over the study area (Haji Djafari, 1982). Features of the hydrodynamic model are:

- 1) Time-variable pumping rates can be modeled.
- 2) The system can be divided into regions, and each region can have different hydraulic conductivity parameters such as the transmissibility, recharge, and storage coefficients.
- 3) Initial hydraulic heads can be specified or can be computed by the program.
- 4) Along any designated element, a line source can be incorporated.
- 5) At any designated node, the value of the hydraulic head can be specified (Dirichlet boundary condition).
- 6) The thickness of aquifer can be updated if required.
- 7) Flow vectors are computed at the center of each element.

Finite-Element Grid

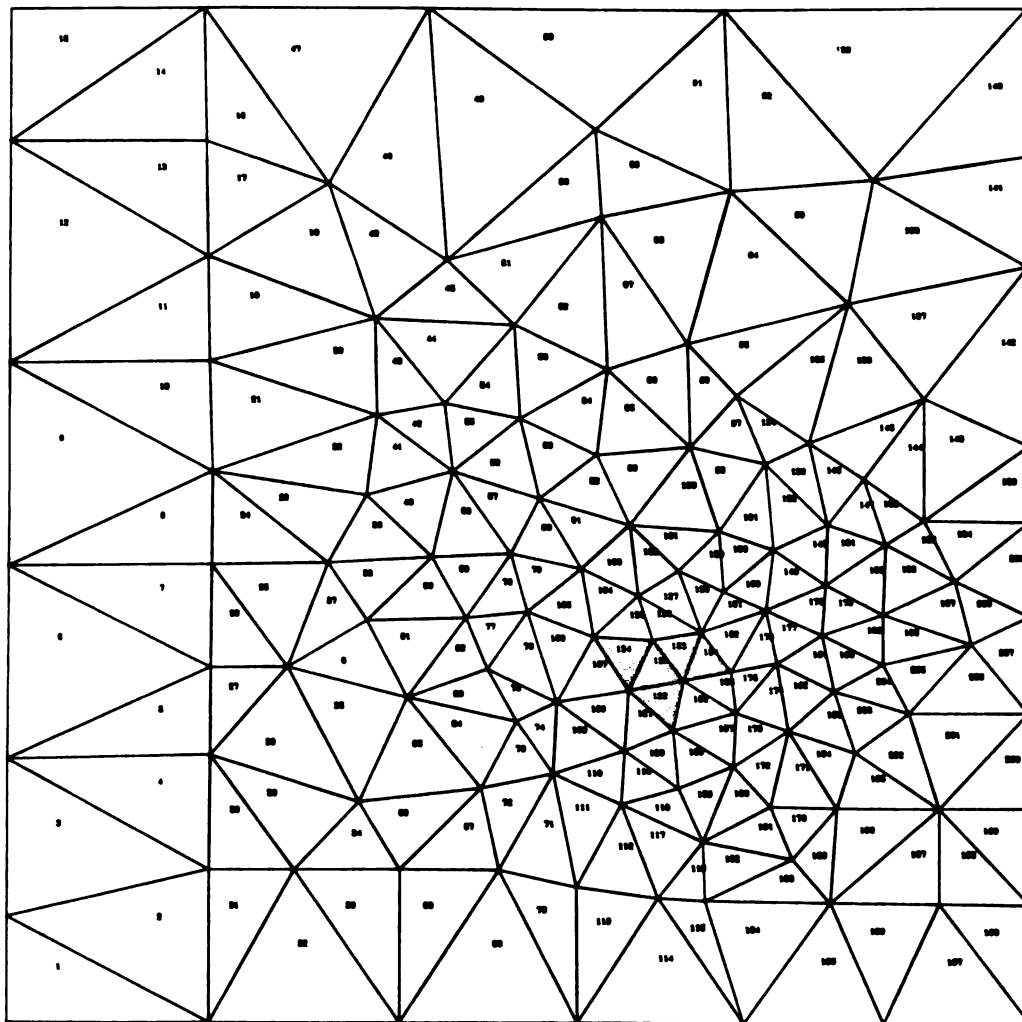
In the design of the finite-element grid (Figure 5) three major factors were considered. First was the aquifer hydraulic conductivity, second was the configuration of the municipal well field, and third was the hydrologic character of the drift. In regards to the aquifer hydraulic conductivity the elements of the grid were designed to conform to the facies relationship between the two conductivity zones as previously defined. To establish more detail near the municipal well field, the elements were also made smaller nearer the well field and generally larger farther away. The finite-element grid delineating pumping centers is included in the Appendix (Figure 6). The heterogeneous nature of the drift was also delineated by assigning a higher vertical conductivity value over the elements representing the suspected high recharge area (Figure 5). In total, the finite-element grid consisted of 209 elements and 118 nodes.

Assumptions

The main assumptions that were used to implement the program were as follows:

- 1) The aquifer was treated as being under semi-confined conditions. Although the drift is acting as the overlying confining bed, there is significant amount of leakage through the

Figure 5. Finite-Element Grid. Shaded area represents identified high recharge area.



53 element no.

0 5 miles



heterogeneous deposits characterizing the drift to the Saginaw aquifer.

- 2) Each node along the boundary of the study area was assumed to have a constant hydraulic head as prescribed by the Dirichlet condition (Prescribed Potential Boundary Condition) since historical data has shown that the potentiometric surface around the perimeter of the study area has not changed appreciably since 1900.
- 3) The thickness of the aquifer was assumed constant at 375 feet (114 m). An isopach map of the thickness of the Saginaw Formation was constructed from oil well logs in the area. The thickness was found to vary from 300 feet in the south to greater than 400 feet in the northeast. From the map it was determined that assuming a constant thickness of 375 feet for the Saginaw aquifer over the study area was reasonable.
- 4) Leakage from the rivers was only assumed to be important in those areas where the well logs recorded continuous units of sand and gravel to the bedrock.
- 5) Initial hydraulic heads were determined from historical data in 1900 because at that time there were no significant stresses imposed on

the groundwater system (Wheeler, 1967).

- 6) Resistivity values for drift were computed assuming an average thickness of 87.5 feet (27 m), where resistivity is defined as the thickness of a semi-confining layer divided by its' hydraulic conductivity. An isopach map of the thickness of the drift was computer generated. From this map an average drift thickness of 87.5 feet was chosen.

CALIBRATION

Calibration of the finite-element model was achieved with application of the indirect approach (Neuman, 1973) whereby input parameters are adjusted with repetitive application of the model until computed heads matched the observed field values (Freeze and Cherry, 1979). Over 60 simulation runs were made adjusting the resistivity of the drift, the hydraulic conductivity of the aquifer, the storage coefficient, and the boundary conditions to obtain a solution which best approximated the actual hydrologic conditions. Included in the Appendix are the various pumping schedules used in calibrating the finite element model (Table 1) and a sample computer run.

The model was initially calibrated to simulate 1945 steady-state conditions. Conductivity parameters were assigned uniform values so as not to force artificial complexities on the system. Therefore, the hydraulic conductivities of the drift and the aquifer were assigned the estimated values of 4.1×10^{-9} ft/sec (1.25×10^{-7} cm/sec) and 9.8×10^{-6} ft/sec (3.0×10^{-4} cm/sec), respectively. The pumpage in 1945 totaled 14 mgd over the study area.

Initial hydraulic heads were input as ϕ (head) zones based on data for the year 1900 when the aquifer was virtually undeveloped. The ϕ zones and hydraulic conductivity of the aquifer were then adjusted singularly keeping the hydraulic conductivity of the drift uniform

until the simulated potentiometric surface (Figure 7) matched the historical potentiometric surface (Figure 8) fairly well. Five initial ϕ zones proved to be a good representation of initial conditions in 1900. In addition, differentiating the hydraulic conductivity of the aquifer into two zones (7.9×10^{-5} ft/sec for the sandstone facies and 2.3×10^{-5} ft/sec for the shale facies) as originally anticipated, produced a good fit for the 1945 conditions.

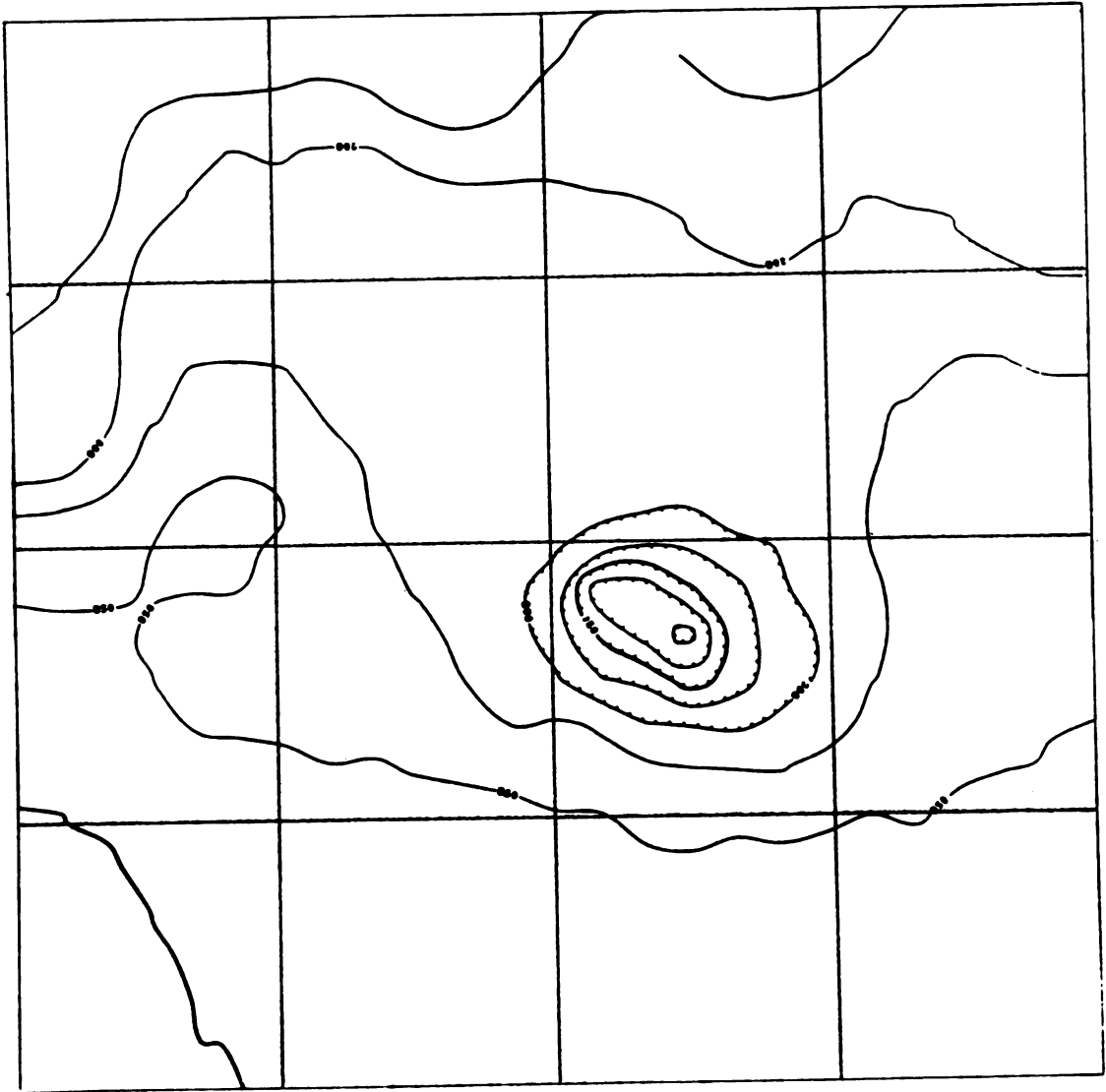
After synthesis of the 1945 potentiometric surface a transient analysis was performed for the period 1945-64. In this analysis the initial ϕ zones were altered slightly in addition to checking the model sensitivity to various time steps. The values of drift conductivity and aquifer conductivity as determined from the 1945 calibration were used in this calibration. In addition, a storage coefficient of 10^{-4} was assigned to the bedrock aquifer. The model was not sensitive to the time step as values of 1, 2.5, and 5 years produced very little difference in the resulting hydraulic heads for 1964. Therefore, a 1 year time step was considered adequate. In 1964 27.5 mgd of water was withdrawn from the aquifer. The fit obtained from this calibration was reasonable considering that the drift heterogeneity had not yet been addressed. Figure 9 shows the 1964 simulated potentiometric surface and Figure 10 shows the actual 1964 potentiometric surface.

With the 1964 match completed, a transient analysis was performed for the period 1956-79. In this analysis the

Figure 7. Simulated 1945 Potentiometric Surface.

Figure 8. Actual 1945 Potentiometric Surface.

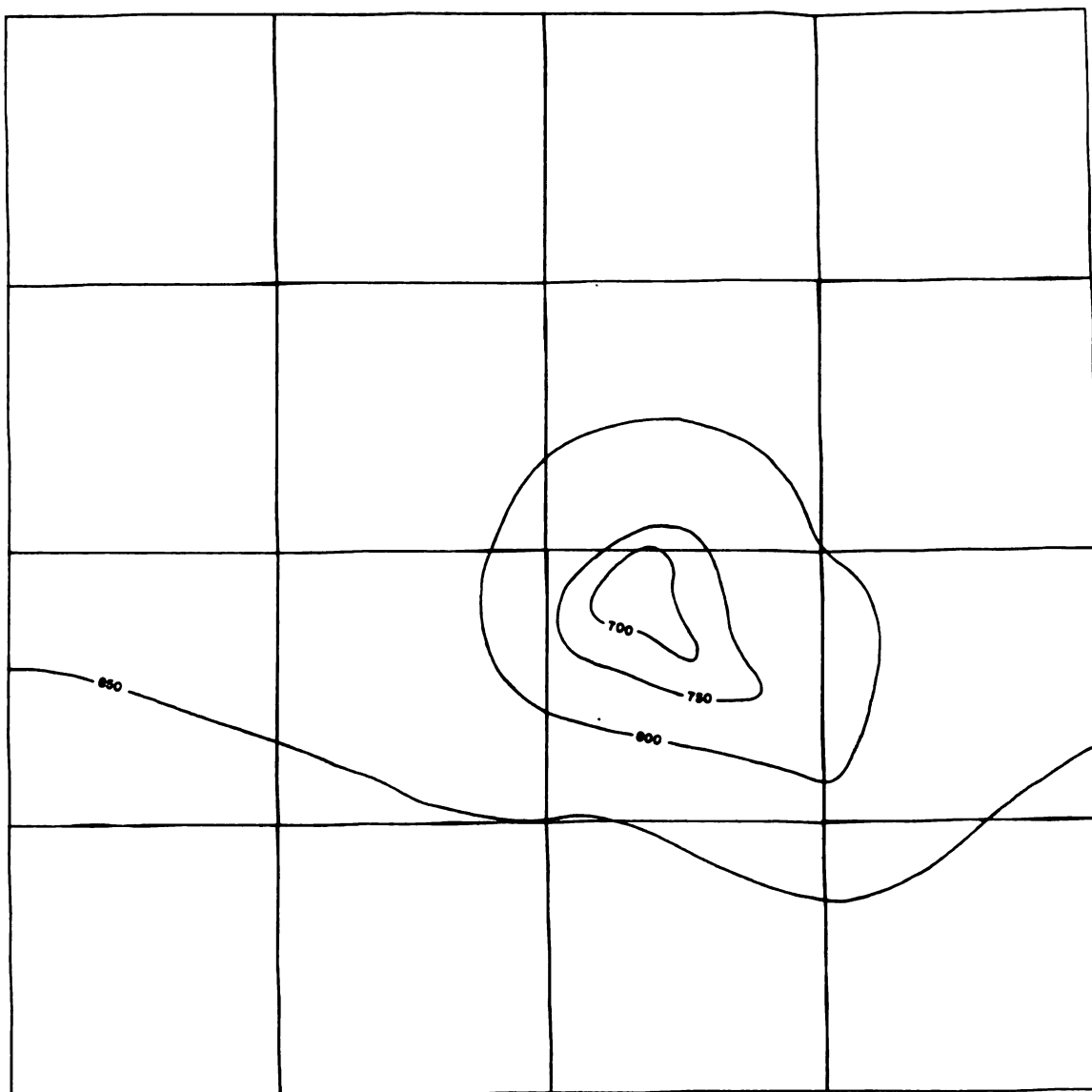
Figure 9. Simulated 1964 Potentiometric Surface.



2
↑

0 1 2 3 4 5 6 miles

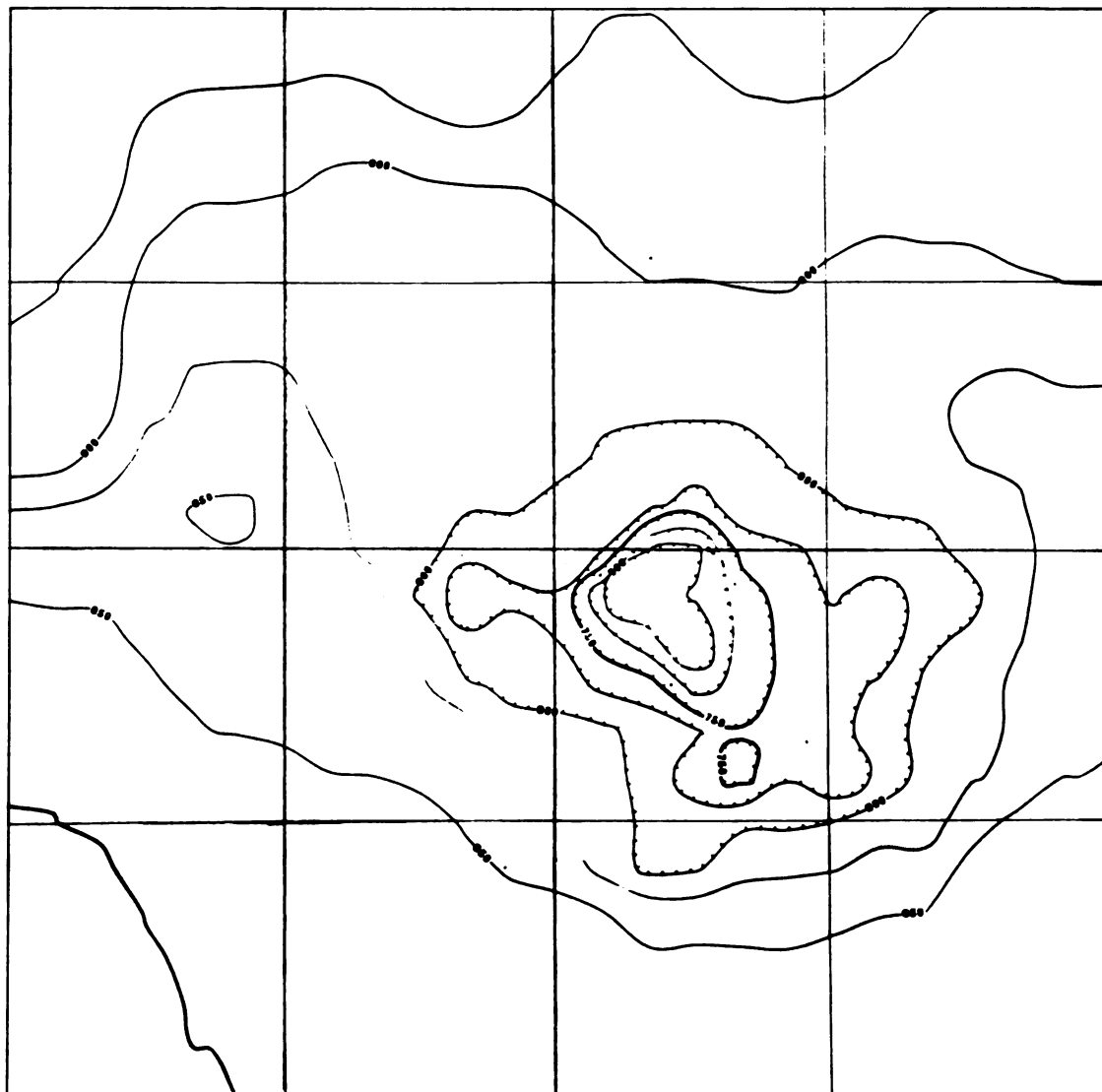
Figure 10. Actual 1964 Potentiometric Surface.



0 5 miles

drift hydraulic conductivity was addressed. Using the values of the aquifer conductivity, storage coefficient, time step, and ϕ zones as determined from the 1945 and 1964 calibrations, the drift hydraulic conductivity was adjusted until the simulated potentiometric surface matched the actual 1979 conditions. Values of 1.5×10^{-8} to 3.2×10^{-8} ft/sec (4.6×10^{-7} to 9.8×10^{-7} cm/sec) were assigned to the conductivity of the till; whereas, values of 5.3×10^{-8} to 2.3×10^{-5} ft/sec (1.6×10^{-6} to 7.0×10^{-4} cm/sec) were assigned to the conductivity of the suspected high recharge areas. From this analysis it was found that assigning a differential conductivity value to the permeable sand and gravel deposits of smaller areal extent produced a mounding effect. Therefore, these deposits may be important for recharge locally but on a regional level they were determined to be insignificant. The area where the Red Cedar and Grand Rivers converge, however, was determined to be an important avenue for recharge. Conductivity values of 1.8×10^{-7} ft/sec (5.5×10^{-6} cm/sec) for the river deposit and 1.8×10^{-8} ft/sec (5.5×10^{-7} cm/sec) for the till produced a good fit when compared with water well level data for 1979 from USGS observation wells (Huffman, 1980), Lansing Board of Water and Light municipal wells, and State of Michigan water well records. Figure 11 shows the 1979 simulated potentiometric surface and Figure 12 shows the 1979 actual potentiometric surface. The slight irregularity in the contours of the simulated

Figure 11. Simulated 1979 Potentiometric Surface.



0 5 miles

Figure 12. Actual 1979 Potentiometric Surface.

potentiometric surface near the center of the cone of depression can be attributed to artificial influence caused by clustering pumping wells into a limited number of pumping nodes. Pumpage for 1979 totaled 39.6 mgd in the Lansing area.

With an acceptable match obtained, sensitivity analyses were performed by varying the aquifer conductivity and storage coefficient. Values ranging from 6.2×10^{-5} to 9.3×10^{-5} ft/sec (1.9×10^{-3} to 2.8×10^{-3} cm/sec) and 1.5×10^{-5} to 3.1×10^{-5} ft/sec (4.6×10^{-4} to 9.4×10^{-4} cm/sec) were assigned to the sandstone facies and shale facies respectively, to determine the best estimate of the aquifer conductivity. The model was only slightly sensitive to these changes and produced the best fit using conductivity values of 7.9×10^{-5} ft/sec (2.4×10^{-3} cm/sec) for the sandstone and 2.3×10^{-5} ft/sec (7.1×10^{-4} cm/sec) for the shale. The model was run with storage coefficients ranging from 10^{-2} to 10^{-6} , however, the simulations showed no sensitivity to the storage coefficient. Therefore, a storage coefficient of 10^{-4} was assumed adequate for calibration. With the model calibrated to the 1979 data the 1945-64 transient analysis and the 1945 steady state analysis was again performed to check the calibration. Figure 13 shows the final simulated 1945 potentiometric surface and Figure 14 shows the final simulated 1964 potentiometric surface. In both cases the computed

Figure 13. Final Simulated 1945 Potentiometric Surface.

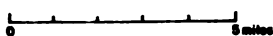
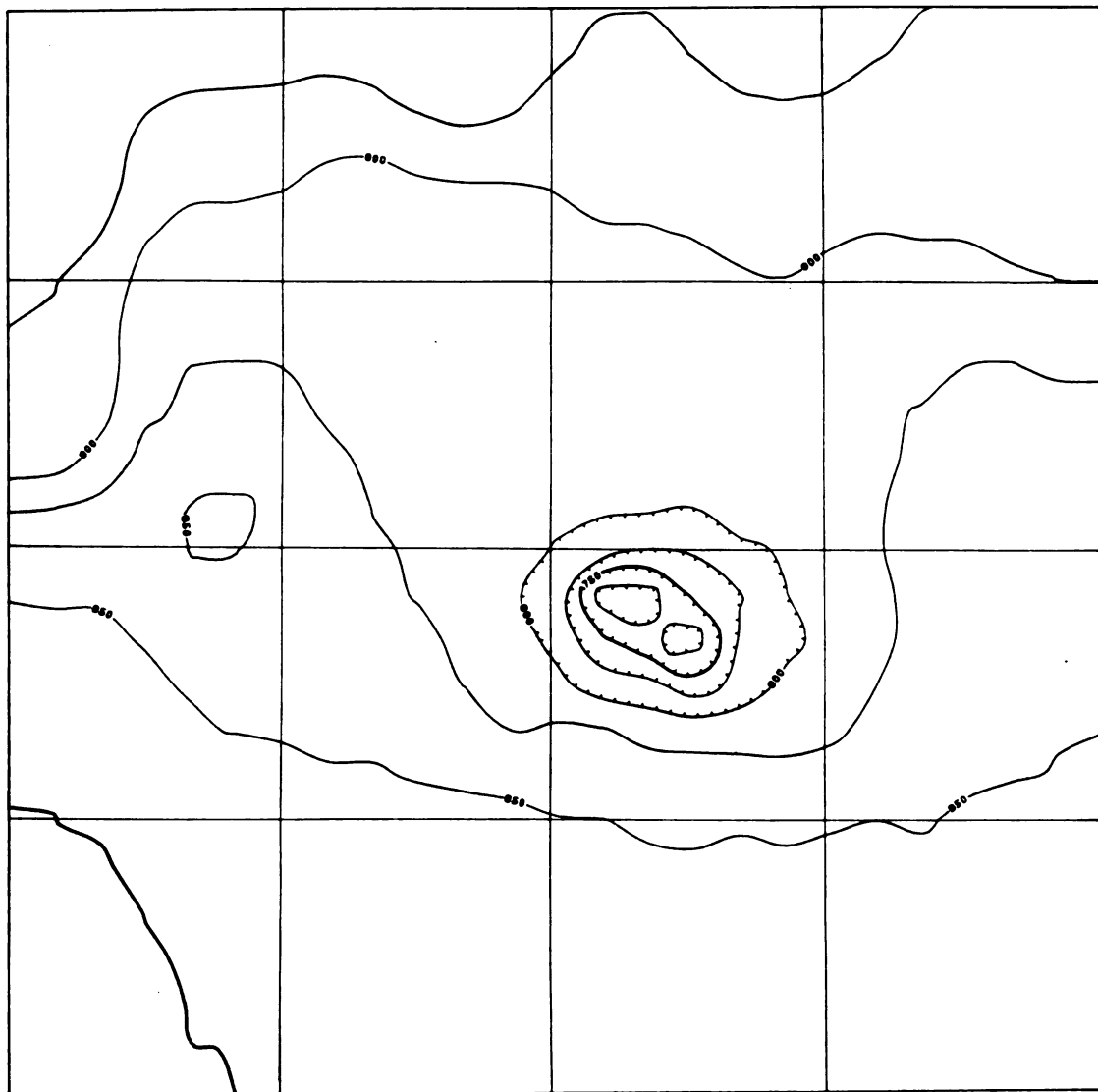


Figure 14. Final Simulated 1964 Potentiometric Surface.

potentiometric surface matched well with the actual measured surface.

DISCUSSION

Electric-Analog Model

Approximately 15 years ago, an electric-analog model (Wheeler, 1967; Vanlier and Wheeler, 1968) was utilized in the Lansing area to predict the change in the potentiometric surface with time due to increased pumpage. In the application of the analog model leakage was contributed along the total length of the rivers. However, even though the amount of pumpage input to the analog model in 1975 (43 mgd) virtually matched that actually pumped in 1979 (39.5 mgd), drawdown in the downtown Lansing area in 1975 was overpredicted by 50 feet when compared to actual 1979 conditions (Huffman, 1980; Lansing Board of Water and Light municipal well data). Furthermore, the drawdown in the East Lansing area predicted by the 1975 analog simulation was 83 feet greater than the actual 1976 conditions (Van Til, 1977).

To assess the hydraulic character of the drift, leakage values were used in the analog analysis, whereas, in the finite-element analysis hydraulic conductivity values were used. However, the recharge values input to the analog model appear to have been underestimated since all other parameters used in the analog model are fairly close to those used in the finite-element model.

Dewatering

In 1979 the potentiometric surface in the northern Lansing area was reported to be 75 feet (23 m) below the surface of the Saginaw aquifer, while in the southern Lansing area, near the identified high recharge area, it was reported to be as much as 25 feet (8 m) below the surface of the aquifer (Keck Consulting; Huffman, 1980). These data which were collected from wells that penetrate at least 155 feet (47 m) into the aquifer would suggest that the Saginaw aquifer is locally being dewatered.

In order to evaluate the significance of dewatering in the Lansing area, a dewatering option (DEWA) was applied to the finite element model. The potentiometric heads calculated when using this option were found to be nearly identical (± 5 ft) to those simulated when no dewatering option was applied. This lack of significant head difference would suggest that perhaps the finite element grid is not refined enough in the vicinity exhibiting dewatered conditions to adequately simulate such conditions. An alternative explanation is that the observation well data may reflect mainly vertical pressure differences due to shale lenses occurring within the aquifer, rather than actual dewatering conditions.

To determine which of the two explanations is correct, shallow observation wells would have to be drilled to test if dewatering is occurring. In addition, a more refined



grid could be applied to the area exhibiting dewatered conditions.

If dewatering is occurring in the Lansing area, water may be drawn into the Saginaw aquifer from the underlying Bayport Limestone. This could result in contamination of the Saginaw aquifer since the Bayport Limestone produces saline water in areas where it is overlain by the Saginaw Formation (Vanlier, et al., 1973).

CONCLUSIONS

The methods which have been most commonly used to determine the hydraulic conductivities of a leaky aretesian system in a glaciated area are aquifer test analyses and laboratory permeability studies. These techniques, though, can not adequately determine the variability in hydraulic conductivity of drift deposits that is expected over a broad regional area. Application of a finite-element model, however, can assess this variability and should provide a more accurate description of the hydrologic regime of a region characterized by heterogeneous deposits.

The hydraulic conductivity values for drift obtained in this investigation can be compared to values obtained by other researchers in the United States and Canada (Table 2). The conductivity values of drift as determined from previous investigations range from 3.3×10^{-13} to 3.3×10^{-7} ft/sec (10^{-11} to 10^{-5} cm/sec). The final values determined from this research are 1.8×10^{-8} ft/sec (5.5×10^{-7} cm/sec) for till deposits and 1.8×10^{-7} ft/sec (5.5×10^{-6} cm/sec) for sand and gravel deposits. These hydraulic conductivity values are clearly within the range defined by others.

TABLE 2
HYDRAULIC CONDUCTIVITY OF DRIFT (cm/sec)

SOUTHEASTERN MANITOBA (Grisak and Cherry, 1975)	SOUTHWESTERN ONTARIO (Desaulniers, et al., 1981)	ILLINOIS, OHIO AND SOUTH DAKOTA (Norris, 1962)	SOUTHEASTERN ILLINOIS (Walton, 1960)	SOUTHCENTRAL MICHIGAN (Kehres, et al., 1983)
Lab 9.0×10^{-11} - 2.0×10^{-10}	Lab 9.3×10^{-8} - 8.3×10^{-7}	Field and Lab	Field	FEM
FEM 1.8×10^{-7}	Field 7.9×10^{-8} - 2.4×10^{-7}	9.5×10^{-9} - 4.3×10^{-5}	3.8×10^{-6} - 7.6×10^{-5}	6.0×10^{-7} - 6.0×10^{-6}

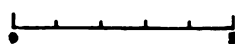
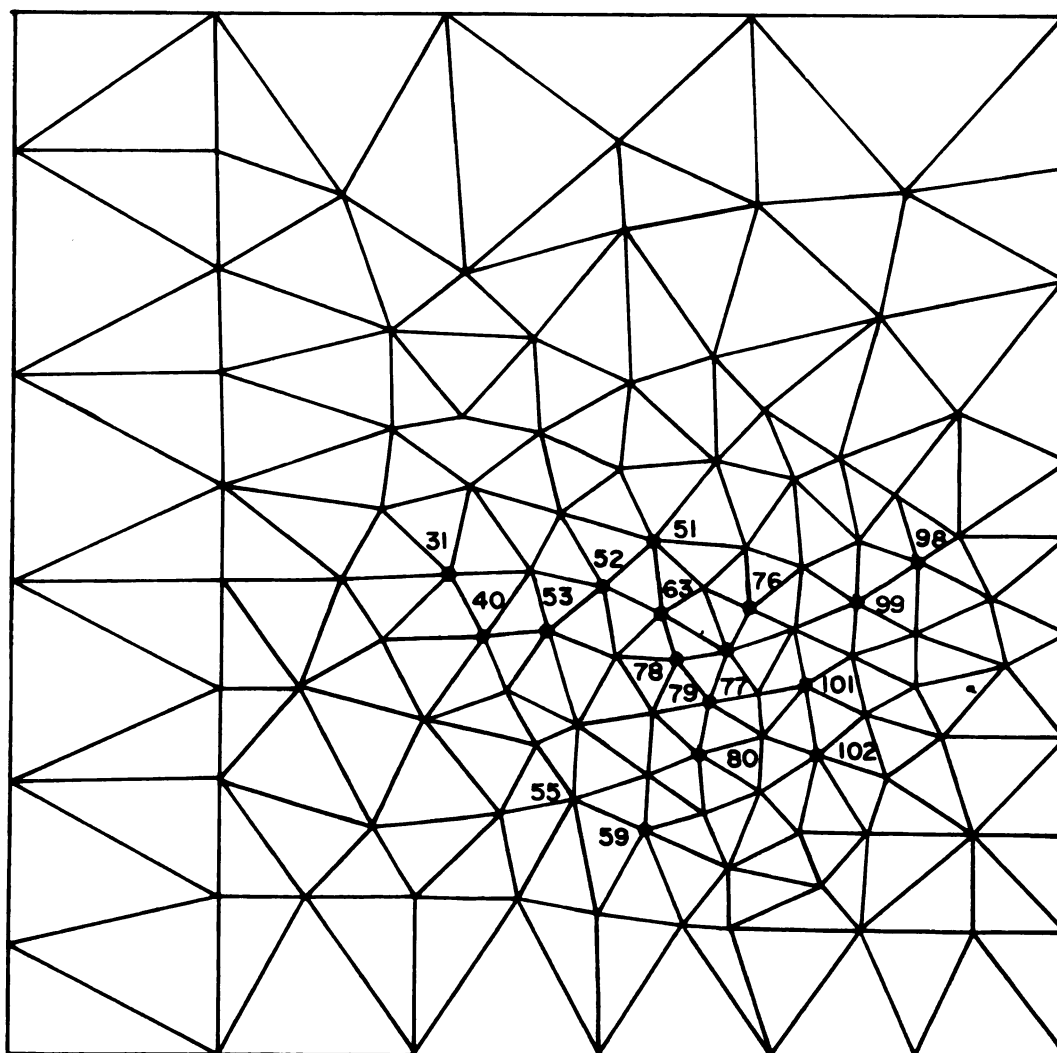
BIBLIOGRAPHY

- Desaulniers, D. E., Cherry, J. A., and Fritz, P., 1981. Origin, Age and Movement of Pore Water in Argillaceous Quaternary Deposits at Four Sites in Southwestern Ontario. *Journal of Hydrology* 50, p.231-257.
- Firouzian, A., 1963. Hydrological Studies of the Saginaw Formation in the Lansing, Michigan Area - 1962. Michigan State University, Unpub. M.S. Thesis.
- Freeze, R. A. and Cherry, J. A., 1979. *Groundwater*. Prentice-Hall, Englewood Cliffs, New Jersey.
- Grisak, G. E. and Cherry, J. A., 1975. Hydrologic Characteristics and Response of Fractured Till and Clay Confining a Shallow Aquifer. *Can. Geotech. Jnl.* 12, p.23-43.
- Haji-Djafari. S., 1982. Users' Manual, GEOFLOW Ground Water Flow and Mass Transport Computer Program. D'Appolonia Waste Management Services, Inc., Pittsburgh, Pa.
- Huffman, G. C., 1980. Ground-Water Data for Michigan, 1979. U.S.G.S. Open-File Report 80-1212.
- Kelly, W. A., 1936. The Pennsylvanian System in Michigan. Michigan Dept. of Conservation, Geol. Surv. Div. Pub. 40, Geol. Ser. 34.
- Long, D. T., Larson, G. J., and Slayton, D. S., 1981. Field Evidence for Shale Membrane Filtration. *Geol. Soc. Amer. Abstr. with Programs* 13, p.499.
- Lovato, J. and Larson, G. J., 1979. The Geohydrologic Characteristics of the Glacial Drift within the Water Quality Management Project Site, Michigan State University. Inst. of Water Research, Tech. Report No. 47.
- Monaghan, G. W. and Ritter, M. H., 1982. Identification and Verification of Natural Recharge Areas in Glaciated Terrain. *Geol. Soc. Amer. Abstr. with Programs* 15, no. 7.
- Neuman, S. P., 1973. Calibration of Distributed Parameter Groundwater Flow Models Viewed as Multiple-Objective Decision Process under Uncertainty. *Water Resource Res.* 9, p.1006-1021.
- Norris, S. E., 1962. Permeability of Glacial Till. *Geol. Sur. Prof. Paper* 450-E, Short Papers in Geology, Hydrology, Topography Art.224, p.E150-E151.

- Stuart, W. T., 1945. Progress Report on the Ground-water Resources of the Lansing Area, Michigan. Mich. Geol. Sur. Prog. Rept. 13.
- Vanlier, K. E. and Wheeler, M. L., 1968. Analog Simulation of Ground-Water Development of the Saginaw Formation, Lansing, Michigan area. Prepared by the U.S.G.S. in cooperation with Tri-County Planning Commission and Mich. Geol. Sur.
- Vanlier, K. E., Wood, W. W., and Brunett, J. O., 1973. Water-Supply Development and Management Alternatives for Clinton, Eaton, and Ingham Counties, Michigan. Geol. Sur. Water-Supply Paper 1969.
- Van Til, R. L., 1977. An Analysis of Hydrologic Simulation and Groundwater Withdrawal Patterns in Eastern Lansing, Michigan. Michigan State Univ., Unpub. M.S. Thesis.
- Walton, W. C., 1960. Leaky Artesian Aquifer Conditions in Illinois. Illinois State Water Survey Rept. of Invest. 39.
- Wheeler, M. L., 1967. Electric Analog Model Study of the Hydrology of the Saginaw Formation in the Lansing, Michigan Area. Michigan State Univ., Unpub. M.S. Thesis.

APPENDIX

Figure 6. Finite Element Grid with Pumping Centers.
Darkened circles represent pumping nodes.



51
TABLE 1
PUMPING SCHEDULES

**1945 STEADY STATE
PUMPING SCHEDULE***

<u>NODE</u>	<u>AMOUNT PUMPED</u>
52	6.55×10^5
63	5.48×10^5
77	1.87×10^5
78	2.94×10^5
76	2.67×10^4
99	6.68×10^4
101	9.36×10^4

**1945 - 1964 TRANSIENT ANALYSIS
PUMPING SCHEDULE***

<u>NODE</u>	<u>1945-1949</u>	<u>1950-1954</u>	<u>1955-1959</u>	<u>1960-1964</u>
52	6.95×10^5	6.82×10^5	6.15×10^5	6.15×10^5
63	6.95×10^5	6.28×10^5	4.52×10^5	4.41×10^5
78	2.27×10^5	2.91×10^5	3.88×10^5	4.81×10^5
77	2.27×10^5	2.27×10^5	2.94×10^5	2.94×10^5
51	0	2.46×10^5	4.01×10^5	6.42×10^5
76	2.67×10^4	5.35×10^4	6.68×10^4	5.35×10^4
99	6.68×10^4	9.36×10^4	1.60×10^5	2.14×10^5
101	1.47×10^5	2.54×10^5	6.68×10^4	6.68×10^4
102	0	0	2.41×10^5	3.21×10^5
53	4.01×10^4	1.20×10^5	1.47×10^5	1.74×10^5

52
TABLE 1 (Continued)

**1964 - 1979 TRANSIENT ANALYSIS
PUMPING SCHEDULE***

NODE	1956-1963	1964-1971	1972-1979
52	5.82×10^5	6.02×10^5	5.07×10^5
63	4.74×10^5	5.40×10^5	5.32×10^5
78	4.57×10^5	5.75×10^5	5.52×10^5
77	3.18×10^5	2.81×10^5	3.26×10^5
51	4.99×10^5	6.68×10^5	5.59×10^5
76	5.35×10^4	6.68×10^4	0
99	1.86×10^5	2.54×10^5	3.68×10^5
101	6.68×10^4	8.02×10^4	1.66×10^5
102	2.86×10^5	4.01×10^5	4.49×10^5
53	1.75×10^5	2.00×10^5	0
98	0	0	2.19×10^5
79	0	0	3.44×10^5
80	0	0	5.37×10^5
55	0	0	1.02×10^5
59	0	0	3.22×10^5
31	0	0	2.46×10^5
40	0	0	6.42×10^4

* Values in ft^3/day

SAMPLE COMPUTER RUN

המחברת מודה כי היא לא יודעת להעריך את ההשפעה של התהליך הזה על התוצאות, ולכן היא לא יכולה להעריך את ההשפעה של התהליך הזה על התוצאות. היא מודה כי היא לא יודעת להעריך את ההשפעה של התהליך הזה על התוצאות, ולכן היא לא יכולה להעריך את ההשפעה של התהליך הזה על התוצאות.

1. The first step is to identify the problem or question that needs to be answered. This involves understanding the context and the specific requirements of the task.

[illegible]

7777 666

[illegible][illegible][illegible][illegible]

() () () () () () () () () () () () () () () ()

.....

.....

.....

.....

.....

.....

.....

.....

.....

***** MODAL DATA *****
MODAL COORDINATE DATA WILL BE MULTIPLIED BY AFAC1 = +961.00 AND YFACT = +99.100
MODAL TYPE X Y H140
1 1 0.57 0.75000 POTENTIAL

POTENTIAL
POTENTIAL
POTENTIAL
POTENTIAL
POTENTIAL

SOURCE
SOURCE

POTENTIAL
87

POTENTIAL
75

SOURCE
87

POTENTIAL
76

SOURCE
SOURCE
SOURCE

POTENTIAL
87

SOURCE
SOURCE

POTENTIAL
77
80

POTENTIAL
77
80

SOURCE
SOURCE
SOURCE

[illegible]

1-7-79 1-8-79 1-9-79 1-10-79 1-11-79
1-12-79 2-1-80 2-2-80 2-3-80 2-4-80

STEADY-STATE SOLUTION

THE VALUES OF PIEZOMETRIC HEADS AT TIME $t =$

၂၀၀၆

THE UNIVERSITY OF CHICAGO PRESS

757 748 679 1437
878 887 887 888 888
.....
197 197 197 197 197
175 445 67 888 888

DISCHARGE VECTORS PRINTED AT TIME = 0.00

DISCHARGE VECTORS ARE CALCULATED AT THE CENTER OF ELEMENT
THEY ARE THE ANGLE THAT TOTAL VELOCITY MAKES WITH POSITIVE
ELEMENT NO. 9X TOTAL 9Y

X-COORDINATE
THETA

X-CORD

Y-COOK

12345678910111213141516171819202122232425

DISCHARGE VECTORS PRINTED AT TIME = 0760.00

DISCHARGE VECTORS ARE CALCULATED AT THE CENTER OF ELEMENT
THETA IS THE ANGLE THAT TOTAL VELOCITY MAKES WITH
ELEMENT NO. UK X-COORDINATE Y-COORD

14:17:14 04/15/83 TH16302 1162 LIMS PRINT.
17 PAGES PRINT. COST AT R62 IS \$ 1.05

14:17:14 04/15/83 TH16302 1162 LIMS PRINT.
17 PAGES PRINT. COST AT R62 IS \$ 1.05

14:17:14 04/15/83 TH16302 1162 LIMS PRINT.
17 PAGES PRINT. COST AT R62 IS \$ 1.05

14:17:14 04/15/83 TH16302 1162 LIMS PRINT.
17 PAGES PRINT. COST AT R62 IS \$ 1.05

14:17:14 04/15/83 TH16302 1162 LIMS PRINT.
17 PAGES PRINT. COST AT R62 IS \$ 1.05

14:17:14 04/15/83 TH16302 1162 LIMS PRINT.
17 PAGES PRINT. COST AT R62 IS \$ 1.05

14:17:14 04/15/83 TH16302 1162 LIMS PRINT.
17 PAGES PRINT. COST AT R62 IS \$ 1.05

14:17:14 04/15/83 TH16302 1162 LIMS PRINT.
17 PAGES PRINT. COST AT R62 IS \$ 1.05

MICHIGAN STATE UNIV. LIBRARIES



31293106817368