

The University of Wisconsin Library
Manuscript Theses

Unpublished theses submitted for the Master's and Doctor's degrees and deposited in The University of Wisconsin Library are open for inspection, but are to be used only with due regard to the rights of the authors. Bibliographical references may be noted, but passages may be copied only with the permission of the authors, and proper credit must be given in subsequent written or published work. Extensive copying or publication of the thesis in whole or in part requires also the consent of the Dean of the Graduate School of The University of Wisconsin.

This thesis by DIDIER FRANZ EDGAR PENNEQUIN
has been used by the following persons, whose signatures attest their acceptance of the above restrictions.

A Library which borrows this thesis for use by its patrons is expected to secure the signature of each user.

NAME AND ADDRESS

DATE

GROUNDWATER CIRCULATION AND GROUNDWATER BUDGET
FOR LAKE WINGRA, MADISON, WISCONSIN

BY

DIDIER FRANZ EDGAR PENNEQUIN

A thesis submitted in partial fulfillment of the
requirements for the degree of

MASTER OF SCIENCE

(Geology)

at the

UNIVERSITY OF WISCONSIN-MADISON

1982

AWO
P4125
D535

ABSTRACT

The groundwater circulation pattern, as well as seasonal variations in groundwater levels and a groundwater budget were established for the late spring and summer of 1981, for one of the Madison lakes: Lake Wingra. Mini-piezometers and conventional piezometers, emplaced both in the lakebed and in the shore sediments, showed that groundwater generally flows into the lake from the west, north, and south shores, and leaves the lake through the east shore. The groundwater seepage pattern in and out of the lake is not compatible with the seepage pattern produced by the model developed by McBride and Pfannkuch (1975), partly because the groundwater system in the Lake Wingra area has been modified by urbanization. Daily and weekly water level measurements showed that the overall configuration of the groundwater system around and below the lake is stable. However, the Lake Wingra hydrogeologic system is permanently in a transient state in that short-term fluctuations of the groundwater levels are controlled by precipitation. The importance of precipitation on fluctuations of water levels is evident from a visual inspection of water level records and rainfall data, and was also demonstrated with a computer program that calculated and correlated the antecedent precipitation index with groundwater levels. Long-term fluctuations of the groundwater levels are probably caused at least in part, by pumping of the deep sandstone aquifer, but more data are needed to confirm this.

The groundwater budget for the lake for August and for the summer

was established using mini-piezometers, in-lake and onshore piezometers, and a standard Darcy's law approach. A finite-difference cross-sectional model was used to estimate hydrogeologic and geologic parameters needed to calculate the groundwater budget for the lake. Groundwater was found to seep into the lake at an average rate of about $0.3 \text{ ft}^3/\text{s}$ (both for August and the whole summer), and out of the lake at rates varying from $0.08 \text{ ft}^3/\text{s}$ (August) to $0.09 \text{ ft}^3/\text{s}$ (summer). These new in- and out-seepage rates are significantly lower than those estimated by Oakes et al. (1975) and Novitzki and Holmstrom (1979), but are believed to be more accurate because of the more numerous field data available for use in this study.

ACKNOWLEDGMENTS

This research was funded by the Department of the Interior, Office of Water Research and Technology, project # 14-34-0001-1153. This assistance was greatly appreciated.

I would like to thank my advisor Dr. Mary P. Anderson for having initiated the project, allowing the free use of her auditron and typewriter, and especially for the careful and excellent editing job she has done for this thesis.

Dr. Charlie Byers of my committee is thanked for his constructive comments and being so flexible in accepting to receive my thesis only two days before my defense, and Dr. Ken Potter is thanked for the valuable suggestions he gave me regarding the use of the antecedent precipitation index.

Harley Young of the U.S.G.S. is thanked for making accessible field data for the Lake Wingra area. Director Katharine Bradley of the University of Wisconsin-Madison Arboretum, Manager Randy Smith of the Nakoma golf course, Sister Jeannette Feldballe, Chairperson of the Biology Dept. at Edgewood College, and Superintendent Dan Stapay of the Madison Parks Department are thanked for their kind cooperation and having allowed me to use their property to install piezometers. Water Resources Supervisor Larry Deibert of the Madison Water Utility is thanked for having made accessible pumping data. Ilona Loser is thanked for having typed some of the worst tables of Appendix D.

I would like to thank all my fellow students for being such a nice

group to work with, including Joe Yelderman for introducing me to Texas humor, which I'll never understand. Ken Bradbury is especially thanked for the help and company he provided in the field, and also for the valuable advice he gave me regarding field methods. The entire Water Resources crew is thanked for their kind assistance and Kari Sherman is especially thanked for her always readiness to help, whether it was to reserve a truck or to answer questions regarding the word processor.

Last, but not least, I would like to particularly thank two people: my father, Albert E. Pennequin for the very much needed help he gave me with drafting the final figures and with surveying all the wells and gauges, and Béatrice Nicot for the great help she provided nights after nights with typing and retyping this thesis, and even more, for her patience and understanding during a most difficult period. Merci Béa.

TABLE OF CONTENTS

	PAGE
-ABSTRACT	i
-ACKNOWLEDGMENTS	iii
-TABLE OF CONTENTS	v
-LIST OF FIGURES	viii
-LIST OF TABLES	xi
-SYMBOL INDEX	xii
-I. INTRODUCTION	1
-Purpose and Scope	1
-Geographic Location	3
-Geomorphology, Geology and Bathymetry of Lake Wingra and Surrounding Areas	3
-Geomorphology and surficial geology	3
-Bedrock geology	11
-Bathymetry	11
-General Hydrogeology and Preliminary Groundwater Budgets Established in Previous Studies.	13
-General hydrogeology	13
-Previous groundwater budgets.	16
-II. GROUNDWATER MONITORING NETWORK AND FIELD METHODS	19
-Main Network	19
-Onshore piezometers	19
-In-lake piezometers	23
-Automatic water level recorder, spilling dam, Staff and rain gauges	24
-Surveying	25
-Temporary Network and Springs	26
-Seepage meters	26
-Mini-piezometers	32
-Spring discharges	34
-III. FIELD DATA AND DISCUSSION OF THE VARIOUS GROUNDWATER MONITORING METHODS USED	39
-Winter Data; In-Lake Piezometers	39
-Summer Data	41
-Onshore and in-lake piezometers	41
-Staff gauges	44
-Automatic water level recorder	44
-Precipitation	44
-Seepage meters	45
-Mini-piezometers	47
-Spring and seep discharges	48

TABLE OF CONTENTS (cont.)

	PAGE
-IV. THREE DIMENSIONAL SHALLOW GROUNDWATER CIRCULATION AROUND AND BELOW LAKE WINGRA	50
-Introduction	50
-Groundwater Circulation	51
-Overview	51
-East	62
-West	64
-South	69
-North	80
-Lake	83
-Low water tables	83
-Causes for the Short Term Fluctuations of the Groundwater Levels: Precipitation	86
-Introduction	86
-Problems in correlating groundwater levels with rainfall in the absence of continuous records.	92
-Antecedent precipitation index	93
-Alpha	95
-Computer program and initial antecedent precipitation index, A_0	98
-Results and discussion	104
-Causes for the Long Term Variations of the Groundwater Levels: Large Scale Pumping of the City Wells.	110
-Conclusions	111
 -V. GROUNDWATER FLOW MODEL	 114
-Introduction	114
-Purpose and Scope	115
-Model	116
-Grid	117
-Boundary conditions	117
-Steady state calibration	120
-Results and Discussion	122
-Conclusion	129
 -VI. GROUNDWATER BUDGET	 130
-Introduction	130
-Method	131
-Groundwater Inflow	132
-Shore segments and lake sub-areas	132
-Areas of seepage	132
-Hydraulic gradients	137

TABLE OF CONTENTS (cont.)

	PAGE
-Hydraulic conductivities	139
-Groundwater Outflow	141
-Shore segments and lake sub-areas	141
-Areas of seepage	141
-Hydraulic gradients	142
-Hydraulic conductivities	143
-Results and Discussion	143
-Conclusion	153
-VII. CONCLUSIONS AND RECOMMENDATIONS	155
-REFERENCES CITED	162
-APPENDIX A	166
-APPENDIX B	175
-APPENDIX C	178
-APPENDIX D	187
-APPENDIX E	231
-APPENDIX F	235
-APPENDIX G	241

LIST OF FIGURES

FIGURE	TITLE	PAGE
1.	Maps showing the location of Lake Wingra in Madison, Wisconsin.	4
2.	Contour map of the bedrock geology in the Lake Wingra basin.	6
3.	Surficial geology of the Lake Wingra area and legend for Figure 3.	8
4.	General groundwater flow in the Lake Wingra basin.	14
5.	Main network used for groundwater monitoring in and around Lake Wingra.	20
6.	Temporary and accessory network used for groundwater monitoring in Lake Wingra, and spring locations.	27
7.	Details of a seepage meter.	30
8.	Mini-piezometer emplacement technique.	30
9.	Velocity-area technique used to measure spring discharge.	35
10.	Mariotte bottle and tracer dilution technique used to measure spring discharge.	35
11.	Average shallow groundwater flow around Lake Wingra during September 1981.	52
12.	Map summarizing the areas of groundwater in- and out-seepage for Lake Wingra.	54
13.	Mean head distribution along cross section A-A' for September 1981.	56
14.	Mean head distribution along cross section B-B' for September 1981.	56
15.	Mean head distribution along cross section C-C' for September 1981.	58
16.	Mean head distribution along cross section D-D' for September 1981.	58

LIST OF FIGURES (cont.)

FIGURE	TITLE	PAGE
17.	Mean head distribution along cross section E-E' for September 1981.	60
18.	Decrease in vertical hydraulic gradient with distance away from shore measured in several mini-piezometers.	72
19.	Cartoon illustrating the model envisioned for the south and north central shores of Lake Wingra.	78
20.	Cartoon representing the groundwater circulation modified by plants in parts of the northwest, west and southwest shores of Lake Wingra during an extended dry period.	84
21.	Daily precipitation and automatic water level recorder charts plotted in 6 hour-increments for: a. June, b. July, c. August and d. September.	87
22.	Graphs showing the effects that different alphas have on the correlation coefficients.	96
23.	Antecedent precipitation index computed with four different A_0 -values (2,3,4 and 5 inches) for an α -value of .80.	100
24.	Antecedent precipitation index computed with four different A_0 -values (2,3,4 and 5 inches) for an α -value of .98.	102
25.	Groundwater circulation around Lake Wingra for August 1981, and location of cross section B-B'.	118
26.	Finite-difference grid used for cross section B-B'.	123
27.	Comparison between the results of the computer simulation and field observations for August in cross section B-B'.	123

LIST OF FIGURES (cont.)

FIGURE	TITLE	PAGE
28.	Map showing the Lake sub-areas and shore segments used to calculate the groundwater budget for Lake Wingra, for August and for the period July through September, 1981.	133

LIST OF TABLES

NUMBER	TITLE	PAGE
1.	Generalized stratigraphy and aquifer system in the Lake Wingra area, from Oakes et al. (1975).	12
2.	Mean monthly vertical hydraulic gradients observed in and around Lake Wingra during the summer of 1981.	65
3.	Comparison of the winter and summer groundwater levels in some eastern and western in-lake piezometers.	68
4.	Comparison between the vertical hydraulic gradients observed in both the shallower onshore piezometers close to the shoreline and in the adjacent off-shore mini-piezometers, for August 19 and 25, 1981.	71
5.	Optimum correlation coefficients with their associated alpha-values obtained when the groundwater levels are regressed against the antecedent precipitation index.	105
6.	Comparison between hydraulic heads predicted by the model for August 1981 and the August heads observed in the field for the 12 piezometers included in cross section B-B'.	125
7.	Horizontal hydraulic conductivity (K_h) and ratio of vertical to horizontal hydraulic conductivities ($K_v:K_h$) used for each geologic layer in the calibrated model.	126
8.	Parameters and gradients used to calculate the groundwater budget for August 1981, and results.	144
9.	Parameters and gradients used to calculate the groundwater budget for the three-month period extending from July through September 1981, and results.	145
10.	Comparison between the groundwater seepage rates in and out of Lake Wingra obtained in this study and in two previous studies.	151

SYMBOL INDEX

Piezometers

OS = Onshore site.

IL = In-lake site.

Sh = Shallow piezometer.

ShI = Shallow-intermediate piezometer. This is the shallower of the intermediate piezometers at any piezometer nest where more than one intermediate depth piezometer exists.

I = Intermediate piezometer.

DI = Deep-intermediate piezometer. This is the deeper of the intermediate piezometers at any piezometer nest where more than one intermediate depth piezometer exists.

D = Deep piezometer.

Examples

OS-1 = Onshore site #1.

OS-1Sh = Shallow piezometer at onshore site #1.

IL-2D = Deep piezometer at in-lake site #2.

Seepage meters

S = Seepage meter station.

S1 = Seepage meter station #1.

1-8 = Seepage meters : 1 closest to shoreline, 8 farthest from shoreline.

Example

S3-4 = 4th seepage meter from shoreline at seepage meter station #3.

Mini-piezometers

MP = Mini-piezometer station.

1-10 = Mini-piezometer station #.

a-d = Mini-piezometers; a, closest to shoreline, d, farthest from shoreline.

Example

MP-6a = Mini-piezometer "a" at mini-piezometer station #6.

Springs

Sp = Spring.

1-8 = Spring #.

a,b = Used if a spring consists of two sub-springs

Example

Sp-la = Sub-spring "a" of spring #1.

I. INTRODUCTION

Purpose and Scope

The objective of this study is (1) to determine in detail the groundwater circulation around and below Lake Wingra, (2) to analyse and determine the causes of fluctuations of the groundwater levels in the Wingra area, and (3) to quantify the amount of groundwater entering and leaving the lake. Lake Wingra is a shallow, glacial lake atypical in the sense that its geology is different from the geology of most shallow glacial lakes that dot much of the mid-west and other northern areas; for example, Lake Wingra has no true lake sediments, reflecting the fact that before being dredged and formed into its present-day configuration, Lake Wingra was a fairly extensive marsh.

Lake Wingra is also a lake enclosed within a city (Madison, Wisconsin), and so the effects of urbanization on the groundwater movement near and below the lake, particularly as it relates to the nutrient budget are of interest to both hydrogeologists and biologists. Furthermore, the unusual groundwater seepage pattern will interest hydrogeologists. Biologists who study zoo- and phytoplankton communities in the lake will need to know the quantity of groundwater seepage. Botanists who study the plant distribution along the lake

shore in the University of Wisconsin Arboretum, are interested in knowing the groundwater circulation pattern, the amount of groundwater seepage and the proximity of the water table to the ground surface, especially now that relationships between the biosphere and the hydrosphere have been widely recognized (Meyboom, 1966, Hasler, 1975).

The study began in December 1980, and ended at the end of September 1981. Most of the data were gathered during the summer study (May 15 to September 30, 1981). In-lake and onshore piezometers, mini-piezometers, seepage meters, staff gauges (to measure lake level and lake discharge into Murphy Creek) and an automatic water level recorder were emplaced during the winter, spring and early summer 1981. Except for the seepage meters and for most of the mini-piezometers, roughly weekly measurements were taken at each piezometer and gauge. A rain gauge gave daily precipitation and spring discharges were measured in August.

The detailed groundwater circulation, its variation and the fluctuations of the groundwater potentials in the Lake Wingra area are described and analyzed. A two-dimensional groundwater flow computer model was constructed to obtain vital information to calculate the groundwater budget for the lake. The groundwater budget for the lake is determined and compared with previous estimates of the groundwater budget done by earlier workers (Oakes et al., 1975, Novitzki and Holmstrom, 1979).

Geographic Location

Lake Wingra is located in south central Wisconsin within the limits of the City of Madison (see Fig. 1). It is the smallest and shallowest of the lakes enclosed or bordering the City of Madison.

Geomorphology, Geology and Bathymetry of Lake Wingra and Surrounding Areas.

Geomorphology and surficial geology

Lake Wingra and the adjacent marshes occupy a depression in thick lake, marsh and glacial deposits that fill a preglacial valley trending roughly northeast-southwest as shown by Figure 2 (Oakes et al., 1975). Most of the north-northwest and south-southwest shorelines of the lake consists of low hills controlled mainly by the bedrock valley wall topography. The northeast, east, southeast and west shorelines of the lake consists of relatively flat marshy areas. The unconsolidated surficial deposits in the area average about 18 meters and range between zero to more than 46 meters in thickness. Before the turn of the century, most of the upper surficial deposits consisted of marsh sediments, but since then, a considerable amount of fill material has replaced some to nearly all of the old marsh sediments (e.g. Vilas Park,

Figure 1 : Maps showing the location of Lake Wingra in
Madison, Wisconsin.

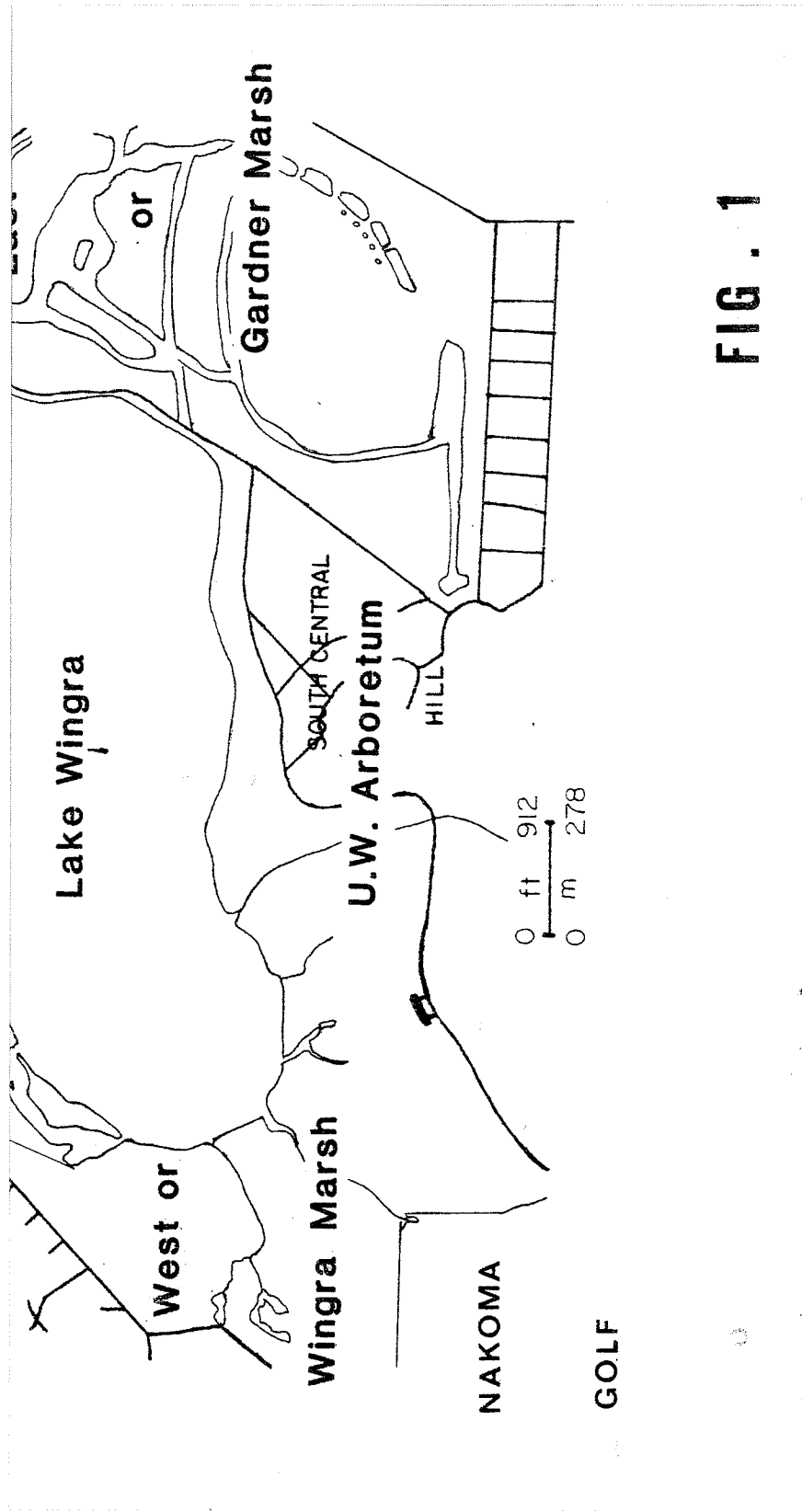
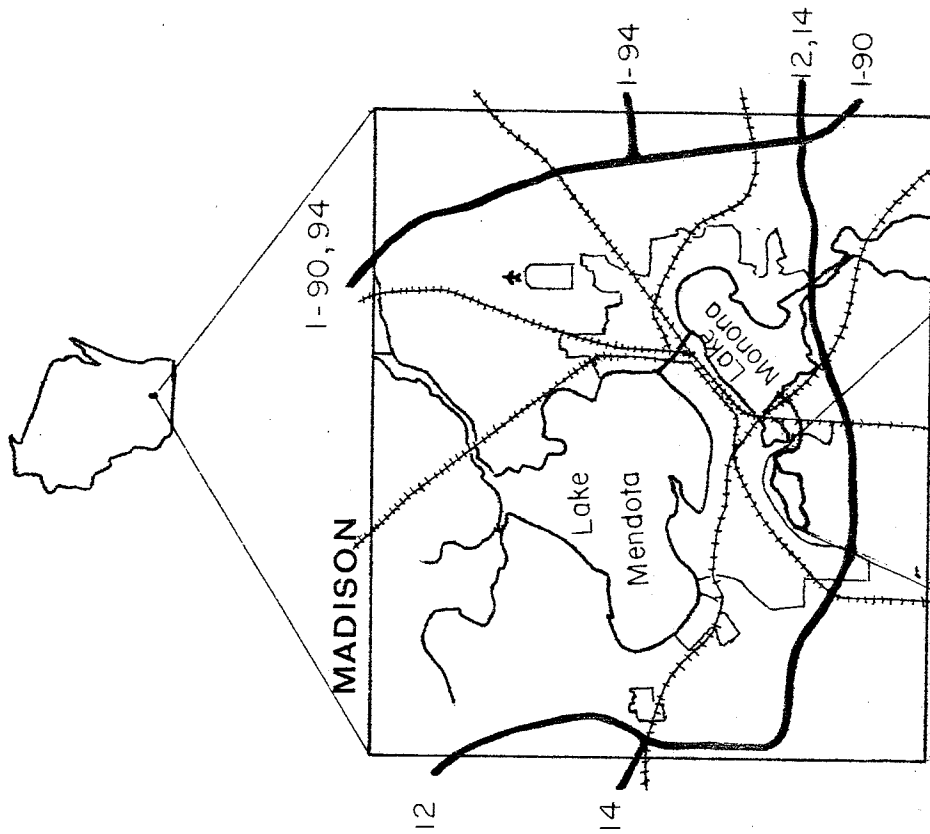
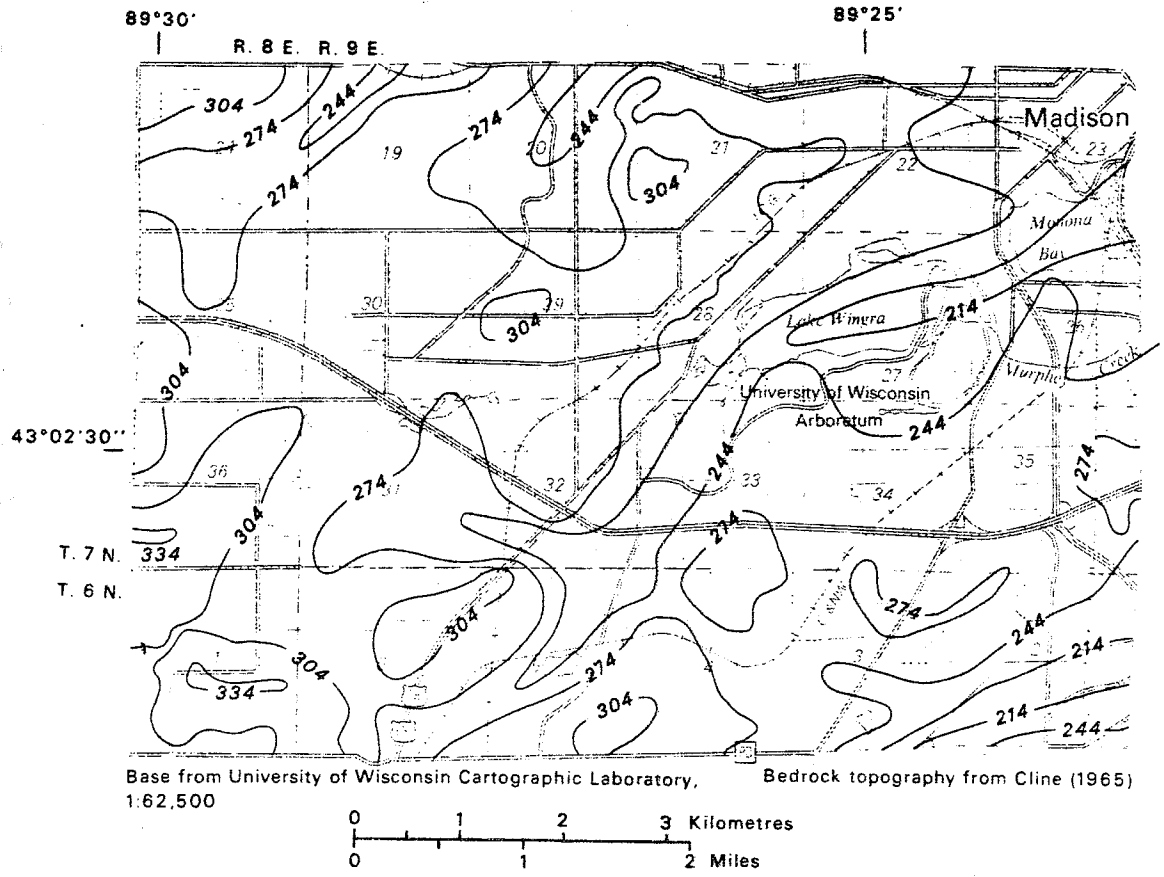


FIG . 1





EXPLANATION

— 274 — Bedrock contour
 Shows altitude of bedrock surface. Contour interval 30 metres (100 feet). Datum is mean sea level

Figure 2 : Contour map of the bedrock geology in the Lake Wingra basin, from Oakes et al. (1975).

Wingra Park) in the north, south and parts of the east shore (Baumann et al., 1974).

Due to the heterogeneous character of these deposits and the paucity of reliable data, it is difficult to detect definite trends in the present surficial geology. However, well logs obtained from Huff and Young (1980) and during this study (see Fig. 3) suggest that the Lake Wingra shore and adjacent area has five different geologic settings. Figure 3 highlights the surficial geology around the lake. Each of the five areas is discussed below.

(1) The west and southwest marshy areas typically consist of a blanket of fine grained material, usually 13 to 18 meters thick; clayey silt or silty clay capped by a thin veneer of peat predominates, but clay lenses and fine sand are locally present. Fine and medium sand underlies this upper layer of fine grained material. Bedrock is usually located 15 to 45 meters below the ground surface. Huff and Young (1980) reported that, in at least one place, the sand layer was within a few meters of the ground surface, which could possibly indicate the presence of a buried channel (Young, personal communication, 1981). The east shoreline of the lake resembles the west and southwest shorelines in the sense that it too is dominated by marshes and similar fine grained marsh deposits which overlie coarser grained material. The lack of deep well logs for the eastern area prevented direct determination of the thickness of the fine grained sediments, but the log of the deepest eastern piezometer combined with logs from deep wells located one kilometer or so away, suggest that sand is present at a depth greater

LEGEND





















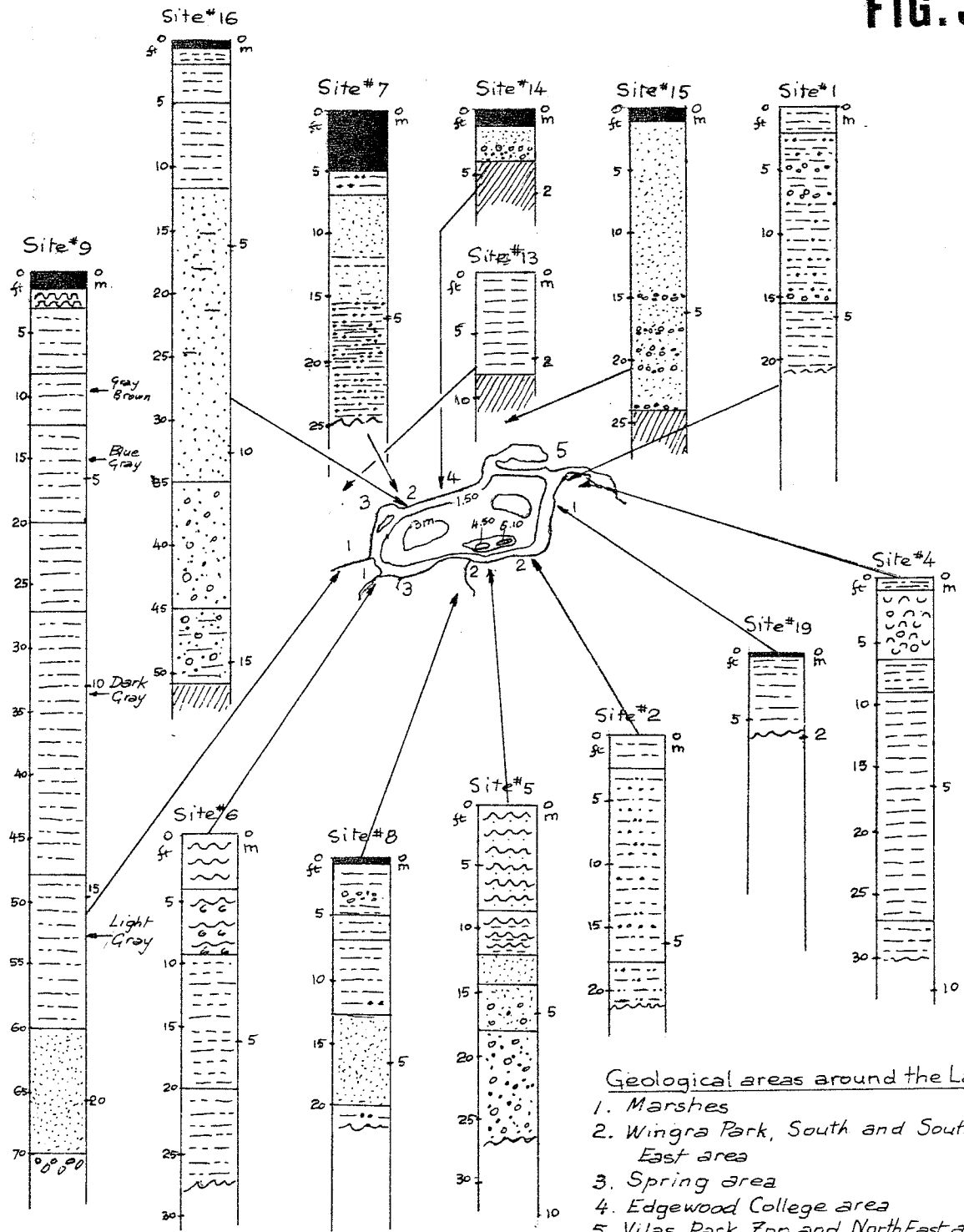
	Soil
	Peat
	Shelly Peat
	Silty Peat
	Sandy Peat
	Artificial Fill
	Clay
	Silt
	Silty Clay
	Clayey Silt
	Fine Sand
	Medium Sand
	Coarse Sand
	Gravel
	Sandy Silt + Clay Lenses
	Clayey Sand + Clay Lenses
	Clayey Sand + Sand Lenses
	Sand + Gravel
	Sand + Gravel + Clay
	Bedrock

Figure 3 : Surficial geology of the Lake Wingra area (next page) and legend for Figure 3. The bathymetry is obtained from this study and from Sale and Helgeland (1961). Logs are obtained from this study, except for the log of site #9, which is obtained from Huff and Young (1980). The geologic subdivision is based on these and additional logs obtained during this study.

FIG. 3



Geological areas around the Lake

- 1. Marshes
- 2. Wingra Park, South and South East area
- 3. Spring area
- 4. Edgewood College area
- 5. Vilas Park, Zoo and North East area

than 10 m below the ground surface, and that bedrock is somewhere between 40 and 50 m below the surface.

Although naturally very similar, there are two basic differences between the east and west marsh. First, the latter lies generally 0.5 to 3 m above the lake surface, while a good portion of the former is below, and is separated from the lake by the Arboretum drive which serves as a permeable dam. Secondly, the northern and western periphery of the east marsh has been considerably altered by the emplacement of artificial fill material, in a futile attempt to build a subdivision and a major road several decades ago; part of this fill was used later for the Arboretum drive.

(2) The second geologic setting includes Wingra Park and the adjacent arboretum area on the west, along with the south and southeast shores of the lake. These shores are characterized by coarser grained material relative to the marshy areas; sandy clays, silt and sand are the dominant sedimentary textures. Clay lenses and gravel are also locally abundant. There is however, a 2 to 3 m thick peat, soil, silty clay and clayey silt layer overlying these coarser grained sediments. Bedrock is located generally 15 to 40 m below the ground surface. In the Wingra Park area, two drilled holes suggest that the bedrock rises from 15 m, near the lake, to 3 or 4 m below the ground surface, 200 m away from the lake.

(3) The geology in the two spring areas adjacent to the lake consists of a mixture of sandstone bedrock and sediments.

(4) The Edgewood College area constitutes the fourth geologic

setting and is dominated by bedrock which is capped in most places, by a thin veneer of soil, fine sand and gravel.

(5) The last area is quite different, in that relatively pure fine sand is the dominant texture; in general, 7 m (in the west) to possibly 30 m (in the east) of medium sand and gravel rests directly over the bedrock in Vilas Park and Zoo area. There are some lenses of finer grained sediments interbedded with the sand.

The geology of the lake bottom itself is more difficult to study. A few lake bottom samples suggest that organic matter, muck, silt and some clay constitute the lake bottom. Locally, at the mouth of some storm sewers, sand is present.

There are small shells (mainly gastropods) and entrapped gases in the fine grained sediments throughout the entire area.

Bedrock geology

The bedrock geology consists of mostly Ordovician dolomites and Cambrian sandstones which overlie Precambrian crystalline rocks (see Table 1). For additional information about the geology of the area, see Cline (1965) and Cullen et al. (1972).

Bathymetry

Lake Wingra is a shallow eutrophic lake. Most of its bottom lies between 0.6 and 3.7 m below the water surface. There are only a couple

Table I : Generalized stratigraphy and aquifer system in the Lake Wingra area, from Oakes et al. (1975)

Era or System	Geologic unit	Dominant lithology	Subsurface hydrologic unit	Saturated thickness (m)
QUATERNARY	Holocene and Pleistocene glacial deposits	Clay, silt, sand		
	Galena Dolomite	Dolomite		
ORDOVICIAN	Decorah Formation	Dolomite		
	Platteville Formation	Dolomite		
	St. Peter Sandstone	Sandstone	Upper aquifer	15-137
	Prairie du Chien Group	Dolomite		
CAMBRIAN	Trempealeau Formation	Sandstone and dolomite		
	St. Lawrence Member			
	Ironton Sandstone Member	Sandstone		
	Galesville Sandstone	Sandstone	Sandstone aquifer	137-274
	Eau Claire Sandstone	Sandstone and shale		
Mount Simon Sandstone	Sandstone			
PRECAMBRIAN	Precambrian rocks undifferentiated	Crystalline rocks	Not an aquifer	

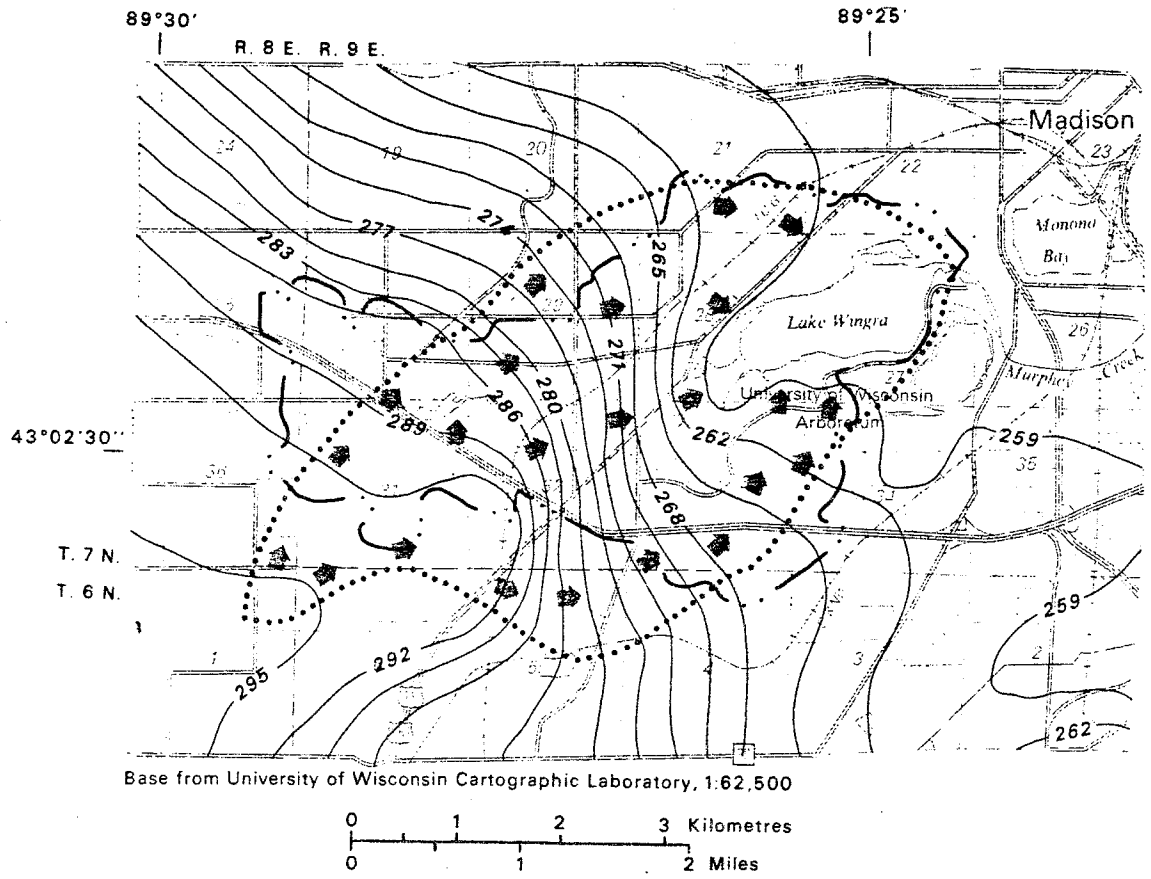
of holes reaching 4.5 and 6.1 m in depth, along the south shore (see Fig. 3).

General Hydrogeology and Preliminary Groundwater Budgets Established in Previous Studies.

In the early seventies, several field investigations of the surface and groundwater in the Lake Wingra area were undertaken by the United State Geological Survey (USGS), the Wisconsin State Geological and Natural History Survey (WGNHS) and, the Department of Water Chemistry of the University of Wisconsin, Madison under the auspices of the Eastern Deciduous Forest Biome (EDFB) Studies of the International Biological Program (IBP) (Oakes et al., 1975, Novitzki and Holmstrom, 1979, Huff and Young, 1980, Cullen et al., 1972, and Kluesener, 1972).

General hydrogeology

According to Oakes et al. (1975) and Novitzki and Holmstrom (1979), Lake Wingra hugs the northeast margin of the Lake Wingra basin which is generally determined, at least in its northeast part, by both the surface and groundwater divides (see Fig. 4). In most of the basin, groundwater flows east and northeast toward the lake, but the flow system is more complex around the lake because of both the geometry of



EXPLANATION

- 295 — **Water-table contour**
Shows altitude of water table during September 1973. Contour interval 3 metres (10 feet). Datum is mean sea level
- · · — **Surface-water divide, as defined by storm sewers**
- ➔ **Direction of ground-water flow**
- · · · · **Ground-water divide**

Figure 4 : General groundwater flow in the Lake Wingra basin, from Oakes et al. (1975).

the system and the withdrawal of large quantities of water by the Madison Water Utility from a deep aquifer (the Cambrian sandstones).

McLeod (1975) assumed that two major aquifers underlaid the Madison area; the deep or sandstone aquifer, comprising all of the Cambrian sandstones stratigraphically below the St. Lawrence member of the Trempealeau formation, and the shallow aquifer which includes all of the consolidated and unconsolidated formations above the St. Lawrence member (see Table 1). These two aquifers are supposedly separated by a low permeability layer located near the base of the St. Lawrence. This layer is not totally impermeable because five springs emerge from near its base, which is exposed on the west side of the lake. Moreover, computer simulations show that both the deep and shallow groundwater circulation is altered by pumping from the deep sandstone aquifer (McLeod, 1975a,b).

In the absence of city pumping Oakes et al. (1975) and Novitzki and Holmstrom (1979) believe that the entire lake area would be a groundwater discharge region, whereas now the vertical flow is partially reversed and recharge from the shallow to the deep aquifer occurs in many places, half of it taking place in the thick bedrock valley beneath the lake. In any case, using the scheme developed by Born et al. (1974), Lake Wingra today resembles a flow-through lake, whereby groundwater generally seeps in through its west side and out through its east side. However, the artificial condition of the lake led Novitzki and Holmstrom (1979) to suspect that groundwater outflow probably occurs over an area larger than the normal potential seepage area of a shallow

natural lake as reported by McBride and Pfannkuch (1975).

Previous groundwater budgets

Groundwater inflow into the lake occurs mainly along its northwest, west and southwest shorelines by means of spring discharge or direct groundwater seepage. A total of ten readily visible springs or seeps, six of which have significant flow, emerge west and southwest of the lake along two arched lines roughly parallel to the walls of the buried preglacial valley (see Figure 2 and also see Figure 6 in Chapter II). In their determination of the water budget for the lake, Oakes et al. (1975) measured, estimated and calculated that the total spring discharge averaged $4,540 \text{ m}^3/\text{day}$, which represents about 37% of the total water flow to the lake for 1972. On the other hand, the direct groundwater seepage into the lake (essentially through the lake bottom) was estimated to be $1,280 \text{ m}^3/\text{day}$ or about 10% of the total water flow to the lake for that same year. This latter figure was obtained by assuming that the average vertical hydraulic gradient (0.24 m/m) based on one set of onshore nested piezometers, and the hydraulic conductivity value (0.02 m/d) derived from five shallow cores of the lake bottom, typified the entire recharge area in the lake. The area of potential groundwater seepage into the lake was estimated to be 0.26 km^2 . The combined groundwater inflow to the lake was estimated to be about $5,280 \text{ m}^3/\text{day}$ or about 47% of the total water inflow to the lake for 1972.

Continuing and extending somewhat the study started by Oakes et al.

(1975), Novitzki and Holmstrom (1979) estimated that the total average groundwater inflow (springs and direct seepage) to the lake over a period of 6 years, amounted to roughly $5,578 \text{ m}^3/\text{day}$ or 35% of the total inflow to the lake over the same period. The average spring discharge during the study period was $3,970 \text{ m}^3/\text{day}$ or 25% of the total water inflow, a figure necessarily lower than what Oakes et al. (1975) would have obtained since only the three major springs are included in the Novitzki and Holmstrom (1979) budget. On the other hand, the average direct groundwater seepage for these six years was estimated to be $1650 \text{ m}^3/\text{day}$; the first $1250 \text{ m}^3/\text{day}$ taken from Oakes et al. (1975), were assumed to be the representative average direct groundwater seepage through the bottom of the lake, and the remaining $400 \text{ m}^3/\text{day}$, taken from Huff and Young (1980), were believed to quantify the average groundwater seepage from the adjacent west and southwest marsh, to the lake.

Out-seepage from the lake into the groundwater system occurs in the southeast, east and northeast portions of the lake, and is believed to be induced by the pumping of the sandstone aquifer. The groundwater outflow term is the most ill-defined component of the water budget for Lake Wingra. Based on one set of nested piezometers showing an average vertical hydraulic gradient of -0.08 m/m , and on the assumption that recharge to the aquifer occurs over all of the lake bottom (1.26 km^2) less the 0.26 km^2 area of groundwater discharge previously mentioned, Oakes et al. (1975) estimated that on the average, the groundwater outflow in 1972 amounted to $1,620 \text{ m}^3/\text{day}$ or about 13% of the total water flow from the lake. The same hydraulic conductivity (K) used in the in-

seepage calculations was used in the groundwater out-seepage calculations.

On the other hand, Novitzki and Holmstrom (1979) assumed that most of the groundwater seepage out of the lake occurred over an arbitrary 0.39 km^2 area in its eastern part, and that this outseepage quantity was solely and directly proportional to the pumpage of the nearby Madison public water-supply wells. They estimated that during the period starting in 1972 and ending in 1977, about $603 \text{ m}^3/\text{day}$ or 4% of the total water outflow from the lake was lost to the underlying aquifer. This is considerably less than the previous estimate reported by Oakes et al. (1975).

Needless to say, the assumptions used in the water budget calculations and the paucity of the data cast some doubts on the validity of the groundwater inflow and outflow estimates.

II. GROUNDWATER MONITORING NETWORK AND FIELD METHODS

Main Network

The main groundwater monitoring network in this study consisted of several onshore and in-lake piezometers, one automatic water level recorder, two staff gauges, a weir to measure the surface outflow and one rain gauge.

Onshore piezometers

Twenty-four stations consisting of 57 piezometers located 0 to 430 m away from the lake, and extending from 1 to about 21 m below the water table, were established around the lake. Nearly all stations included a nest of two to four piezometers sunk at different depths, except where the proximity of the bedrock to the surface prevented deep drilling, as in parts of the north shore. Figure 5 shows the locations of the stations and Table A-1 gives a detailed description of the onshore piezometers.

These piezometers consisted of either 1-1/4" (34.5mm) ID PVC or steel pipe, or 3/4" (20.5mm) ID galvanized steel electrical conduit. Except for a few jetted piezometers, most are finished using a 1-foot

Figure 5 : Main network used for groundwater monitoring in and around Lake Wingra.

Legend for Figure 5 :

- ⊙ In-lake or onshore piezometer.
available for the summer study.
- In-lake piezometer destroyed by the
drifting ice.
- ⊙^R Automatic water level recorder.
- * Staff gauge.
- Ⓢ^{#1} City well unit #1.

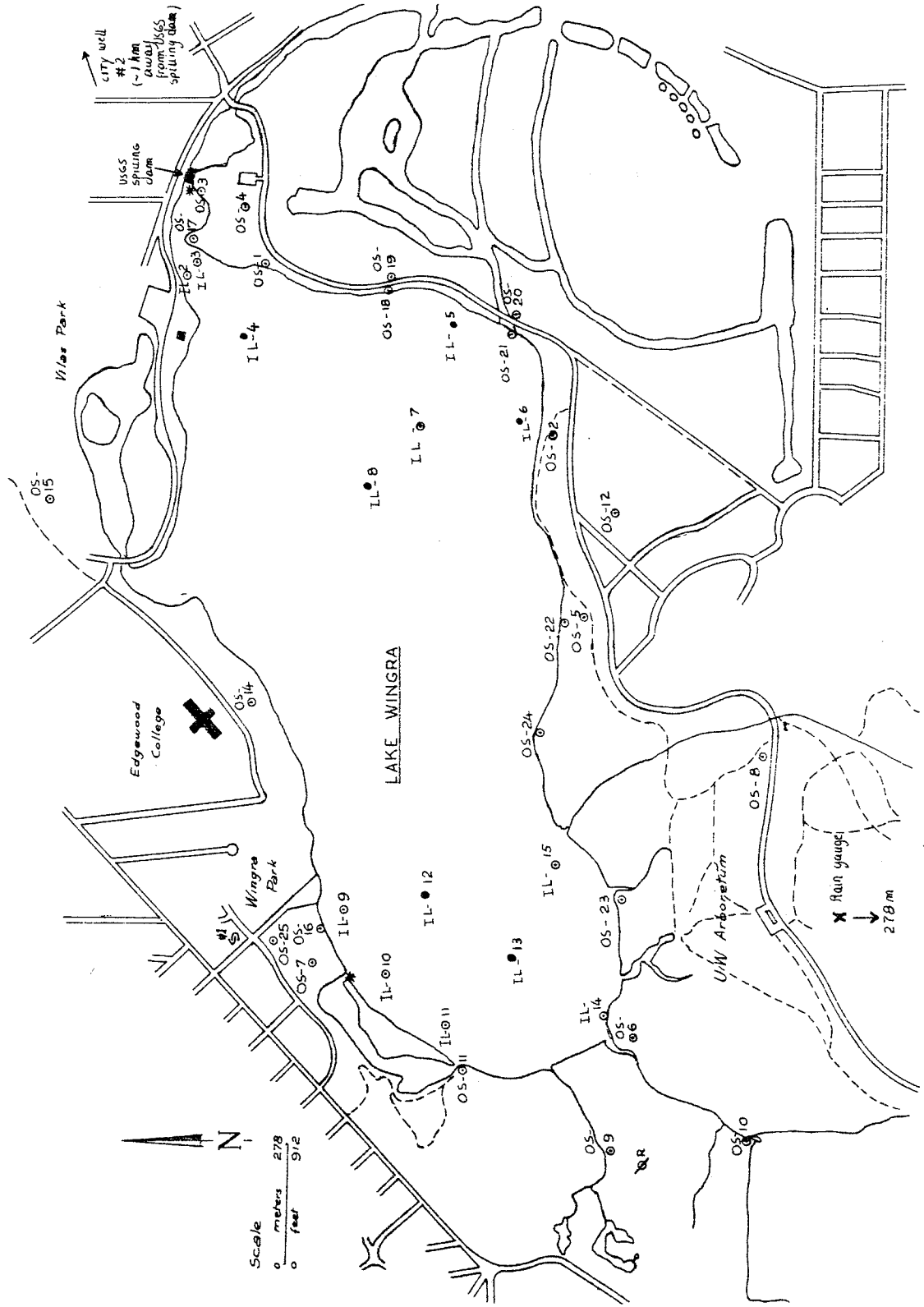


FIG. 5

long 1-1/4" or 3/4" ID PVC sand point, through which water exchange between the piezometer and the aquifer occurs.

The installation of onshore piezometers began in November 1980, and because of various mechanical failures, logistic problems and winter weather, lasted until early July. Due to hostile terrain or delicate ecosystem conditions, a truck mounted rig was used to sink only two of the onshore piezometers near Wingra Park. The remaining onshore piezometers were either installed using a mobil minute man drill with 2-3/4" (69.9mm) diameter augers, a 3" (76.2mm) diameter hand auger, a pipe pounder, a 140 GPM Homelite jetting pump or a combination of these.

Aside from the difficulties mentioned above, borehole collapse, sealing and clogging of some of the points were among the biggest problems encountered in the field. Borehole collapse, however, also helped in sealing the piezometer as will be seen in the following paragraph.

In theory, a true piezometer is supposed to measure the groundwater potential at one specific point in the aquifer. Therefore, a piezometer should have the shortest possible point or screen, and a good seal around the pipe immediately above the intake area. The seal minimizes interference from the overlying aquifer. Unfortunately, a good seal was generally difficult to achieve because of both the proximity of the water table to the land surface, and the small difference between the borehole and the pipe outside diameter. Sealing was particularly difficult for the wells emplaced with the small mobil drill and, even more so for the jetted ones. Nevertheless, a combination of borehole

collapse, pounding in of the last few feet of the pipe and downhole forcing of bentonite powder into the hole probably helped achieve an adequate seal in most cases.

Most of the piezometers were developed using a homemade surger; water was both forced down and sucked up the pipe several times, until a good connection between the well and the aquifer was achieved, or until lack of response led the author to abandon the piezometer. This lack of response was due to a rapid clogging of the point. A total of six intermediate and deep piezometers, five of which were installed by pounding, had to be disregarded early in the study. Reasons for clogging of the point include the compaction of the aquifer material around the point through forceful pounding, and the accumulation of fines in the slots of the screen. A few additional piezometers became partially clogged during the period of study; they are listed in Table A-1).

The two major observations made during the installation of the onshore piezometer network, are the presence of small scale heterogeneities and trapped gases within the sediments. The smell of hydrogen sulfide was especially noticed along the south and southwest shores of the lake.

In-lake piezometers

Piezometers installed into the lake bottom have two advantages : they show the true hydraulic gradients existing both under the lake and

between the lake and the underlying aquifer, and they are automatically nested because the lake surface in most cases is at the same elevation as the water table.

The lake piezometer network consisted of 14 sites and 17 piezometers ranging in depth from 1.5 to 5.2 m below the sediment-water interface (see Figure 5 for their location and Table A-2 in Appendix A for details on the piezometers). These piezometers were emplaced beginning in early January when the ice was sufficiently thick to support body and equipment weight. Holes were augered through the ice, and the piezometers were lowered down into these holes and pushed into the mucky lake bottom. The in-lake wells consisted solely of 1-1/4" (34.5mm) ID PVC pipe terminated by a 1-foot (304.8mm) PVC point. They were developed in the same manner as the onshore piezometers.

Unusually warm weather conditions and a premature rupture of the ice prevented the completion of the initially planned network, resulting in the absence of piezometers in the central and northern portions of the lake. During the spring ice-out, the ice engulfed and destroyed several of the in-lake piezometers (see Table A-2), further reducing the available lake network for the summer study.

Automatic water level recorder, spilling dam, staff and rain gauges

In addition to the piezometers, a Stevens type F water level recorder was used to monitor the water table fluctuations on the north side of Wingra marsh, from June 1 until October 27, when the recording

station was destroyed by vandals (see Table A-1 for more details). A spilling dam with a wier and a staff gauge previously emplaced by the USGS at the northeast side of the lake and at the head of Murphy Creek, controlled and were used to estimate most of the direct lake outflow. Another staff gauge was pounded in at the north entrance to Ho-Nee-Um pond; this gauge was used to measure lake stage. A standard 8" (203 mm) precipitation gauge, read daily by the arboretum staff, recorded rainfall at the University of Wisconsin-Madison's arboretum station (see Fig. 5 for the locations).

Surveying

Surveying was done in three steps in late April and early July 1981. Due to the extent of the network, the surface of the lake served sometimes as a plane of reference. However, closed loop surveys and a loop intersection method indicated that an accuracy of 0.01 foot (0.3 cm) was generally achieved. All elevations were calculated relative to an arbitrary datum exactly 100 feet (30.48 m) below the top of the uncapped shallow piezometer at OS-4 (see Figure 5). Due to vandalism and possibly to heavy rains, three of the shallow wells (OS-4Sh, OS-11Sh and OS-18Sh) were found to have sunk slightly in August, and corrections were applied; however, their relative elevation may be as much as 0.1 foot (3cm) in error compared to the rest of the network.

Temporary Network and Springs

Seepage meters

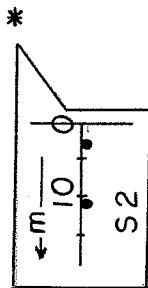
A direct way of measuring seepage flux between surface water and underlying groundwater is through the use of seepage meters (Lee, 1977). This device consists essentially of a plastic bag connected to an open bottom container and are designed to be placed over an area of sediments affected by seepage (see Fig. 7). By making certain that this container of known basal area is full of water at the start of the measurement, and by pouring a known initial volume of water in the plastic bag, it is possible to determine the seepage flux in the covered area, simply by measuring the change in volume of water which occurred in the bag over a known period of time. Many seepage meters so emplaced into the lake bottom could theoretically delineate both the areas of significant seepage and the total flux which took place between the lake and the underlying aquifer.

For this study, ten seepage meters were built by cutting off 15 cm sections from the two ends of 0.208 m^3 (55 gallons) metal drums, 57 cm in diameter. Two holes were drilled in each; one, on the side, to accommodate a 3.78 liter (1 gallon) plastic sample bag and, the other one, in the sealed top to insert an air vent.

The seepage meters were used at only four stations in the lake, during one day in June and three days in August (see Figure 6 for the

Figure 6 : Temporary and accessory network used for groundwater monitoring in Lake Wingra, and spring locations.

Legend for Figure 6 :



Seepage meter station and seepage meter placement within the station. The number on the right (e.g. S2) is the station # and the black dots represent the emplacement of the seepage meters in meters away from the shoreline. The seepage meter closest to the shoreline is assigned a "1"; for example, in this case the black dot closest to the shoreline is S2-1, and that farthest away is S2-2. If more than one seepage meter is installed at one seepage meter emplacement (black dot), then the subscripts a and b are used.

MP-1

⊕

Mini-piezometer station #1 (see Table D-5 in Appendix D for the details on the # of mini-piezometers per station and their location within the station).

Sp-1a

Δ

Spring #1a.

⊙

In-lake and onshore piezometers used during the summer study.

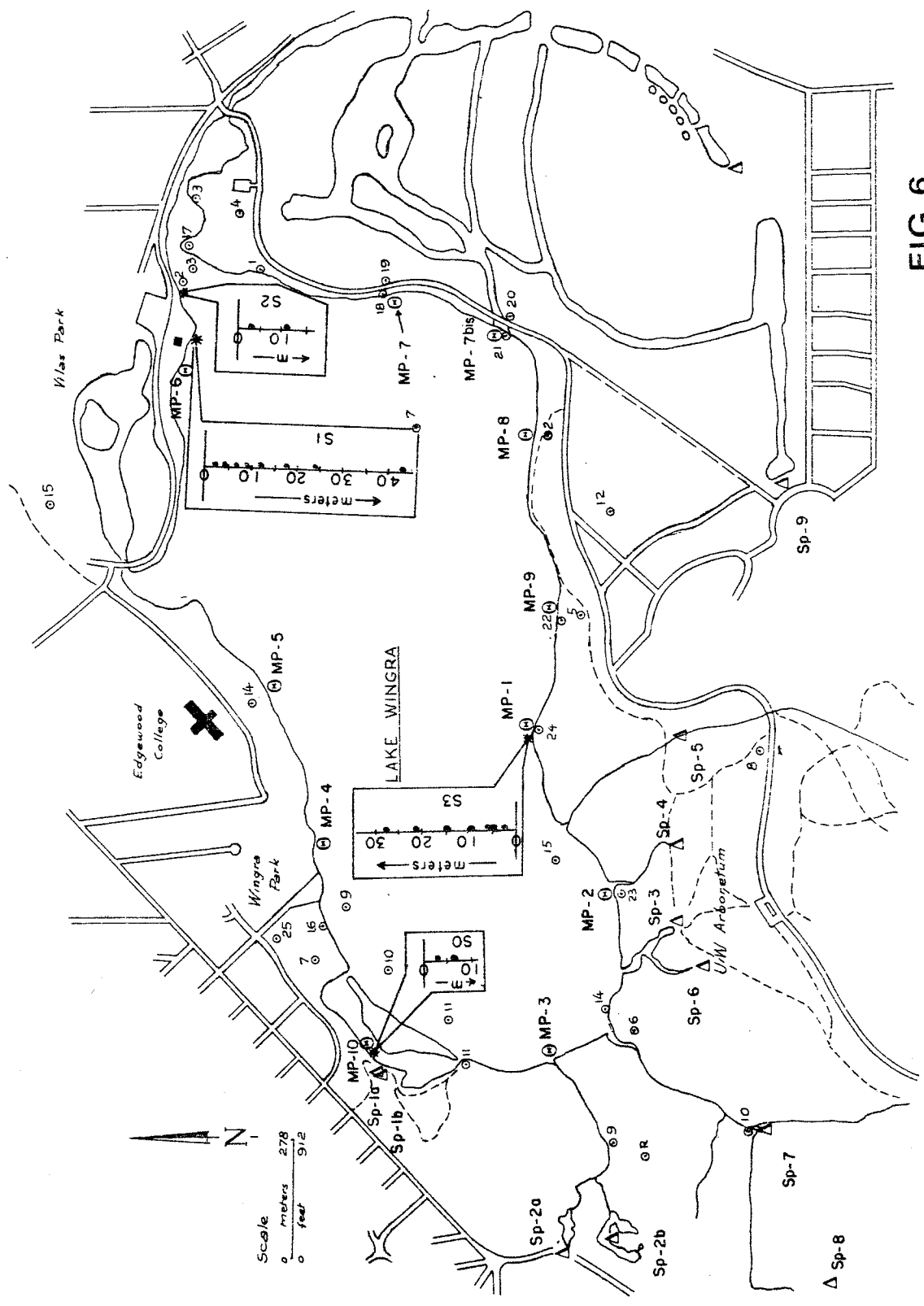


FIG. 6

Figure 7 : Details of a seepage meter.

Figure 8 : Mini-piezometer emplacement technique.

Explanation of drawings A, B and C in Figure 8 :

A/ Pounding in of a steel electrical conduit pipe.

B/ Insertion of the mini piezometer in the conduit pipe.

C/ Removal of the conduit pipe.

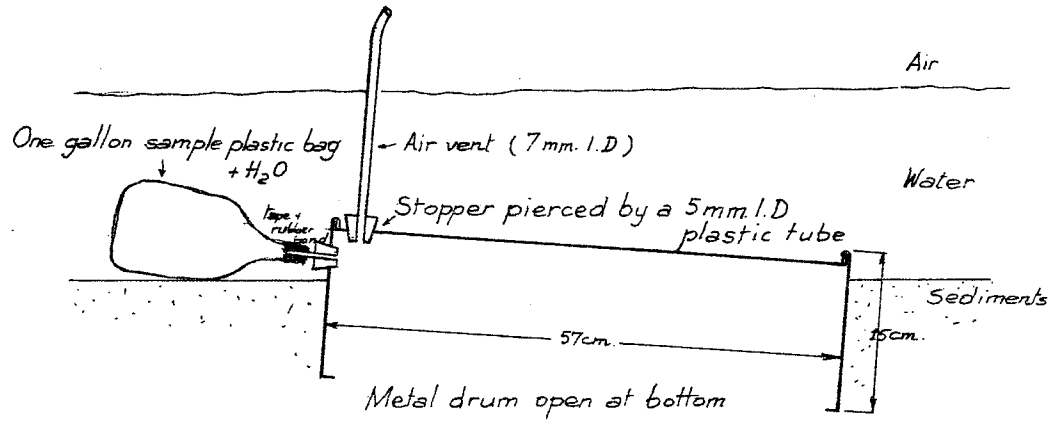


FIG. 7

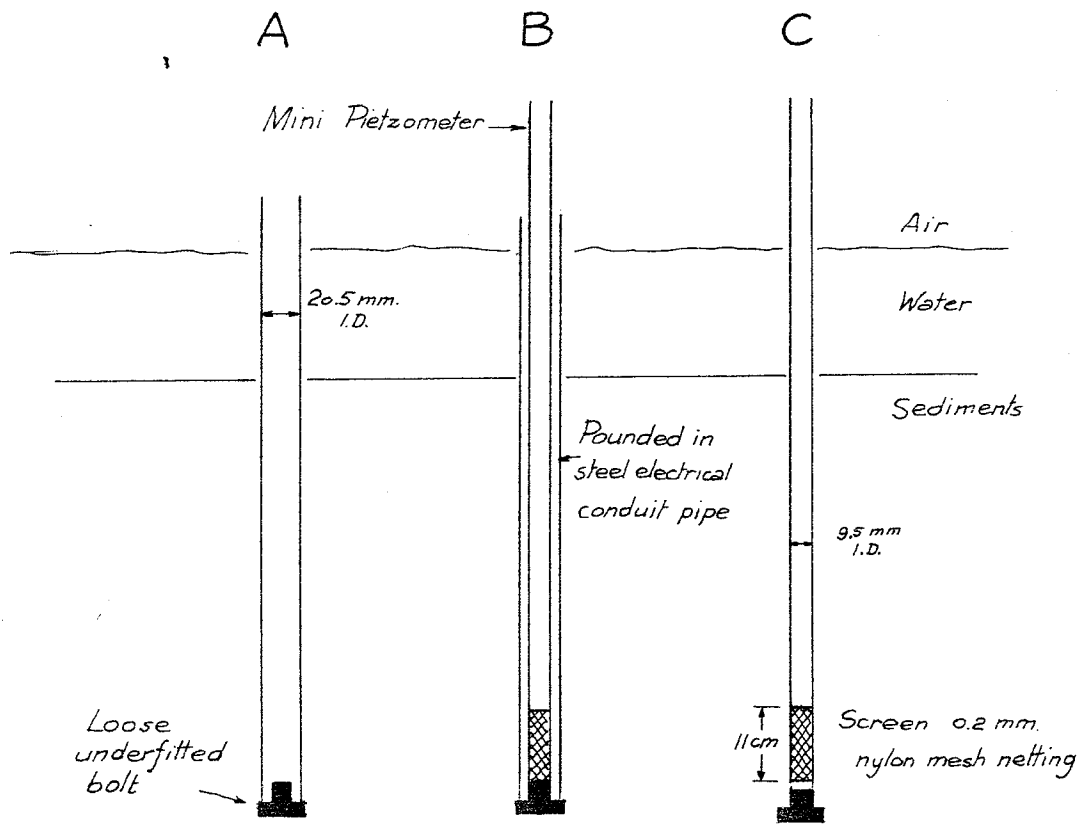


FIG. 8

location of the stations); vandalism and especially mucky lake bottom conditions prevented their long-term and widespread use. In fact, the lake bottom was so soft and ill-defined at some places, that some seepage meters sank and became buried in the sediments under virtually no pressure. Where used, two to eight seepage meters were placed at different distances away from the shore. Each was carefully pushed into the lake bottom, slightly tilted to ensure proper functioning of the air vent and then sealed from the lake water as recommended by Lee and Cherry (1978). The irregularities of the bottom and the muck made it necessary to push most of them 10 to 14 cm into the sediments, to ensure an adequate seal. After the drums had been in place for 10 minutes or so, stoppers pierced by a 5 mm ID plastic tube emerging at the top, were twisted in every drilled hole. Finally, a 7 mm ID air vent and a tube attached to a partially filled, prewetted plastic bag were slid on the 5 mm ID stopper tubes. The bags were totally submerged in water (see Figure 7). Additional informations about seepage meters may be found in Lee (1977), Lee and Cherry (1978) and Erickson (1981).

Mini-piezometers

Mini-piezometers are essentially piezometers reduced in size and installed manually (Lee and Cherry, 1978). These devices serve the same purpose as the inlake piezometers, but their nature and size renders them more economical, easier and faster to install, in addition to allowing more accurate water level measurements and faster response

time. In this study, they consisted of a 9.5 mm ID hard, translucent plastic tube, the bottom of which was plugged and the tip perforated and wrapped with a 0.2 mm nylon mesh netting. They commonly range between 1.83 m (6 ft) and 2.13 m (7 ft) in length, with an intake area 11 cm long (see figure 4).

The installation process included the pounding into the lake bottom of a 20.5 mm (3/4") ID electrical conduit pipe plugged at the bottom by a loose, underfit bolt, followed by the insertion of the mini-piezometer into the pipe, and finally, the removal of the latter, leaving both the bolt and the mini-piezometer anchored into the bottom sediments. This process is illustrated in Figure 8. The screen portion of the device generally was inserted between 0.76 m to 1.37 m below the lake-sediment interface (see Table D-5 in Appendix D for further details).

The likelihood that most of the seepage into and out of Lake Wingra occurs in the near-shore area (McBride and Pfannkuch, 1975), as well as the intense summer boat traffic on the lake led to the use of mini-piezometers in a narrow zone extending to about 22m (72 ft) from the shore. A total of ten stations with two to four mini-piezometers per station, were established in the lake for a few days during the second half of August 1981 (see Figure 6). Stations MP-7bis, MP-9 and MP-10, being in a less popular area of the lake, were kept operating through the remaining part of August and September. The mini-piezometers were emplaced along a line perpendicular to the shore, generally about 5.5 m (one boat length) apart for convenience sake. Each mini-piezometer was developed by filling it with water and applying buccal pressure until a

2-way connection between the device and the sediments was achieved.

Spring discharges

Most springs discharge via a small stream into the west and southwest sides of the lake (see Figure 6). Spring discharges were measured in the middle of August by one or two of three different techniques; the velocity-area, the tracer dilution and the floater method.

In the velocity-area method, the stream is first divided into convenient segments along a transect orthogonal to its flow. Next, depth and width of each segment were measured, and a pigmy current meter was placed in the center of the cross section, as close as possible, but always above the recommended 6/10 of the depth below the surface (United States Department of the Interior, 1967). When the propeller was totally submerged in the water, the number of revolutions per chosen time period was counted several times, averaged and then converted to a current velocity in feet per second. Finally, the volume flow was calculated for every segment and flows were added to obtain the total discharge of the spring (see Figure 9).

The tracer dilution technique involves the injection of a tracer upstream and the monitoring of its increase above background concentration downstream. If the tracer is injected at a constant rate, measurements downstream will initially indicate a concentration similar to the background level (C_0) in the stream. However, when the added

Figure 9 : Velocity-area technique used to measure spring discharge.

Explanation :

The total spring discharge (Q_{sp}) is :

$Q_{sp} = \sum_{i=1}^n W_i Z_i V_i$; where "n" is the number of stream segments (in this case 3) and W, Z and V are the width, depth and current velocity for a particular stream segment, i.

Figure 10 : Mariotte bottle and the tracer dilution technique used to measure spring discharge.

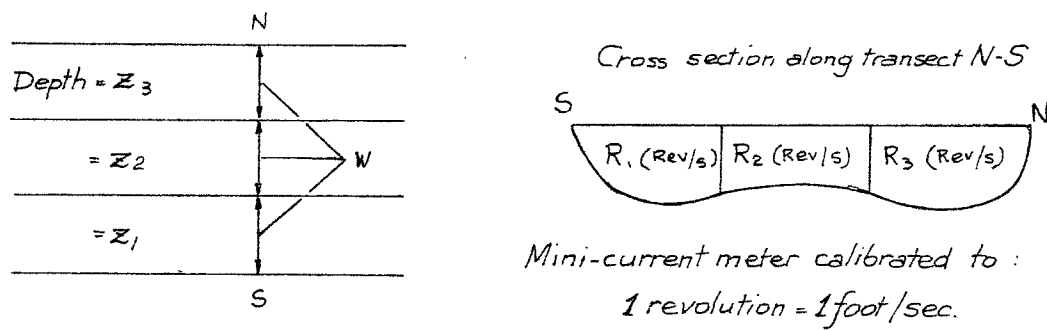


FIG. 9

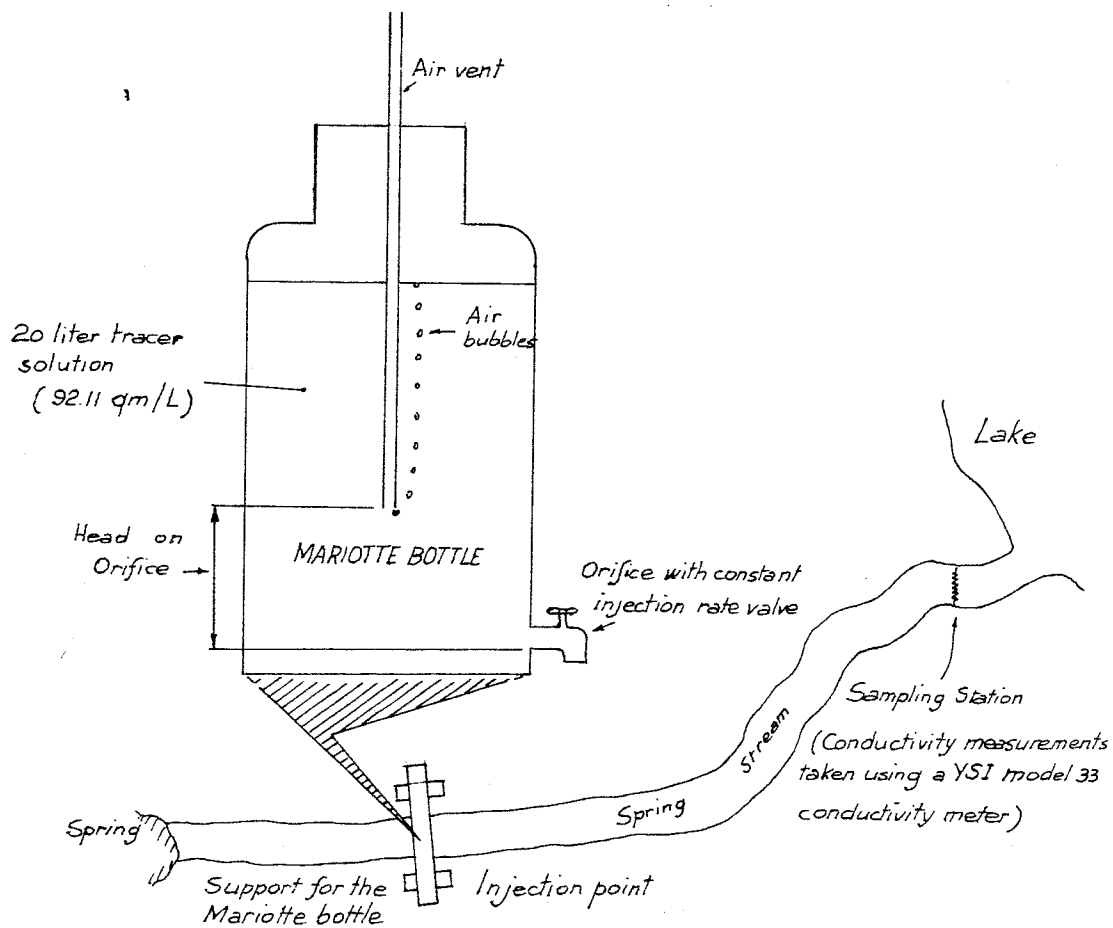


FIG. 10

tracer reaches the sampling station, its concentration will gradually rise to reach a plateau value (C_2), representing the sum of its steady state injected and natural background concentrations. If both C_0 and C_2 are carefully measured and, if in addition, the tracer concentration (C_1) in the injection solution and its injection rate (q) are known, the stream discharge (Q) may be calculated using the following equation (United States Department of the Interior, 1967):

$$Q = q(C_1 - C_2) / (C_2 - C_0) \quad (1)$$

The tracer used was A.C.S. NaCl (crystals). Constant injection flow was assured by use of a 20 liter volume homemade mariotte bottle placed in the middle of the stream, 20 cm above the water surface (see Figure 10). A YSI model 33 conductivity meter was used downstream to detect arrival of the tracer.

When the conductivity stopped rising after injecting the tracer (the plateau concentration was attained), several readings were taken across the stream at the downstream station, in order to verify that mixing of the tracer in the stream was complete and that no significant pond of still water existed between the tracer injection point and the measuring station; such a pond could divert some of the tracer and lead to undercalculation of the spring discharge. These measurements showed that mixing of the tracer was generally complete by the time the tracer reached the downstream station, inspite of the fact that all sampling stations were located only 15 to 30 stream widths below the injection

point; this is far less than the 100 stream widths rule of thumb suggested in the University of Waterloo Field Notes (1981). Therefore, tracer concentration was only a function of spring discharge. A laboratory predetermination of the relationship between conductivity and concentration of the tracer solution allowed for conversion of conductivities into concentrations, and the use of equation 1 (see Appendix B for the details).

Some spring streams had such a low current and were so full of weeds that the two techniques described above were impractical. For these springs, a method involving the use of a floater placed in a small clearing (protected as much as possible from the wind) in the stream. The downstream migration velocity of the floater multiplied by the average stream cross-sectional area through which it passed, permitted a gross approximation of the discharge for these springs. Table D-6 in Appendix D lists the different techniques used at each spring. One spring could not be measured because of its extremely low discharge and the absence of a patch of open water, but its discharge is estimated to be the lowest encountered around the lake.

III. FIELD DATA AND DISCUSSION OF THE VARIOUS GROUNDWATER MONITORING METHODS USED

Winter Data; In-Lake Piezometers

A few days after the installation of the in-lake piezometers in mid-January, the water level in some of them appeared to be at equilibrium. However, by the end of January the water had frozen in each.

Several options were considered to prevent freezing of the water in the pipe. However, most of these turned out to be either impractical or led to uncertain results. For example, injections of antifreeze or salt solutions into the well could lead to erroneously high water levels resulting from the extreme miscibility of these solutions with water. Low density oil is immiscible in water and should solve this problem. However, its use in small diameter off-shore piezometers in Green Bay (Bradbury, personal communication, 1981) met with several problems. Minute amounts of ice in the small diameter tubes (1 cm) prevented the oil from reaching the water inside. Piezometers in Lake Wingra were larger and this problem would be reduced or perhaps eliminated, but the large quantity of oil that would be needed to depress the water level far enough to insure no freezing at any time, combined with its messy

use and higher price relative to rubbing alcohol made this option impractical. Later field experiments however, suggest that in 1-1/4 inch diameter pipes, the use of oil may prove to be superior to alcohol. Electric water heating devices were also considered but were too expensive for a low budget study. A low cost 70% mixture of rubbing alcohol and water, which has the double advantage of being less dense than water and only slightly miscible in it, was used.

At the end of January, 1 pint of the rubbing alcohol mixture was poured into each off-shore piezometer. Water levels were measured twice in February 1981, but a premature rupture of the ice made further work on the lake impossible. Measurements were made with a chalked tape. The water levels obtained were corrected for the density difference between the alcohol and the water, by multiplying the initial height of the alcohol column in the piezometer by its laboratory determined average density of 0.8758 g/cm^3 . Table D-1 in Appendix D summarizes the data collected during the winter.

The use of alcohol raised some concerns regarding the accuracy of the corrected water levels in the piezometers. Namely, would the evaporation, the diffusion or the expansion of alcohol with temperature have a significant effect on the water levels calculated by simply correcting for the density difference? Various laboratory and outdoor experiments were set up to try to answer this question, and it was found that generally, neither diffusion, evaporation nor thermal expansion of the alcohol should be of any concern in the normal range of conditions associated with piezometers emplaced through the ice of a frozen lake.

The temperature of the alcohol in the piezometers normally remained close to 0°C, but rose to 5°C in some wells installed in areas of strong hydraulic gradient, during an unseasonably warm spell. Nevertheless, in the shallowest in-lake piezometers (about 10 ft in length) enough alcohol may have diffused through the screen and influenced the calculated water levels, which appear to be lower than they should be. The details of these experiments and further discussion of the results can be found in Appendix C.

Summer Data

Onshore and in-lake piezometers

Water level measurements in most onshore and in some in-lake piezometers, were taken on the average more than once a week, from May 15, to September 28, 1981. As the late spring and summer progressed, a few new onshore and the remainder of the surviving in-lake piezometers were included in the water level measuring round. The reason for the late inclusion of the in-lake piezometers into this measuring routine is two fold: 1. Except for IL-9, IL-10 and IL-11, the alcohol in the off-shore piezometers was not removed until early July. 2. After the removal of the alcohol, the water levels recovered much more slowly than expected, and in one case (i.e., IL-7) it never reached equilibrium during the study period.

Water level measurements were made with either a chalked tape or a popper. A precision of better than 0.01 ft was achieved in onshore piezometers; each measurement was repeated two or three times. Measurements made in in-lake piezometers suffered from a combination of wave and wind action, and so they were repeated several times until accuracy was reached. In general however, a precision of 0.01 to 0.02 ft was obtained. Accuracy, may not be as good in some slow responding in-lake piezometers. Well-concealed onshore piezometers remained out of the hands of vandals (thanks to the Arboretum's voracious army of mosquitoes), but in-lake piezometers were vandalized, and by early July covers on most had been removed, which allowed direct precipitation and lake water to enter the standpipe. Many boaters and swimmers used these piezometers as an anchor or resting rod, and since they are flexible they bent easily under water. This could be a problem for some of the wells with a particularly slow response time (i.e., IL-7, IL-15) because the water level may not have recovered by the time the measurement was taken. Nevertheless, most of the water levels recorded during the study period appear not only to maintain a consistent pattern, but also are compatible with earlier findings of Oakes et al. (1975) and Novitzki and Holmstrom (1979), and thus the problem was probably minor.

The data collected during the weekly water level measurement rounds are summarized in Table D-2A in Appendix D. Figures D-2 to D-14 are graphical representations of Table D-2A and should aid in obtaining a quick visual understanding of the lake hydrogeological system as discussion proceeds in the following chapters. These graphs were

created using SHOW, a graphics program available at MACC (Madison Academic Computer Center). Additional water level data were gathered at some sites during the summer and appear in Table D-2B.

The water level data for a few partially plugged piezometers need some explanation. First, the piezometer at IL-7 probably never reached equilibrium during the summer study. However, its water level is assumed to be at about the same elevation as the lake surface, as suggested by data collected during the winter (see Table D-1) and by the fact that the water levels in the eastern piezometers were closer to the lake surface during the summer compared to during the winter (discussed in Chapter IV). Secondly, the deep piezometer at OS-19 was essentially clogged, but a combination of one slug and one bail test (which lasted all summer) definitely showed that the area is affected by a significant upward hydraulic gradient. Furthermore, the head toward which both tests converged seemed to be similar to that found at the same time in the deep piezometer of OS-18 across the road. Because of the lack of better data, the hydraulic head in the deep piezometer at OS-19 is assumed to mimic that in the deep piezometer of OS-18. Finally, the water level measurements recorded in the deep piezometer at OS-8 are included in Table D-2A, but caution should be exercised when interpreting these numbers, because the well probably became significantly plugged in mid-June.

Staff gauges

In addition to the measurements taken at the piezometer sites, measurements at the two staff gauges were also made during the measuring rounds. The Ho-Nee-Um pond gauge recorded the lake stage fluctuations, and the USGS dam gauge allowed estimates of the variations of the surface water discharge through the wier to be made. The data are listed in Table D-2A.

Automatic water level recorder

The water level in the water table well (referred to as R on Fig.5) monitored by the automatic water level recorder was checked about once a week, in a effort to ensure the proper functioning of the recorder. The data also appear in Figure 21 in Chapter IV.

Precipitation

Daily precipitation measurements were taken at 8:00 am by one of the Arboretum staff. Table D-3 in Appendix D includes the data relevant to the summer study. A discussion of precipitation data is also included in Appendix E. Note in particular that 1981 had a winter drier than normal, and a spring and summer wetter than normal.

Seepage meters

Seepage meters were used on June 15 and during August 13 to 15, but failed to yield meaningful results in Lake Wingra. The results listed in Table D-4 of Appendix D would suggest an extreme heterogeneity of the lake bottom. However, many inconsistencies among the results, as well as disagreements with adjacent mini-piezometers or nearby in-lake wells, led to an early abandonment of the seepage meter approach to the groundwater budget problem. Although, some of these results may be valid, most of them are believed to be erroneous, and so were not used in this study.

Many possible reasons could account for this failure. 1. The extremely mucky nature of the lake bottom and the abundance of gases trapped in the lake sediments made it easy for the seepage meters to sink slowly into the bottom and water to be forced into the plastic bag. 2. Gases could also influence the exchange of water between the bag and the drum, if they are released under the drum and are unable to escape through the small diameter air vent. 3. The seepage meters were not in place long enough before the measurements were taken. Hence the system was in disequilibrium and organisms trapped under the drum may have affected the measurements (Erickson, 1981). 4. The two main seepage meter stations are in areas of very weak seepage, and the groundwater velocity may have been too low to measure with this technique; seepage meters are said to work best in zones with moderately high to high seepage rates (Lee, 1977). 5. The irregular bottom of the lake and the

abundance of aquatic plants and weeds may have prevented an adequate seal around the drums, inspite of efforts to ensure proper sealing.

Beside these potential problems there exists a score of other problems associated with seepage meters. Erickson (1981) in his study of seepage meters concluded that:

1. The presence of the plastic bag increases the natural hydraulic head.
2. Prewetting the plastic bag changes the results.
3. A greater initial volume of water in the bag yields smaller in-seepage rates.
4. The emplacement of the seepage meter affects the hydraulic gradients in the natural flow field, and even more so after placing the stopper and the tubing on the drum.

The last conclusion is particularly significant in Lake Wingra since the seepage meters were pushed deeper into the sediments than the 8 cm recommended by Lee and Cherry (1978). Walking around the drums also may have disturbed the natural flow field.

Nevertheless, knowledge acquired later suggests that a more careful search for better sites combined with a longer period of time allowed for equilibrium to be restored and entrapped organisms to escape from the drum, could significantly improve the results, but not enough to justify their use in Lake Wingra or a lake with similar bottom conditions.

Mini-piezometers

Use of mini-piezometers was moderately successful in Lake Wingra. In general, these devices were used from August 19 to 22, but those at MP-7bis, MP-9 and MP-10 remained operational until the end of September (see Figure 5). Water level measurements in the mini-piezometers were taken anywhere from 2.5 hours to several days after installation. As a rule, the water levels remained at the same elevation for at least 1-1/2 hour before the reading was taken, in order to ensure that equilibrium was established. A ruler graduated in millimeters was used to measure the difference in elevation between the lake surface and the water level in the mini-piezometer.

Although many mini-piezometers yielded what is believed to be reliable water levels, some did not, specifically those from stations in the west, east and northwest. Unreliable head measurements could arise because of non-equilibrium or more probably, were due to the presence of gas bubbles trapped in the tube. At most of these stations, gas bubbles were observed to lift the water column, and gas could not readily escape unless a series of small shocks was applied on the mini-piezometer. Reasons for the immobility of these bubbles may include static electrical forces and surface adhesion. More details and all the significant data pertaining to the water levels observed in the mini-piezometers appear in Table D-5 in Appendix D.

Although the mini-piezometers were only moderately successful, they represent an invaluable complementary tool that can provide a great deal

of information on shallow groundwater movement into or out of a shallow, relatively small lake. Their performance could probably be improved if slightly larger diameter tubes were used to aid gases to escape, and if the minimum time allowed to reach equilibrium inside the tube were extended to at least one day, for eutrophic lakes with shallow, mucky bottoms.

Spring and seep discharges

The shallowness and tortuosity of the small spring streams rendered difficult the task of finding a place suitable for discharge measurements. Nevertheless, the discharge at most springs was measured on August 17.

Whenever the tracer dilution and the velocity-area methods were used in the same spring, the discharges obtained usually agreed to within 12 to 25%. Reasons for the small discrepancies are many. The discharge measured by the tracer dilution technique could be affected by lag waters in still ponds located between the tracer input and the measuring station. Incomplete mixing of the tracer at the downstream station, or errors induced in reading the conductivity or generated by the device itself could also affect results. Except perhaps at Sp-3, the first two sources of error seem unlikely for reasons explained in Chapter II. However, the fact that the plateau concentration attained after the injection of the tracer never exceeded the background concentration by more than 10 to 16%, instead of the 30% recommended in

the University of Waterloo Field Notes (1981), could have led to mistakes in reading the conductivity.

The velocity-area method also has potential sources of errors, two of which may have influenced the results at Lake Wingra. The first involves the shallowness of most streams and the necessity of placing the meter propeller above the recommended $6/10$ of the depth below the water surface, which supposedly is the depth where the velocity best represents the average current velocity in the stream (United States Department of the Interior, 1967). Placing the meter propeller above the recommended depth could probably lead to overestimates of the discharge, since in general the propeller turns faster near the surface than at depth. Finally, in view of the numerous obstacles (e.g., weeds and rocks) in the streams, the chosen points of measurements may not have yielded representative velocities. Nevertheless, the general consistency of the results argues that overall the measured discharges closely reflect the real discharges. Spring discharges are summarized in Table D-6 of Appendix D.

Three measurements at Sp-1a in 1981 and 1982, suggest that the discharge at this spring varies with time, and in fact can double within a few days after a major storm. If this variation is typical of the discharge of all springs around the lake, it is significantly greater than the discharge fluctuation reported by Oakes et al. (1975).

IV. THREE-DIMENSIONAL SHALLOW GROUNDWATER CIRCULATION AROUND AND BELOW LAKE WINGRA

Introduction

The shallow groundwater flow in and around Lake Wingra is relatively complex as illustrated by the wide variety of positive (upward) and negative (downward) vertical hydraulic gradients which characterize the flow system. The first part of this chapter presents an overview of the general groundwater circulation, followed by a more detailed description of the groundwater movement and of its variations in specific areas in and around the lake. Maps and cross sections for September were chosen to illustrate some of the points developed below, because the September data are more complete than other monthly data sets, and because the overall configuration of the groundwater circulation does not change very much, even on a daily basis. The second part of this chapter is concerned with the factors responsible for the fluctuations of the water table and groundwater potentials. To understand the numbering scheme used for the groundwater monitoring network, the reader is referred to the symbol index at the beginning of the thesis.

Groundwater Circulation

Overview

Figures 11 and 12 show the general groundwater circulation around and below Lake Wingra for September 1981. Groundwater flows toward the lake in the north, west and south, and away from the lake in the east. The groundwater movement in the west shore area is mostly vertical and upward (see Fig. 13 and 15), and mostly horizontal in the north and south shores (see Fig 15, 16 and 17). In the east shore area, groundwater flow is essentially across the Arboretum drive, then becomes slightly upward immediately east of the drive (see Fig. 13), before becoming mostly downward in Gardner Marsh (see Fig. 14).

The groundwater movement below the lake is summarized in Figure 12; most of the western portion of the lake bottom as well as two relatively narrow bands of the lake bottom adjacent to the north and south shores, are affected by upward groundwater flow from the aquifer. On the other hand, in a fairly extensive area of the lake bottom adjacent to the eastern shoreline and in Vilas Pond, the groundwater movement is downward into the aquifer. Figure 12 is based on the data acquired from in-lake piezometers during February and during the summer of 1981, and from ten mini-piezometer stations installed in mid- August. It should be emphasized that this is in some parts an approximation, and that more data are needed to define more accurately, and possibly to redefine some

Figure 11 : Average shallow groundwater flow around Lake Wingra
during September 1981. Contour interval, 1 foot.

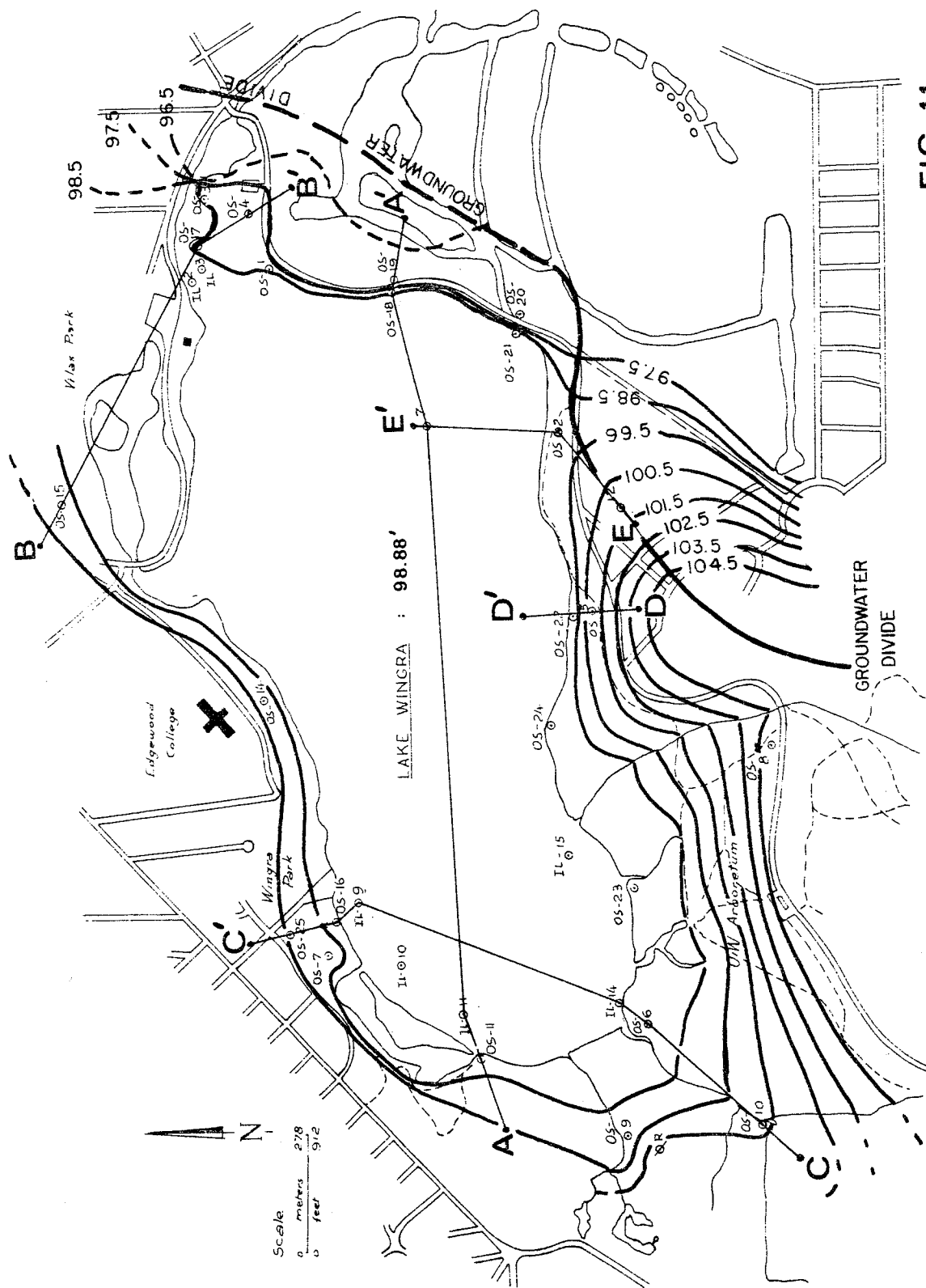


FIG. 11

Figure 12 : Map summarizing the areas of groundwater in- and out-seepage for Lake Wingra. The dashed lines show an estimate of the extent of the lake bottom affected by significant seepage, based on data obtained from in-lake piezometers and mini-piezometers during February and the summer of 1981.

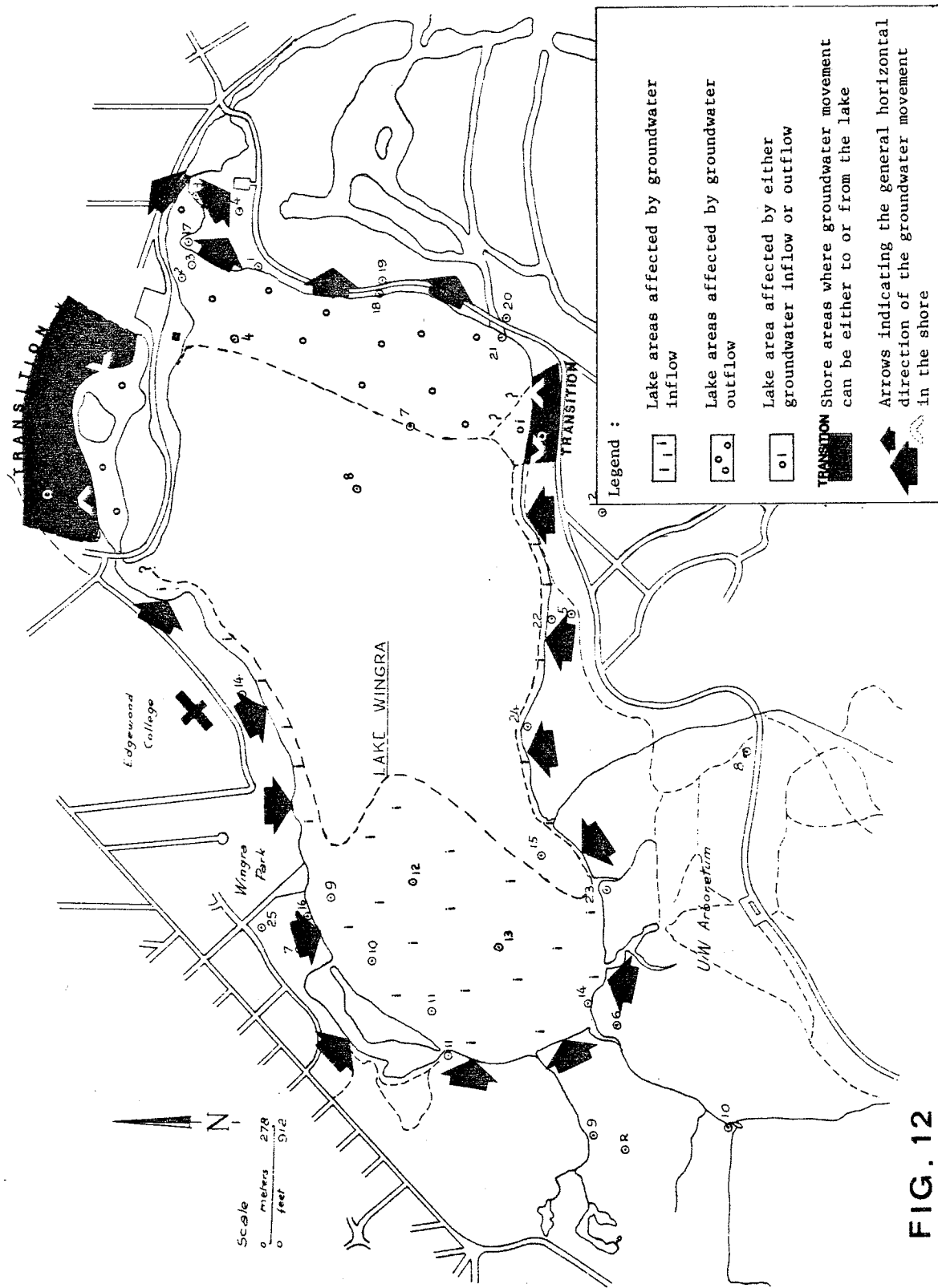


FIG. 12

Figure 13 : Mean head distribution along cross section A-A' for September 1981 (see Figure 11 for the location of the cross section).

Figure 14 : Mean head distribution along cross section B-B' for September 1981 (see Figure 11 for the location of the cross section). From about the southeast side of the pond to the north side of Lake Wingra, the upper portion of the groundwater flow is diverted progressively more eastward, a result of the USGS spilling dam at the head of Murphy Creek (see text).

Figure 15 : Mean head distribution along cross section C-C' for September 1981 (see Figure 11 for the location of the cross section).

Figure 16 : Mean head distribution along cross section D-D' for September 1981 (see Figure 11 for the location of the cross section).

Figure 17 : Mean head distribution along cross section E-E' for September 1981 (see Figure 11 for the location of the cross section). The groundwater flow is essentially northeastward around OS-12, but becomes southeastward (out of the page) at OS-2 (see text).

of the northern and eastern portions of the lake bottom affected by significant seepage, here arbitrarily defined as areas where the vertical hydraulic gradient exceeds ± 0.003 .

The general aspect of the groundwater circulation does not undergo significant modifications on a mean monthly or even daily basis, at least during a wet summer (see Appendix E), without large scale pumping. Comparing Figures 11 and 14 for September with Figures 25 and 27 (in Chapter V) for August supports arguments for the stability of the overall configuration of the groundwater system, in spite of the progressive rise and continuous fluctuations of the groundwater levels throughout the summer (see Figure D-2 to D-14 in Appendix D).

East

From the onshore hydraulic gradient pattern (see Figure 11, Table D-2A and Table 2), it appears that south of OS-1 lake water and deeper groundwater seep from the west across and beneath the Arboretum drive (which serves as a permeable dam), into the marsh, creating a relatively narrow zone of upward flow in the marsh area immediately adjacent to the road (see Figure 13 and 14). North of OS-1, a similar situation exists, although on a smaller scale; water flows from the lake east and south into the shoreline sediments, where the water table is in some places more than 1 ft lower than the lake stage, and a narrow zone of weakly upward seepage characterizes the area immediately adjacent to the shoreline. The remaining portion of the marsh appears to show a

generally downward groundwater movement, with perhaps a significant horizontal component in the upper 10 to 15 ft of the sediments, close to the lake.

The in-lake area is not as well known. As already mentioned, close to shore in-lake piezometers and mini-piezometers recorded a downward vertical hydraulic gradient. During the winter 80-81, in-lake piezometers showed a slight downward gradient at more than 1000 ft west of the central portion of the east shore (see Table D-1 in Appendix D). However, it seems safe to assume that most of the recharge to the aquifer takes place within the area adjacent to the shore, as suggested by two lines of evidence: first, the minimum vertical gradient recorded during the winter at IL-5 (-.12) was about 27 and 42 times greater than the maximum gradient observed at IL-7 (-.0045) and IL-8 (-.0029) respectively. Secondly, MP-7a (see Figure 6) showed a strong negative gradient (-.18) in mid-August, while the piezometers at IL-3 and probably IL-7 showed no or relatively low vertical gradients.

In the fine grained sediments of the northwest portion of the east marsh, a relatively slow and steady decline in the horizontal hydraulic gradient occurred during the summer; the September gradient was about half of what it had been in June. No reversal of the horizontal groundwater flow occurs on either a monthly or daily basis. Relative to the horizontal gradients, the vertical hydraulic gradients are more variable; on a monthly basis, they change from a few percents in the piezometers installed in finer grained sediments, to more than a factor of 4 in those emplaced in coarser grained material. No vertical flow

reversals occur on a monthly basis (see Table 2). However after a major storm (1-2 inches or more) the narrow zone of upward flow north of OS-1 may become a temporary zone of downward flow.

The water levels in the piezometers located along the Arboretum drive vary essentially as the lake or marsh level does. No horizontal groundwater flow reversals occur (the marsh level remains at all times lower than the lake surface), but sometimes after a major storm, the vertical flow may reverse temporarily, but only in piezometers on the lake side of the Arboretum drive. The daily water level fluctuations in all the eastern onshore piezometers generally range between 0 and 2.5 ft, with greater variations in the shallower piezometers.

The few data available for the off-shore area suggest that this area probably does not undergo significant changes with respect to the groundwater movement in the sediments, but the extent of the seepage zone may fluctuate somewhat from month to month in response to storms and droughts, or to pumping of the deep aquifer by the City of Madison.

Vertical gradients in the marsh show a marked increase with depth below 10 or 15 ft; this is due to the presence of a low hydraulic conductivity layer at depth (see Chapter V).

West

In the western portion of the study area, groundwater generally flows from the northwest, west and southwest, either directly into the lake or indirectly via a small stream. As in Gardner Marsh, the flow in

Table 2. Mean monthly vertical hydraulic gradients observed in and around Lake Wingra during the summer of 1981.

Site #	Points of measurements	June ¹	July	August	September
OS-1	Sh-D	+0.041	+0.038	+0.043	+0.021
OS-2	Sh-D	-0.0023	-0.0052	-0.0035	-0.0069
	Sh-I	+0.0005	-0.0042	-0.00084	-0.00084
OS-4	Sh-D	-0.12	-0.12	-0.13	-0.11
	Sh-DI	-0.051	-0.043	-0.064	-0.045
	Sh-ShI	-0.019	-0.032	-0.034	-0.011
OS-5	Sh-D	+0.012	+0.0094	+0.0067	+0.0047
OS-6	Sh-D	+0.082	+0.080	+0.078	+0.075
	Sh-I	+0.023	+0.018	+0.022	+0.020
OS-8	Sh-D	+0.0064	+0.026 ²	+0.032 ²	-0.010 ²
OS-9	Sh-D	+0.16	+0.15	+0.16	+0.16
OS-10	Sh-D	-	+0.19	+0.20	+0.21
OS-11	Sh-D	+0.34	+0.33	+0.34	+0.33
OS-15	Sh-D	-	-0.030	-0.0067	-0.023
OS-16	Sh-D	+0.042	+0.047	+0.050	+0.051
	Sh-I	+0.0039	+0.017	+0.020	+0.010
OS-17	Sh-D	+0.021	+0.020	+0.010	+0.014
OS-18	Sh-D	-0.028	-0.061	-0.015	-0.047
OS-19	Sh-D	+0.066 ³	+0.065 ³	+0.062 ³	+0.057 ³
OS-20	Sh-D	+0.17	+0.17	+0.17	+0.15
OS-21	Lake-Sh	-0.037	-0.039	-0.047	-0.042
OS-22	Sh-D	+0.027	+0.026	+0.026	+0.029
OS-23	Sh-D	-	+0.018	+0.023	+0.021
	Sh-I	+0.0031	.0000	+0.0027	+0.0027
OS-24	Sh-D	+0.0018	+0.0036	+0.0031	+0.0040
	Sh-I	.0000	+0.0011	+0.0022	+0.0044
OS-25	Sh-D	-	+0.025	+0.033	+0.025
IL-2	Lake-D	-	-0.031 ⁴	-0.028	-0.022
	Lake-Sh	-	.0000	-0.0018	-0.0000
IL-3	Lake-D	-	-0.0062 ⁴	-0.0047	-0.0027 ⁵
	Lake-Sh	-	-0.0029	-0.0022	-0.0022
IL-7	Lake-Sh	-	-	.0000 ³	.0000 ³
IL-9	Lake-D	+0.0043	+0.0050	+0.0057	+0.0071
IL-10	Lake-D	-	.0000	+0.0016	+0.0033
IL-11	Lake-D	-	+0.044	+0.040	+0.039
IL-14	Lake-D	-	-	+0.037 ⁶	+0.042
	Lake-Sh	-	.0000	.0000	+0.0030
IL-15	Lake-D	-	-	.0000 ⁶	.0000
MP-9	Lake-a	-	-	+0.089 ⁷	+0.11
	Lake-b	-	-	+0.056 ⁷	+0.068
	Lake-c	-	-	+0.031 ⁷	+0.038
MP-10	Lake-a	-	-	+0.071 ⁷	+0.078

- 1 June gradients computed only from the water levels measured on the 6/6, 6/13, 6/18, 6/25 and 6/30.
- 2 Gradients affected by the near total clogging of the point.
- 3 Gradients estimated only (see section E in chapter I).
- 4 Gradients based on water levels recorded in the second half of July only.
- 5 Water level for 9/11 not included in the calculation of the gradient.
- 6 Gradient based on water levels recorded in the second half of August only.
- 7 Gradients obtained from one late August water level measurement.

the west marsh and wetlands is more vertical than horizontal, certainly below a depth of about 2 to 4 ft, but unlike the east, the west has a strong upward component of groundwater flow as evidenced by the many flowing wells which dot the area. The strongest vertical hydraulic gradients are found in this region (see Table 2), and as in the east also increase in magnitude with depth. There is nonetheless a substantial lakeward horizontal flow in the first couple of feet of the surface sediments, in the root zone.

The western portion of the lake is also characterized by positive vertical gradients, the strongest found in the lake, and thus groundwater seeps up through the bottom sediments and into the lake. In-lake piezometers emplaced all over the western portion of the lake showed that this upward seepage occurred over most of the the area and probably beyond, during the winter (see Table D-1). No summer data are available for the central portion of the west part of the lake, and so the lateral extent of the summer upward seepage is not known for certain at this time. However, in-lake piezometers measured during the summer seem to reveal hydraulic gradients roughly similar to those obtained during the winter (see Table 3). It is therefore believed that nearly all of the western part of the lake is continuously affected by significant upward seepage, although most of it takes place in Ho-Nee-Um Pond and in the lake area closest to the marshy shore.

The configuration of the groundwater circulation in the west is very stable; flow reversals never occur, even in times of prolonged drought or after major storms (more than 2 inches). However, during

Table 3. Comparison of the winter and summer groundwater levels in some eastern and western in-lake piezometers.

AREA	SITE	PIEZOMETER	AVERAGE WINTER GROUNDWATER LEVELS RELATIVE TO THE LAKE SURFACE (FEET)	AVERAGE SUMMER GROUNDWATER LEVELS RELATIVE TO THE LAKE SURFACE (FEET)
East	IL-2	D	-.63	-.41
	IL-3	D	-.18	-.063
	average:		-.41	-.24
West	IL-9		+.07	+.07
	IL-10		+.12	+.01
	IL-11		+.48	+.56 ¹
	IL-14	D	+.83	+.69 ¹
	IL-15		+.03	+.00 ¹
average:		+.31	+.27	

1.- Only data for August and September are included.

severe storms, the shallow groundwater system may be significantly perturbed (but not reversed) by discharging storm sewers (i.e., the Nakoma Road storm sewer). The vertical and horizontal hydraulic gradients commonly vary by less than 10 to 20%, and the water table usually fluctuates by less than one foot, except after major storms when water level rises of 1.5 ft have been observed.

South

Most of the south shore is characterized by a predominantly horizontal, lakeward groundwater movement with a slight upward component superimposed, except in the east. A closer look at the vertical gradients and at the distribution of springs reveals that the vertical component is generally greatest and positive in the west, negligible in the center and becomes negative in the east. For example, the vertical hydraulic gradients obtained from the piezometers at OS-6, 23 and 24, show a trend of decreasing gradients toward the east, and at OS-2, the vertical gradient is negative. There is a higher gradient at OS-22, most likely caused by the proximity of a relatively steep sided hill (on the south central shore) located upflow of the groundwater movement, and which is an isolated exception to this general trend. The springs change from major springs exiting directly from bedrock outcrops in the western south shore, to minor sand boils percolating upward through the bottom sediments of a small creek in the central south shore (see Figure 6 and Table D-6 in Appendix D).

As groundwater crosses the vertical plane of the south central shoreline, it is rapidly deflected upward and percolates through the bottom sediments into the lake (see Figure 19a). This is shown by the increase in the late mid-August vertical hydraulic gradients observed between the shallower onshore piezometers close to the shoreline, and their adjacent off-shore mini-piezometers (see Table 4).

The southwest portion of the south shore does not conform to this trend; the piezometer nest at OS-23 does not exhibit an increase in vertical gradient between the shore and the lake (see Table 4). Nevertheless, the gradient increases at a depth of about 20 ft below the water table, suggesting the existence of a deeper upward flow system, and possibly seepage into the lake occurs farther off-shore. This is not inconsistent with the conceptual model developed for the western lake area, calling for upward seepage through most of the lake bottom.

In the portion of the lake adjacent to either the south central or southwestern shore, the vertical seepage decreases rapidly and linearly, with distance away from shore (see Figure 18a,b,c). The bottom of the lake affected by most of the upward seepage is therefore a zone roughly 75 ft wide adjacent to the shoreline (see Figure 19a), as suggested by the consistent results obtained from three mini-piezometer stations and the zero-vertical gradient recorded at IL-15. In addition to the upward groundwater seepage, there may be an important horizontal seepage component into the lake, and so the total groundwater seepage from the south shore should not be computed only from the vertical hydraulic gradients recorded by the mini-piezometers.

Table 4. Comparison between the vertical hydraulic gradients observed in both the shallower onshore piezometers close to the shoreline and in the adjacent off-shore mini-piezometers, for August 19 and 25, 1981.

Mini- Piezometers ¹	MP-2 (Aug.19)	MP-1 (Aug.19)	MP-9 (Aug.25)
LAKE c	+0.0000	+0.0034	+0.032
b	+0.0010	+0.0055	+0.055
a	+0.0017	+0.0074	+0.089
Onshore Piezometers	OS-23 (Aug.19)	OS-24 (Aug.19)	OS-22 (Aug.25)
SHORE	+0.0017	+0.0016	+0.023

¹Mini-piezometer "a" is the closest to shore and mini-piezometer "c" is farthest from shore (see Table D-5 for more details).

- Figure 18 : Decrease in vertical hydraulic gradient with distance away from shore measured in several mini-piezometers (see Figure 6 for their locations):
- a. MP-9 on four different days in late August and September 1981;
 - b. MP-1 on August 19, 1981;
 - c. MP-2 on August 19, 1981;
 - d. MP-4 on August 19, 1981;
 - e. MP-5 on August 20, 1981.

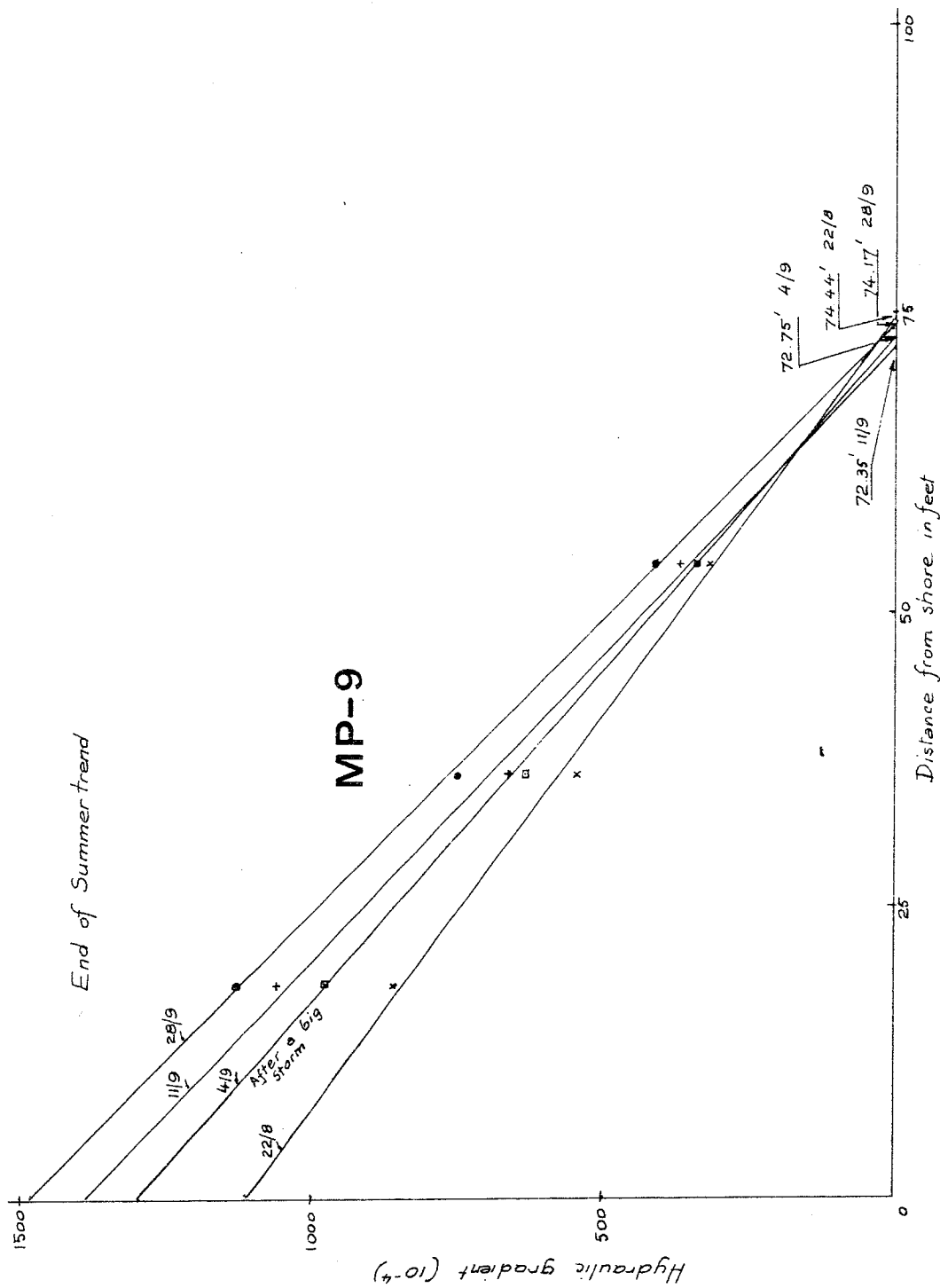


FIG. 18a

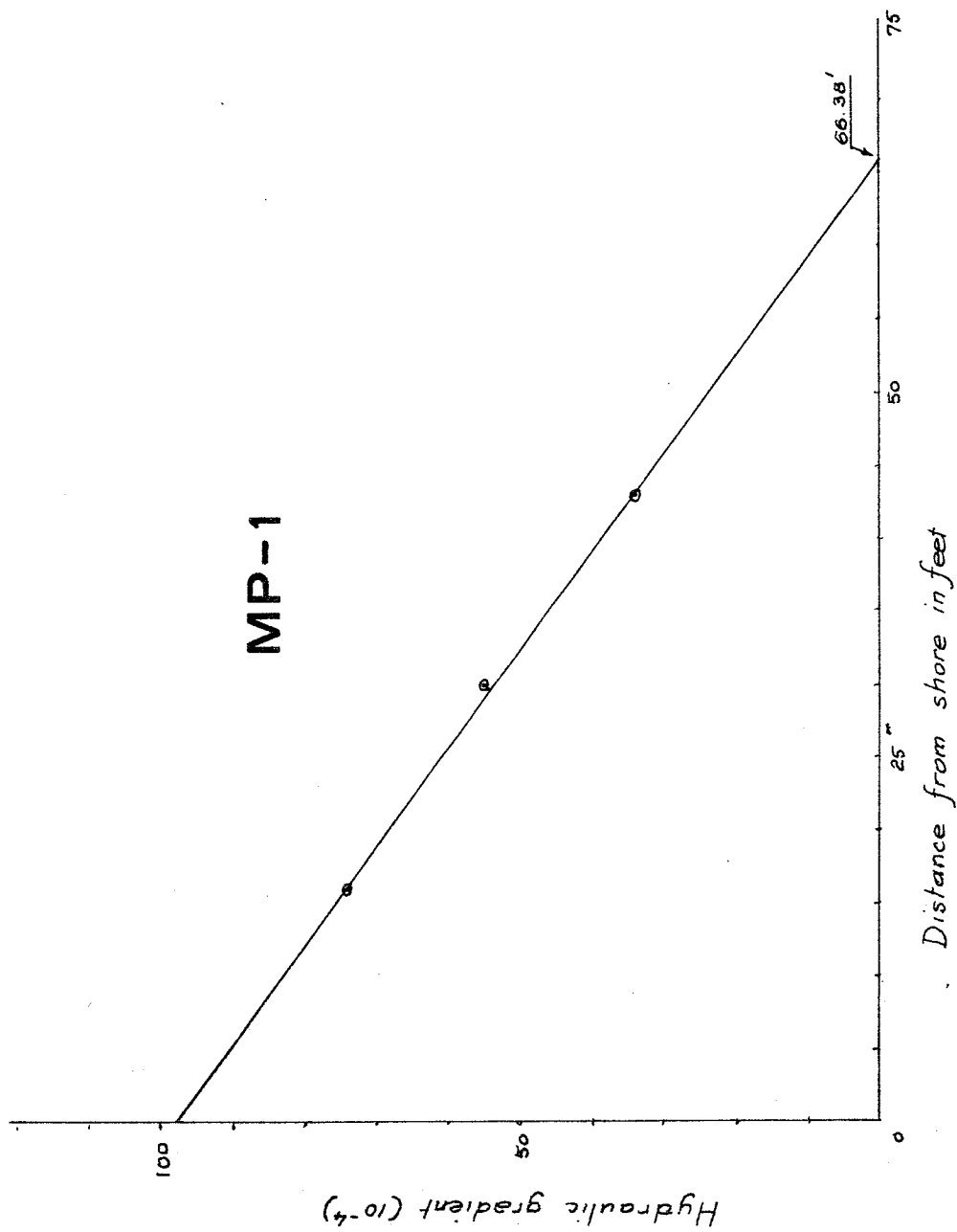


FIG. 18b

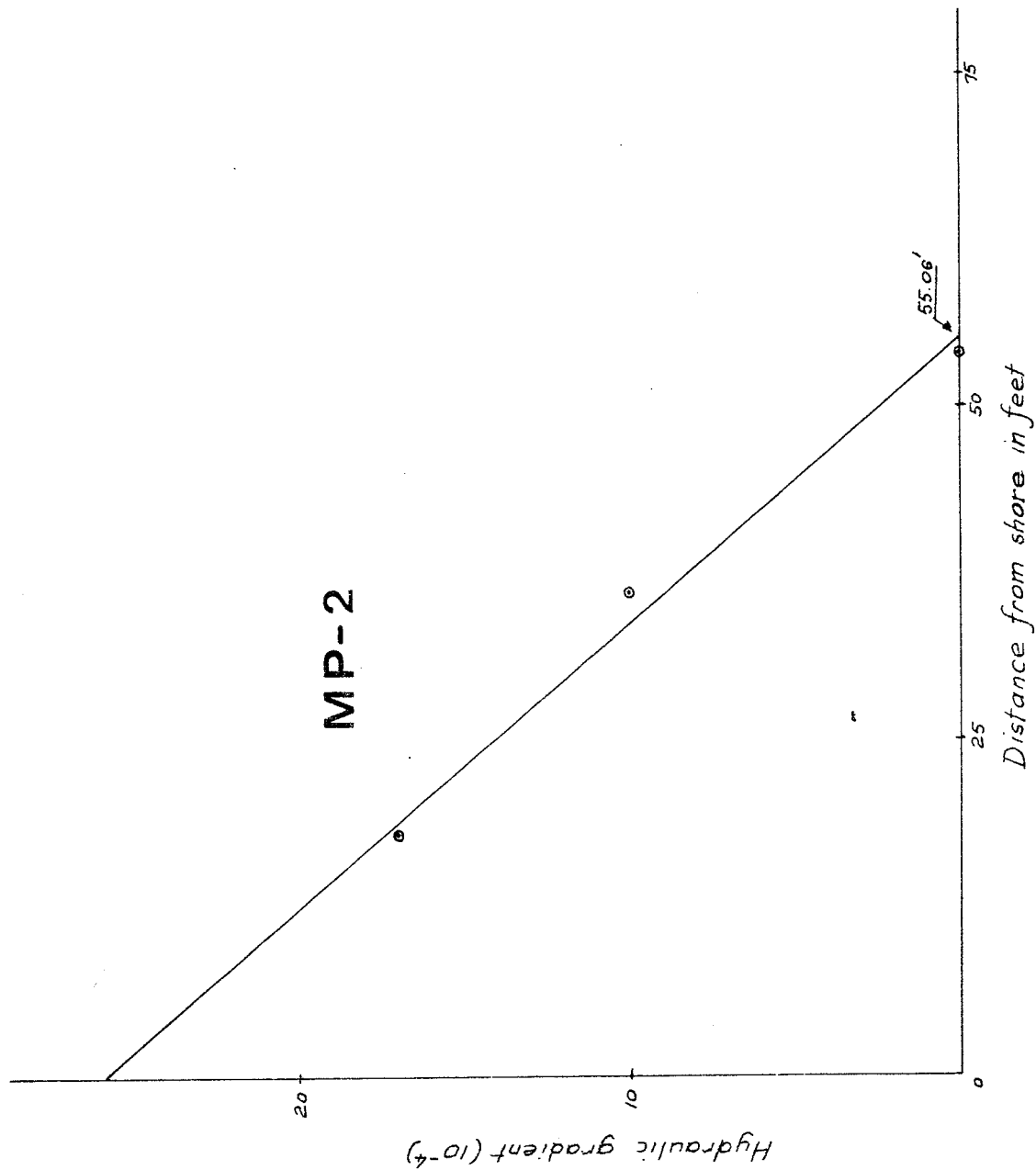


FIG. 18c

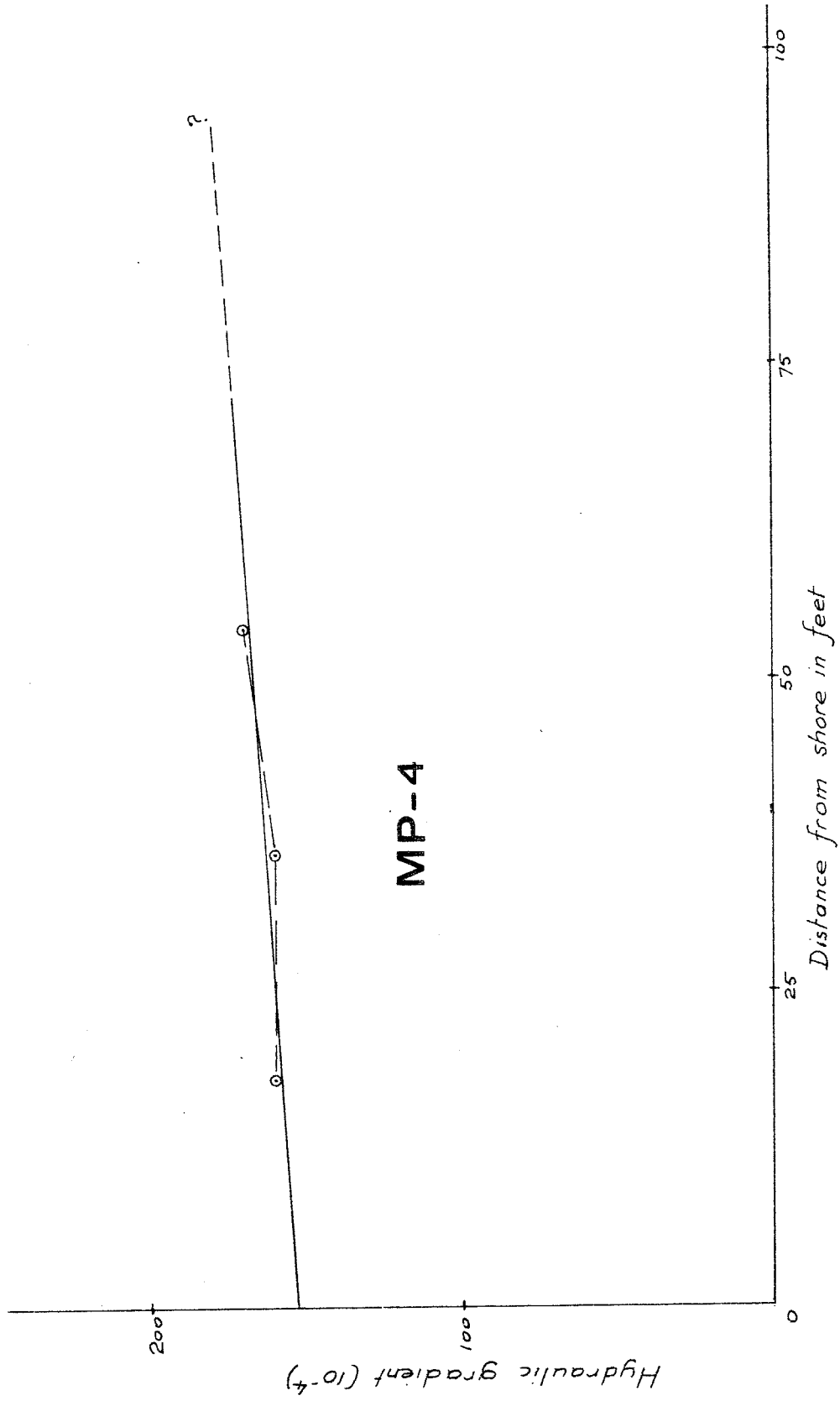


FIG. 18d

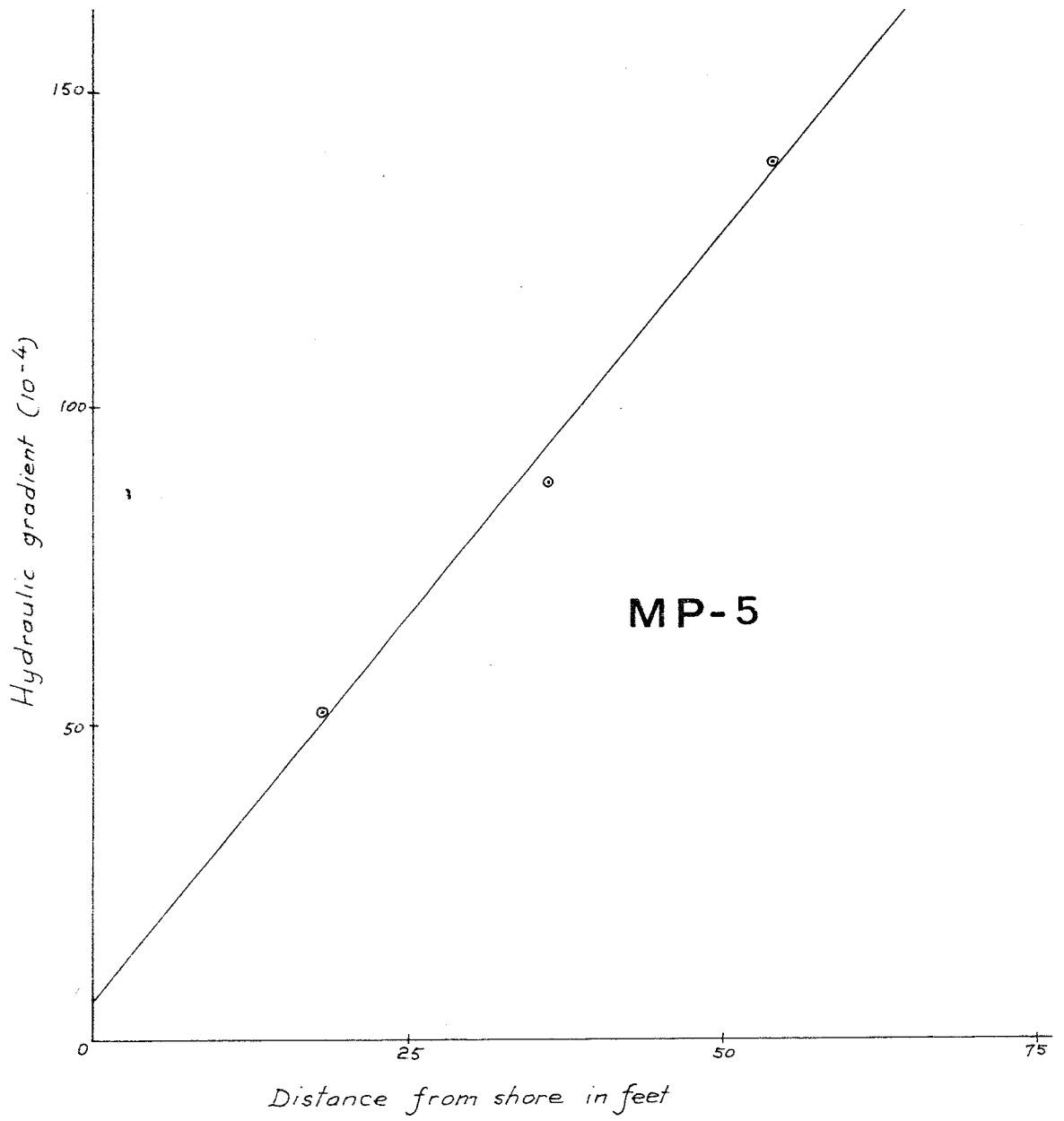
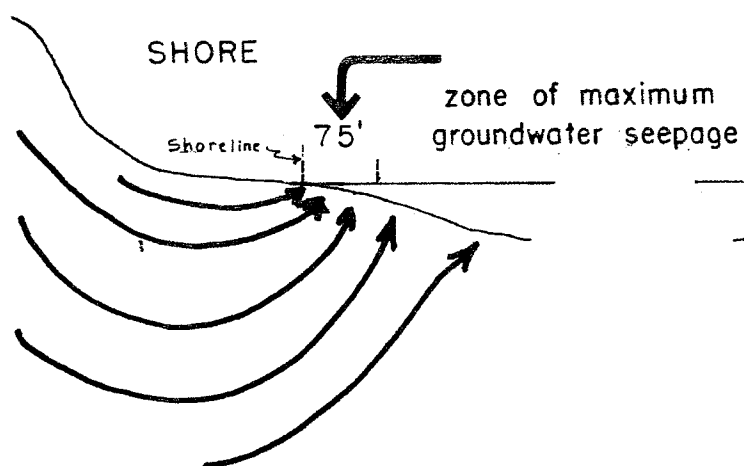


FIG.18e

a. SOUTH CENTRAL



b. NORTH CENTRAL

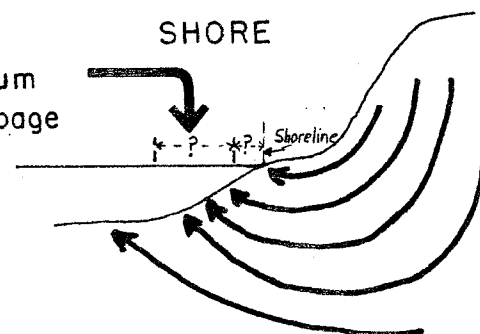


Figure 19 : Cartoon illustrating the model envisioned for the south and north central shores of Lake Wingra (see text for the details).

Because of a groundwater divide present in the eastern portion of the south shore (see Figure 11), the eastern south shore is hydrogeologically more complex; from west to east, the groundwater progressively changes from flowing into the lake, to parallel to the lake and finally away from the lake and into the east marsh (see Figure 11). The transition occurs in the vicinity of OS-2, where the vertical gradient is usually weakly negative, but can become slightly positive at times. Figure 17 attempts to depict the circulation between OS-12, OS-2 and the lake, however a true representation of the groundwater movement in this area cannot easily be represented by a two-dimensional cross section, since the flow lines are curved and change through time, and so the reader should note that generally the groundwater flow is toward the right (on Figure 11) or northeast at OS-12, but becomes progressively more out of the page or southeast around OS-2. Along the eastern south shore, the water exchange between the lake and the aquifer is minor, as the area is a transition zone between groundwater flowing into and out of the lake.

Along most of the south shore the horizontal groundwater flow always remains toward the lake, except in the east where the groundwater movement is dependent on the weather patterns; the flow is from the lake after a prolonged dry spell and toward the lake after a wet period, but it is in both cases negligible. The vertical groundwater flow component along the entire south shore changes and reverses according to the prevailing weather, but it remains minor at all times.

North

In terms of groundwater flow, the north shore resembles the south shore in many ways. Except in the Vilas Park and Zoo area, the north shore is characterized by a mostly horizontal lakeward groundwater movement. Superimposed on the horizontal flow is a slight upward component.

As in the south shore, the groundwater flow in the north central and northwestern shore becomes more vertical as the shoreline is crossed, however in this case, the zone of maximum upward (and perhaps also lateral) seepage into the lake is displaced farther off-shore, and does not occur right next to the shoreline as in the south shore (see Figure 19a,b); the groundwater seepage zone in the north is probably also relatively narrow. Indeed, mini-piezometers installed in the lake bottom along the north central and northwestern shore show an increase in vertical gradient with distance away from shore (see Figure 18d,e). Nonetheless, by the time IL-9 is reached, the vertical gradient has decreased.

One way of explaining this unexpected trend is by comparing the north central with the south central shore. In the latter, the vertical hydraulic gradient progression from OS-5, to OS-22 and finally MP-9, seems to indicate that the groundwater movement mimics the classic flow from a hill to a lake, where the flow is downward at the top of the hill (recharge area), then becomes horizontal near its foot (about 30 ft south of OS-5), and lastly seeps upward into the nearshore area of a

downgradient lake (discharge area) (see Figure 19a). The situation in the north central shore appears very similar eventhough the bulk of the Edgewood hill being probably composed of bedrock inferred to be intensively fractured, from observations made at surrounding outcrops. The apparent difference in the vertical gradient pattern observed between the two shorelines could possibly be explained by the following argument. Because the Edgewood hill is only 500 ft away from the shoreline (compared to 1000 ft for the south central hill), the groundwater flow in the north could be forced farther off-shore and, in doing so, the zone of maximum upward seepage into the lake could be displaced a short distance away from the shoreline (see Figure 19b). In other words, what is suggested here is that the ratio of the lengths of the vertical to the horizontal groundwater paths exceeds a critical threshold, inducing a lateral extention of the groundwater circulation some distance beyond the shoreline. Although this does not conform to the findings of McBride and Pfannkuch (1975), it is not incompatible with the suggestion made by Munter (1979), that the distribution of groundwater seepage into a lake depends on the geometry of the system. Differences in the ratio of the horizontal to vertical hydraulic conductivity of the lakebed materials may also effect the distribution of groundwater seepage in a lake (Munter,1979; Barwell and Lee, 1981).

Groundwater movement may also be controlled by fractures. However, Bradbury (1982) showed that the densely fractured Niagara Dolomite in northwestern Wisconsin behaves as a porous medium with respect to groundwater movement. This may be true of the fractured sandstone found

around Lake Wingra as well, and so would not account for the trend.

In the northwest shore, the increase in vertical gradient with depth at OS-16 (see Table 2) and Figure 15 suggest that in addition to the shallow horizontal groundwater seepage into the lake, a deeper groundwater flow percolates upward into the lake some distance away from the shore, within the groundwater in-seepage area in the western portion of the lake, shown on Figure 12. This situation for the northwest shore is in fact similar to that of the southwest shore.

The northeast shore of the lake is quite complex in that it represents, as does the southeast shore, a transition zone between groundwater flowing into and out of the lake. In west Vilas Park, the flow of groundwater is essentially southeast and downward toward and under Vilas Pond (see Figure 11 and Figure 14). In the east Vilas Park and Zoo area, the shallow groundwater flow seems to be deflected away from the lake, east and northeastward, in order to bypass the USGS spilling dam (not shown on Figure 14 since water would flow into the page). On the other hand, the deeper groundwater flow is forced under the shallow northeast arm of the lake, southeastward toward the east marsh. The northeast portion of the lake appears to gain some groundwater on the west side of Vilas Pond, while most of its remaining portion loses water to the underlying aquifer. These losses are minor except in the portion of the lake adjacent to the USGS dam.

The configuration of the groundwater circulation in the north central and northwestern shore is basically stable and no significant flow reversal takes place, even after a major storm or a period of

prolonged dry weather (one week or so), both of which occurred during the summer of 1981. In the northeast shore, flow reversals occur both horizontally and vertically in accordance with the prevailing weather: flow is away from the lake or pond and slightly upward during a prolonged dry spell, and toward the lake or pond and downward after a major wet period.

Lake

The groundwater seepage into or out of the lake bottom adjacent to the shoreline is discussed above. For the middle portion of the lake, the few data available suggest that the water exchange between the lake and the aquifer is minimal, e.g., there were no measurable vertical gradients recorded in these piezometers (see Table D-1 in Appendix D). The water level in all the in-lake piezometers fluctuated essentially as the lake level did, with perhaps at times a slight delay.

Low water tables

In several areas of upward groundwater seepage, the water table was consistently lower (e.g., OS-16) or sometimes lower (e.g., OS-11, OS-6, OS-23) than the lake surface. The low water tables are believed to be caused by small cones of depression engendered by the abundant vegetation present on the Arboretum shores of the lake. This idea has been advanced by several investigators including Meyboom (1966, 67) and

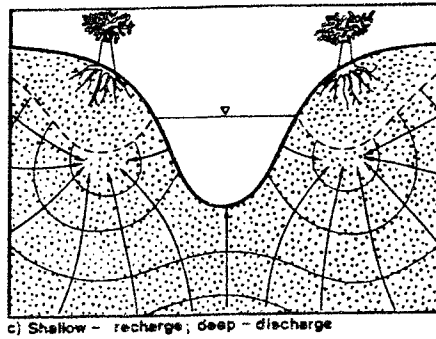


Figure 20 : Cartoon representing the groundwater circulation modified by plants in parts of the northwest, west and southwest shores of Lake Wingra during an extended dry period, taken from Born et al. (1974, p.26).

Born et al. (1974), and is illustrated in Figure 20.

The low water tables generally develop during extended dry weather periods. In those times, the west, northwest and southwest shores of Lake Wingra may have a temporary minor shallow, landward groundwater flow, superimposed on the generally lakeward groundwater movement, as shown in Figure 20.

Causes for the Short Term Fluctuations of the Groundwater Levels :
Precipitation

Introduction

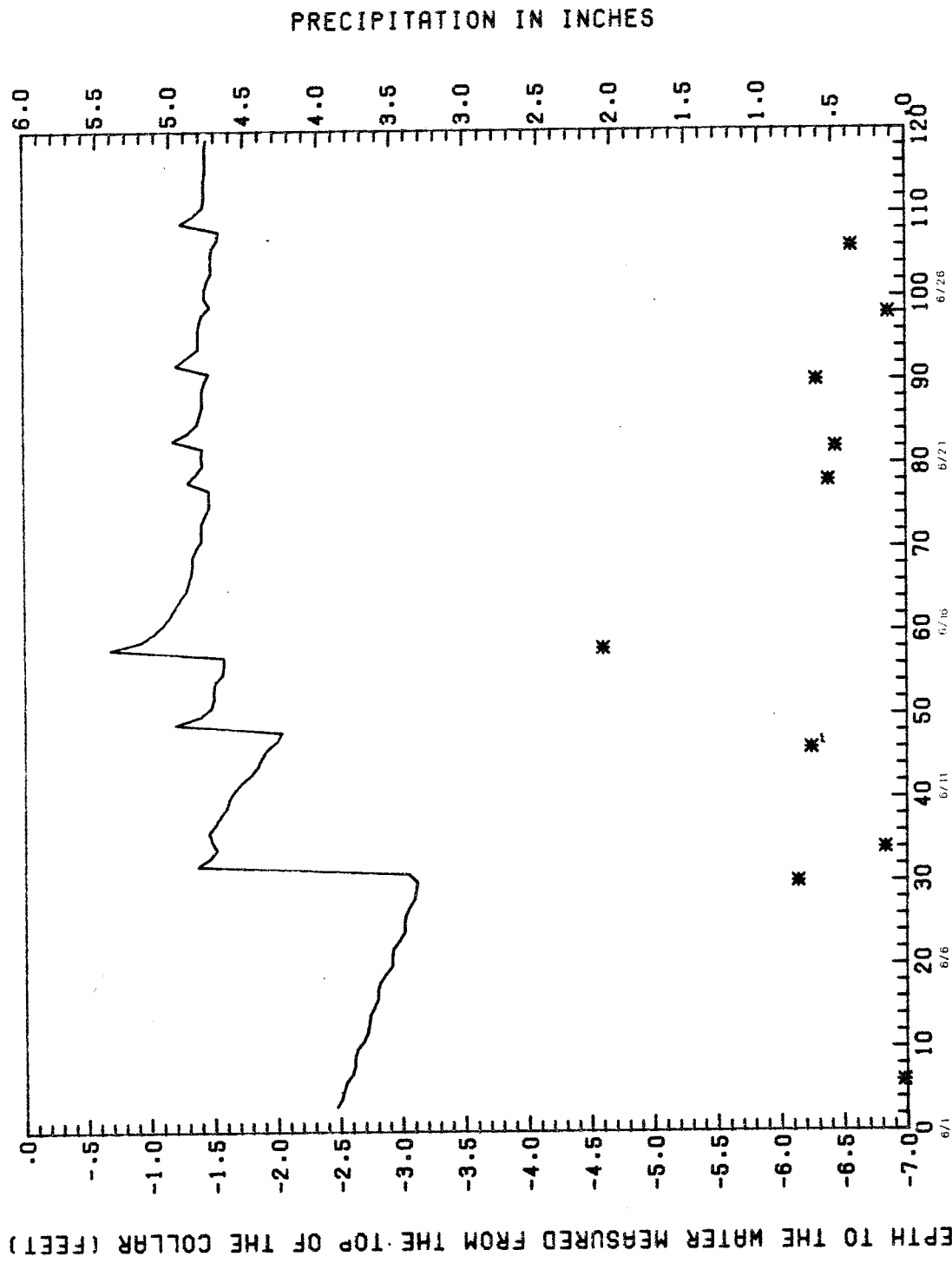
In the case where continuous records of rainfall and groundwater levels are available for all main regions in the area of study, it is relatively easy to evaluate the relationship between precipitation and groundwater levels, either by a simple visual inspection, or by a detailed analysis of both storm hydrographs and corresponding water level charts. However, such an ideal monitoring network was not available for this study and in fact is seldom available in a study with limited funding.

In the case of Lake Wingra, only one automatic water level recorder continuously monitored the water table in the west marsh (see Figure 5). Daily readings of a standard 8 inch precipitation gauge gave the total amount of rainfall which fell over the area during each previous 24 hour period. Figure 21 shows the daily precipitation and the elevation of the water table at the automatic water level recorder site, from June 1 to the end of September. The elevation of the water table obtained from the recorder charts was plotted in 6-hour increments. In this area however, the correlation apparently is reduced because often an extensive lakeward sheet flow forms after a storm, when certain combinations of climatic and soil conditions prevail. Nevertheless, a visual inspection of Figure 21 reveals that the elevation of the water

Figure 21 : Daily precipitation and automatic water level recorder charts plotted in 6 hour-increments for :

- a. June
- b. July
- c. August
- d. September.

The asterisks represent the precipitation in inches, and the solid line represents the water table elevation in feet below the top of the collar of the water table well monitored by the recorder.



PERIOD IN DAYS FOR JUNE, 1981

FIG. 21

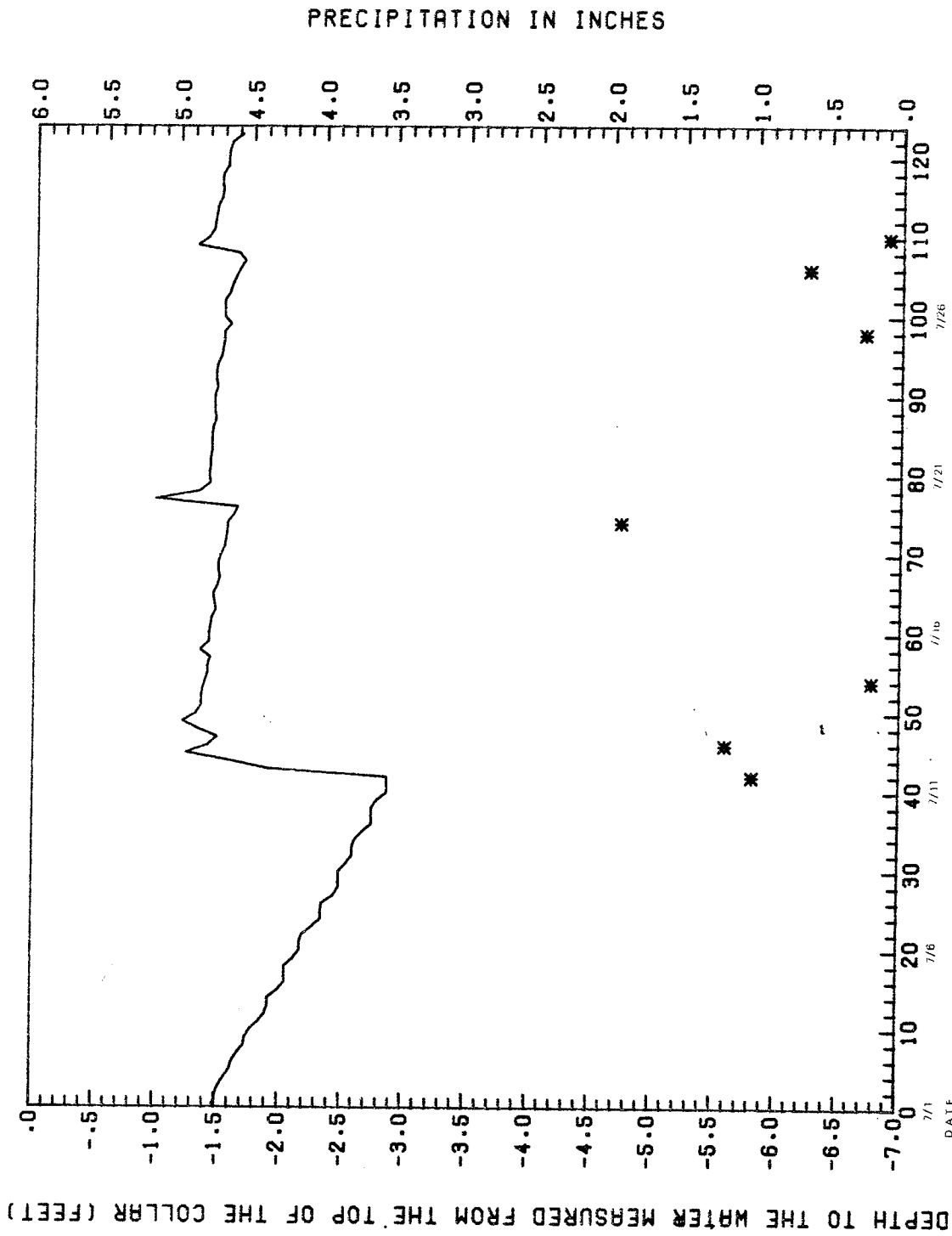


FIG. 21

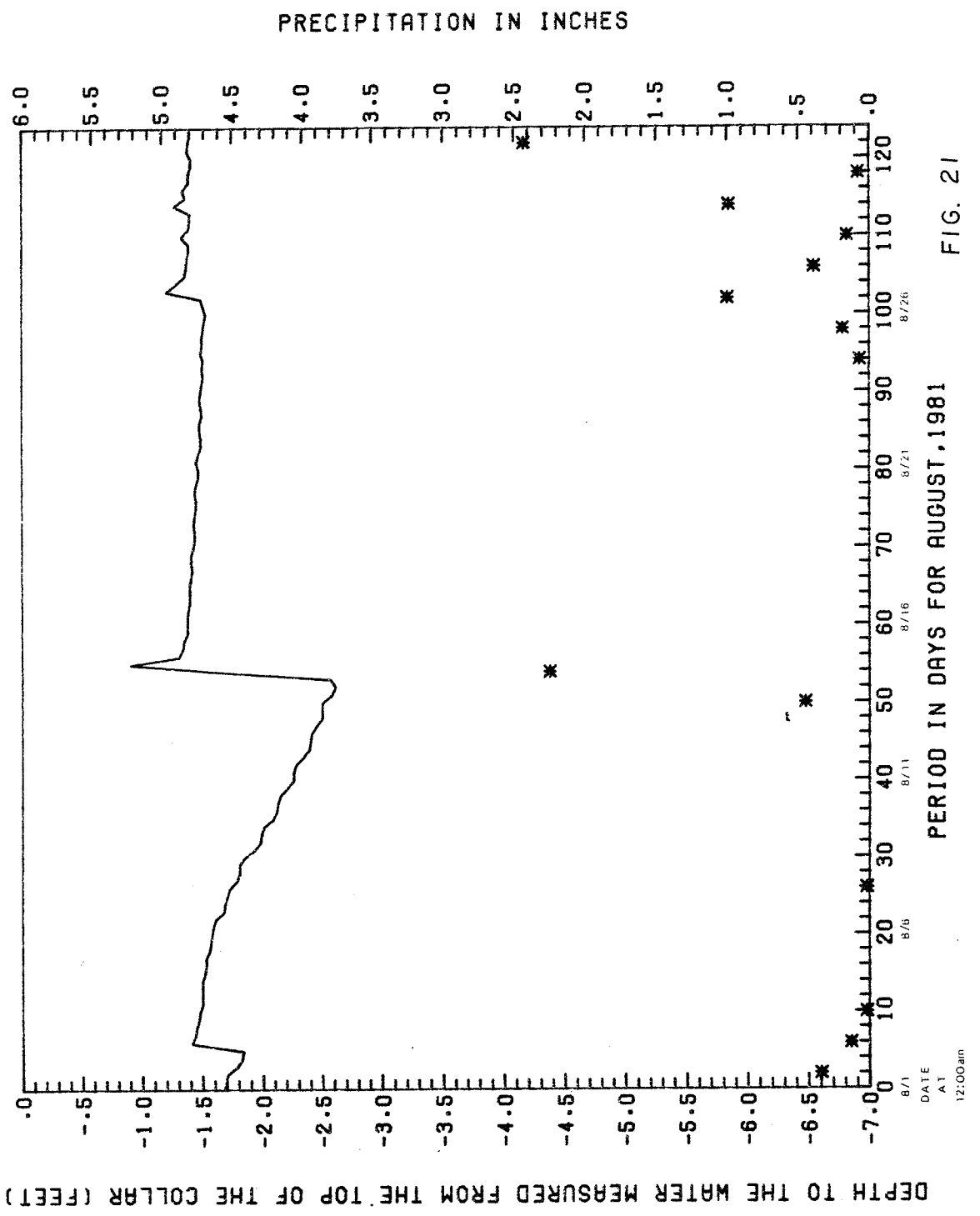


FIG. 21

DEPTH TO THE WATER MEASURED FROM THE TOP OF THE COLLAR (FEET)

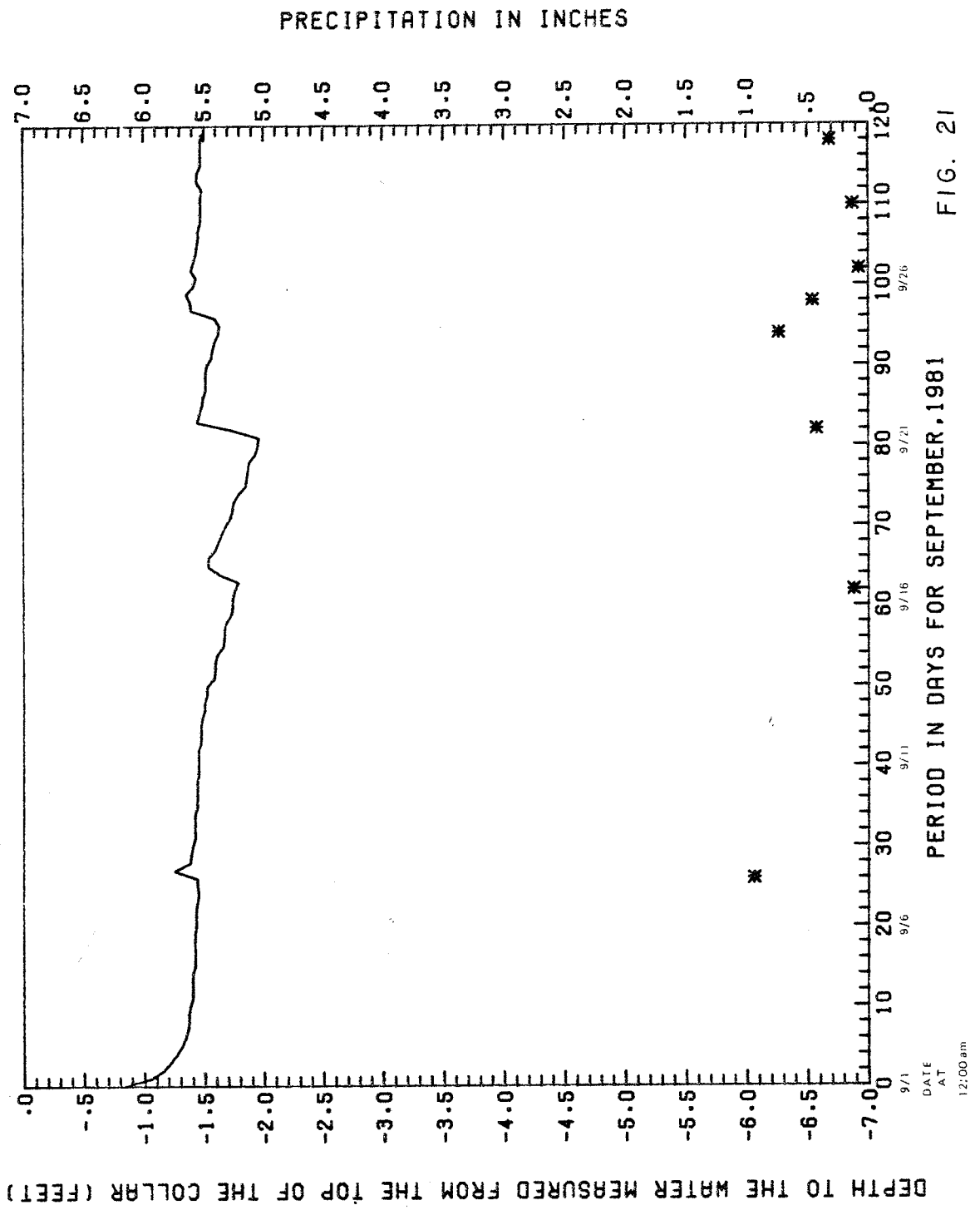


FIG. 21

DEPTH TO THE WATER MEASURED FROM THE TOP OF THE COLLAR (FEET)

table is strongly related to precipitation events. The question now arises as to whether this relation persists all around the lake, or if other factors control or partially control the groundwater potentials as well.

Problems in correlating groundwater levels with rainfall in the absence of continuous records

Since no other continuous groundwater level record is available for the rest of the Wingra area, the water levels measured in piezometers were used. However, the analysis is complicated by the sporadic nature of the rainfall and the discontinuous record available for the groundwater levels. One way to reduce the problem is to average the daily rainfall between water level measurements, but this technique still does not take into consideration the relative time which ellapses between the precipitation event and the measurements in the piezometers. Indeed, a measurement done shortly after a minor rainfall may record an apparently higher water level than if it were taken several days after a major storm. Along the same lines, the averaging method disregards any possible cumulative effects of many storms on the moisture storage in the ground. For example, a major storm at times of low water table could still engender lower groundwater levels than a minor precipitation event that occurs when the water table is high, although the water table has risen significantly more in the former case. Clearly, these problems seriously impede any correlation analysis that would attempt to relate rainfall with groundwater potentials.

What is needed is a formula that provides an estimate of the equivalent amount of precipitation that remains in the ground at any time between rainfalls. Such a formula would depend on both the time of occurrence of the precipitation event and on the initial amount of moisture in storage at the start of the event. One such formula is the antecedent precipitation index.

Before explaining the antecedent precipitation index, a few words should be said about indices in general. Indices are routinely used by engineers, often as black boxes, to describe hydrologic variables or processes which are difficult to measure, or to simplify computational methods by allowing observations to replace field data in time and space (Viessman et al., 1977). Often, these indices work, but why they work is not always fully understood. The following quote from Viessman et al. (1977) discusses the validity of an index, for example when applied to a drainage basin: " The adequacy of an index is based on (1) the ability of the index to adequately describe the physical process it represents, (2) the random variability of the observation, (3) the degree to which the point observation is typical of actual conditions, and (4) the nature of variability between the point measurement and basin means".

Antecedent precipitation index

The antecedent precipitation index, A_t , is roughly proportional to the quantity of groundwater storage and describes the rate of moisture depletion from a particular basin subjected to specific meteorological

conditions (Linsley, 1975). It assumes that most of the precipitation infiltrates into the ground and that the decrease of moisture with time between storms, is logarithmic and can be described by equation 2:

$$A_t = A_0 \alpha^t \quad (2)$$

where A_0 represents the initial value of the antecedent precipitation index, A_t , its value t days later and α , a recession factor normally ranging between .85 and .98 (Linsley, 1975), and representing the effects of both seepage and evapotranspiration (seepage probably dominates in the saturated zone, whereas evapotranspiration is more important in the unsaturated zone). In general, fine grained sediments require α -values that approach 1.00, whereas values closer to zero typify coarser grained material (Potter, personal communication, 1982).

The antecedent precipitation index assumes therefore that the rate of moisture depletion is mostly controlled by evapotranspiration, which is probably valid in most areas of limited surface runoff as suggested by Linsley (1975). Around Lake Wingra, some areas of the western and eastern marshes and wetlands are affected by significant runoff. Other regions where runoff could be important, although never observed, include parts of the flanks of the south and north central hills (see Figure 1). Elsewhere, surface runoff is minor and infiltration predominates; it is in this portion of the Wingra area that the antecedent precipitation index is potentially most powerful, although the approach has been used everywhere around the lake.

If t equals unity in equation 2, it can be seen that for any day

between storms, the index equals that of the previous day multiplied by the recession factor, α (Linsley, 1975):

$$A_1 = \alpha A_0 \quad (3)$$

and in a general form, the antecedent precipitation index on any day i , A_i , is:

$$A_i = \alpha A_{i-1} + P_i \quad (4)$$

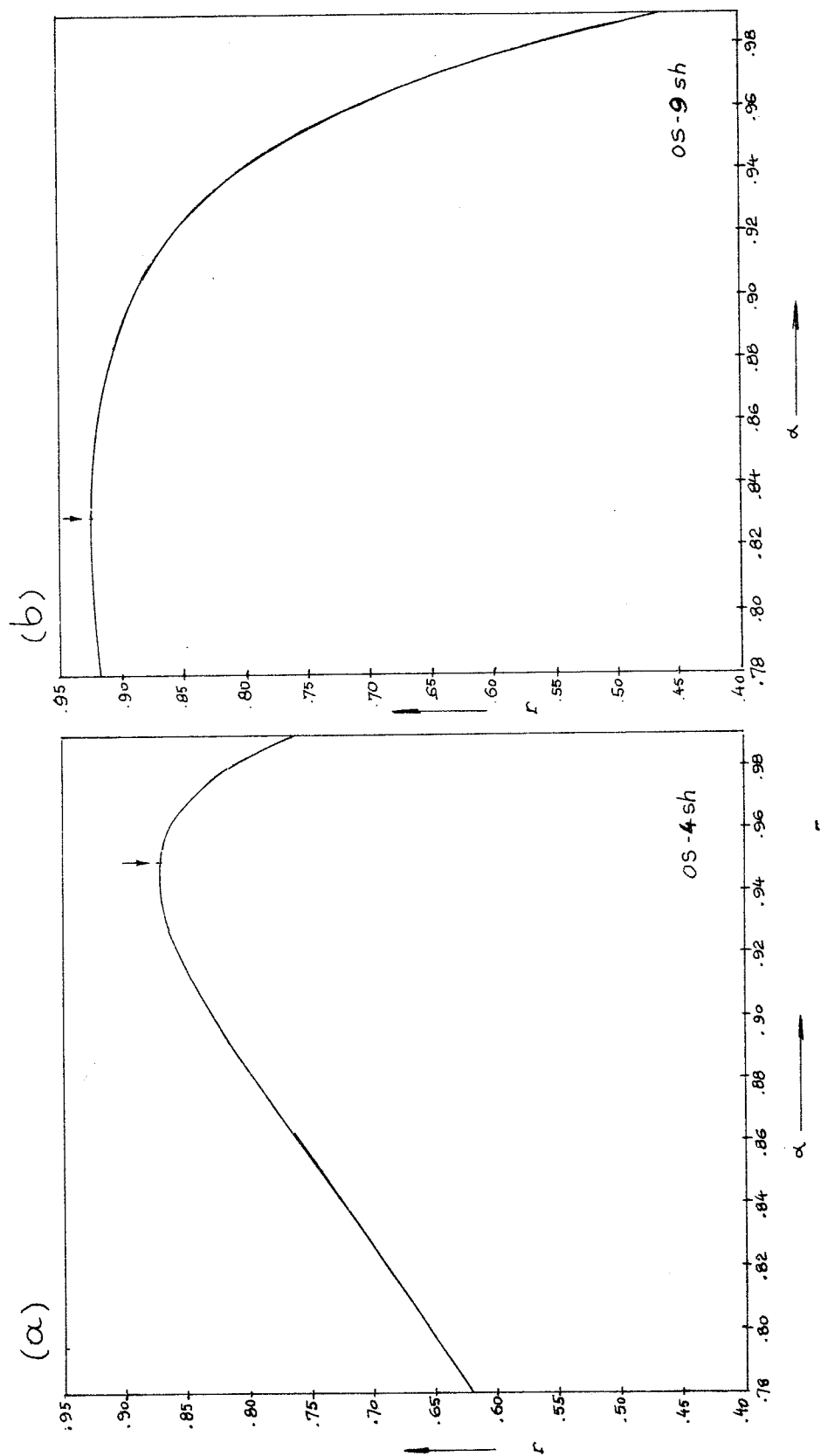
where A_{i-1} is the value of the index for the preceeding day, and P_i , the amount of precipitation which fell during day i . Linsley (1975) recommends the joint use of equations 3 and 4 with calendar date (by small adjustments of α), because the potential evapotranspiration varies with the seasons. But in view of all the possible sources of errors mentioned below, and the facts that α could represent mostly the effects of seepage and that the study basically was restricted to one season, the results obtained in this study did not justify the added computations that would be required for the slight improvement to be gained.

Alpha

If the antecedent precipitation index were known for the days when the water levels were measured in the piezometers, it would then be easy to evaluate the relation that exists (if any) between the groundwater

Figure 22 : Graphs showing the effects that different alphas have on the correlation coefficients. The correlation coefficient, r , which describes the correlation that exists between the antecedent precipitation index, A_i , and the groundwater potentials, depends on the α -value chosen to compute A_i . In some cases, as in Figure 22a for OS-4sh, there is a relatively well defined r optimum produced by basically one value of α . On the other hand, Figure 22b for OS-9sh shows that several α 's engender high values of r . The r optimum and associated α listed in Table 5 are indicated by an arrow on the graphs.

FIG. 22



potentials and the amount of rain that falls over an area, by simply regressing the antecedent precipitation indices against the groundwater levels. Nonetheless, an eventual correlation between the two would depend upon the chosen value of α , since α is probably mostly a function of seepage (saturated zone), and thus indirectly of the sediment type. For example, assuming that storms are the controlling factors in the short term fluctuations of the groundwater potentials, the record of water level variations for a piezometer installed in the sand would probably produce a low correlation coefficient (r), if linearly regressed against the antecedent precipitation index computed from an alpha value of .95. However, if the same water level record is linearly regressed against the index calculated with an α of .85, a higher r would result. Because of this, such a correlation analysis should be performed for several values of α , until an optimum correlation is reached for each piezometer. It is of course assumed that the optimum r reflects reality, and that the alpha(s) which generate(s) this r is/are the value(s) which best describe(s) the depletion of moisture in the sediments surrounding the piezometer. Figure 22 illustrates the point developed above, where different values of α engender different r 's for the shallow piezometers at OS-4 and OS-9.

Computer program and initial antecedent precipitation index, A_0

Because of the massive and tedious computations involved in performing a large scale correlation analysis, a computer program was

written to solve equation 4 and linearly regress the water levels recorded during the study, against the corresponding antecedent precipitation indices. The program appears in Appendix F.

The biggest problem was to determine the initial antecedent precipitation index, A_0 , since the value of the index at anytime theoretically depends on the precipitation over an infinite antecedent period (Linsley, 1975). A reasonable initial A_0 -value must then be chosen in order to avoid significant errors in the succeeding computed indices. Nevertheless, the nature of the index is such that, no matter which value is assumed for A_0 the computed indices will always converge toward their real values after a period of time, and will thereby eventually eliminate the errors introduced by a wrong initial choice. The rate of convergence depends nonetheless on both the assumed A_0 and on the particular recession factor, α ; the more reasonable the A_0 and the smaller the α , the faster the convergence (see Fig. 23 and 24).

To ease the problem of determining A_0 , equation 4 was solved repeatedly from January 8, 1981, to the end of September, for three α -values (.85, .95 and .98) and for 4 different reasonable A_0 -values assigned for January 7 (2,3,4 and 5 inches). The reason to start solving equation 4 so many days before the date of the first water level measurement of interest (May 15), is to take advantage of the natural convergence of the index, and to ensure that, for all practical purposes and for most α 's, convergence is generally achieved by May 15 (beginning of the summer study) regardless of the assumed A_0 . The results show that for an α of .80, the indices computed with the 4 different A_0 -values lie within .01 ft of each other after less than one month (see

Figure 23 : Antecedent precipitation index computed with four different A_0 -values (2,3,4 and 5 inches) for an α -value of .80.

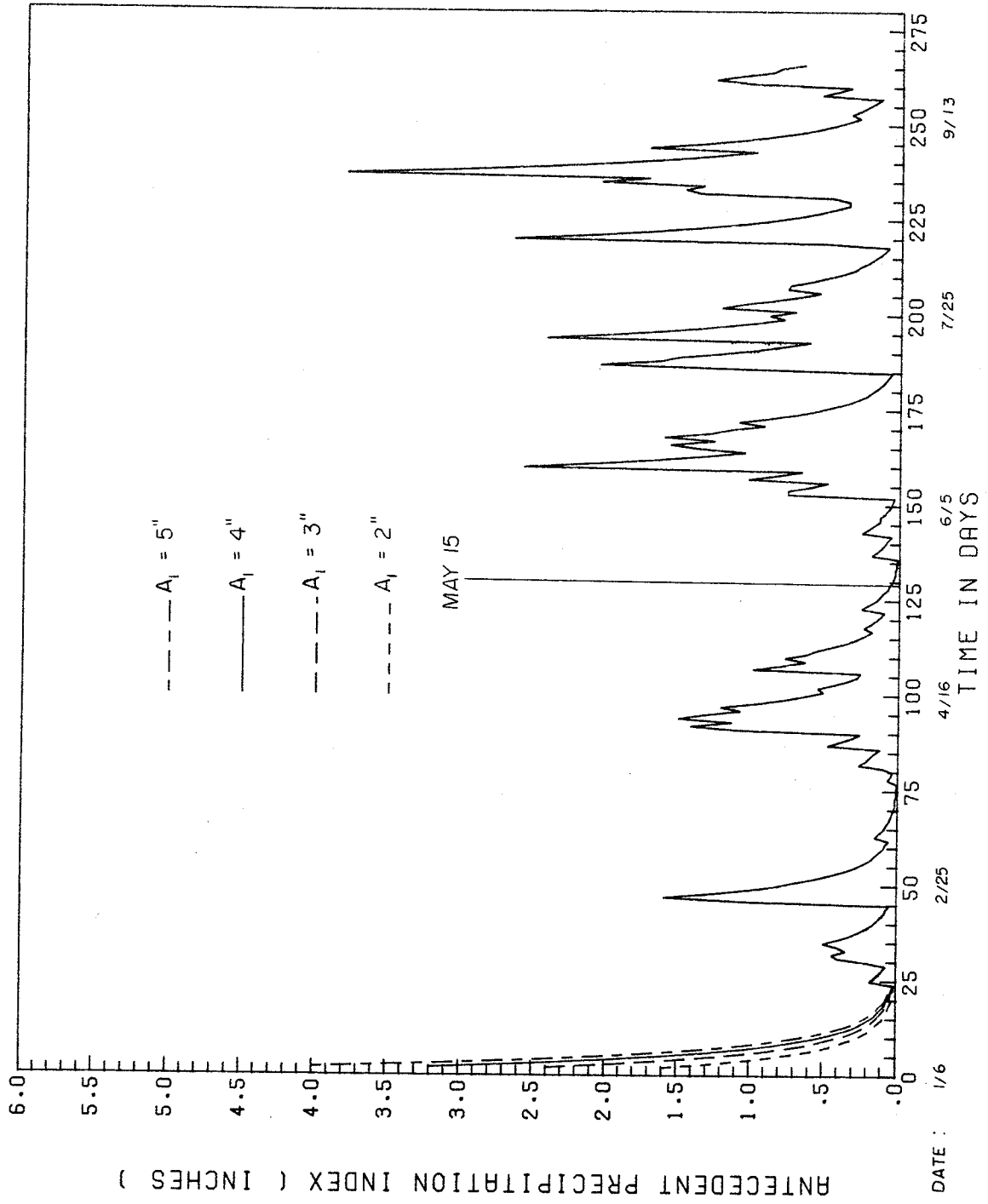
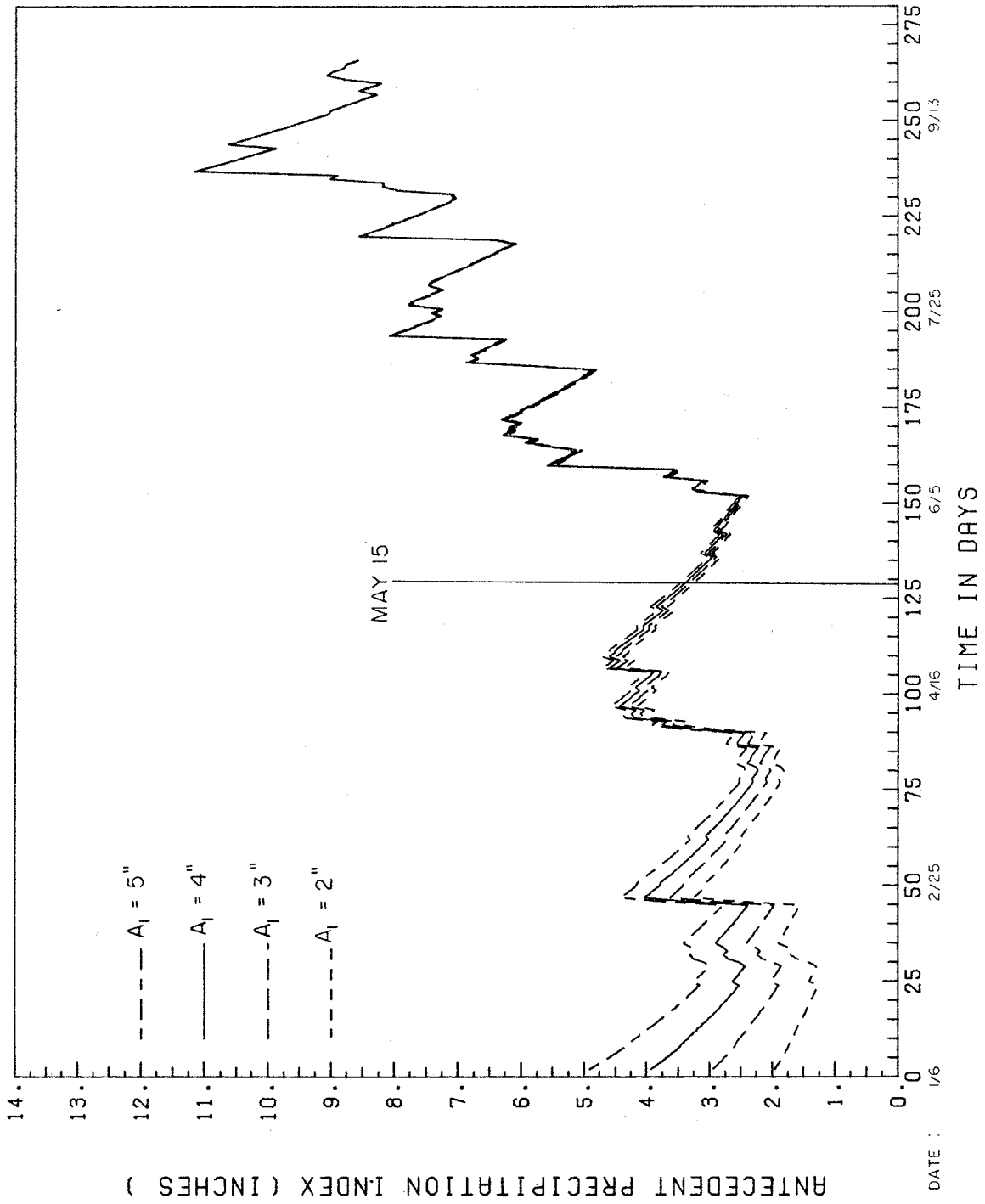


FIG. 23

Figure 24 : Antecedent precipitation index computed with four different A_0 -values (2,3,4 and 5 inches) for an α -value of .98.

FIG: 24



ANTECEDENT PRECIPITATION INDEX (INCHES)

DATE : 1/6

TIME IN DAYS

Fig. 23), and for an α of .95, convergence occurred after about 100 days. On the other hand, for the α of .98 the convergence to .01 ft was not reached until the end of September (see Fig. 24). Nonetheless, after about 128 days or at the beginning of the summer study, the computed indices agreed to within .23 ft of each other, and so the errors that would be introduced by choosing the wrong A_0 are minor, at least for alphas less than or equal to .98 (see Fig. 24). If α is greater than .98, the choice of A_0 becomes more critical. However, the only sediment layer which has an alpha value greater than .98 in the Lake Wingra area, is located at depth in the vicinity of OS-4, and the amount of groundwater storage here shows a very strong correlation with rainfall input when an A_0 of 4 inches is assumed (see Table 5). In view of all of the above arguments, A_0 was chosen to be 4 inches. The program was then run several times for each of 38 piezometers, using different alphas. In each case the output included both the daily antecedent precipitation index for the particular α , and the Pearson product moment correlation coefficient obtained in the regression analysis, for each piezometer.

Results and discussion

On the whole, the results show that a strong correlation exists between the short term fluctuations of the groundwater levels and the amount of rainfall, as shown in Table 5 by the relatively high correlation coefficients, r . These vary typically between .85 and .97 for the bulk of the shallow piezometers and for over half of the deeper

Table 5. Optimum correlation coefficients with their associated alpha-values obtained when the groundwater levels are regressed against the antecedent precipitation index.

SITE	PIEZOMETER	CORRELATION COEFFICIENT	ALPHA
East shore / marsh			
OS-1	Sh	.86	.88
	D	.76	.91
OS-3	Sh	.92	.89
OS-4	Sh	.87	.95
	ShI	.87	.97
	DI	.93	1.00
	D	.97	.99
OS-17	Sh	.87	.89
	D	.87	.93
OS-18	Sh	.93	.82
	D	.76	.87
OS-19	Sh	.89	.83
OS-20	Sh	.92	.85
OS-21	Sh	.90	.79
South shore			
OS-2	Sh	.91	.84
	I	.83	.90
	D	.73	.86
OS-5	Sh	.77	.83
	D	.73	.81
OS-22	Sh	.88	.85
	D	.91	.84
OS-8	Sh	.75	.43, .44
OS-23	Sh	.96	.83
	I	.94	.87
OS-24	Sh	.92	.82
	I	.93	.83
	D	.95	.85
West shore / marsh			
OS-6	Sh	.66	.94
	D	.47	.88
OS-9	Sh	.93	.83
	D	.55	.83
OS-11	Sh	.85	.82
	D	.54	.76

North shore

OS-7	Sh	.89	.87
OS-16	Sh	.82	.87
	I	.91	.85
OS-14	Sh	.86	.93
OS-15	Sh	.90	.88

ones. In addition to showing that a strong correlation exists between rainfall and groundwater levels, these high r 's also suggest that the antecedent precipitation index adequately describes the system. Most of the lower r 's obtained are not believed to reflect a reduction in the influence of rainfall on the short term fluctuations of the groundwater potentials, but rather are the result of any one or combination of errors or factors discussed below.

One of the most obvious factors that often decreases the r is the depth of the piezometer intake. If the screen is located far below the ground surface, there may be a significant lag time between the precipitation event and the adjustment of the local deeper groundwater potential. This situation may become even worse if less permeable sediments lie between the surface and the sand point; such is most likely the case for the deep piezometers at OS-9 and OS-11. Along the same lines, if the piezometer point is significantly clogged with fine sediments, bacteria or organic matter, or if it is located in a local clay lens, the response of the water level in the piezometer would be delayed, and apparently would diminish any correlation that exists between the amount of rainfall and the groundwater potentials. Clogging of the point explains the lower r obtained for the shallow piezometer at OS-16, and perhaps slightly reduces that of other piezometers as well.

In addition to the physical and mechanical factors mentioned above, there are factors directly related to rainfall and climate which significantly impair the results of the correlation analysis. If a piezometer is placed in an area where abundant surface runoff occurs during and after a storm, an apparent lower correlation coefficient or

α -value or both would result. Of the sites analysed, three are known to be prone to surface runoff including OS-6, OS-19 and OS-20 in the west and east marshes, and two others are suspected to be influenced by effects of surface runoff. These are OS-8 and OS-5, both on or near the slopes of the south central hill (see Fig. 1 for location). Variable evapotranspiration rates are another source of error which could influence the results. However this problem probably only becomes significant in studies that extend over several seasons. There are many other factors which could reduce the correlation, including variable groundwater seepage rates and errors in precipitation or water level measurements, but in view of the generally high r 's generated, they must be minor.

A few words should be said about the α -values and their distribution around the lake. In theory it is possible to extract information about the geology of an area by looking at the distribution of the α -values, since they partially depend on the grain size of the sediments (if α is mostly a function of seepage). In practice however, caution should be exerted in deducing geological information from these α -values; the r optimum obtained with one particular alpha is often not significantly greater than those generated by the adjacent α , and thus there are actually several α 's which may produce high r 's (see Figure 22b). For this reason and others mentioned above, several of the alphas listed in Table 5 should be regarded as one in a range of possible values.

Nevertheless, some trends are noticeable in Table 5, and support the geologic observations discussed in Chapter I. For example, on the

whole the sediments of the northeast shore area (northern portion of the east marsh sediments) seem to be represented by higher α -values than those of the north and south shores which are coarser grained.

Piezometers at OS-18, 19, 20 and 21 must be disregarded here since they are emplaced next to the Arboretum drive, and thus are affected by more permeable artificial fill. Higher α -values may also be encountered in the west marsh sediments. However the fact that only the three piezometers at OS-6 are installed in the true marsh sediments renders questionable any statement that could be made to that effect.

Beside precipitation events, intermittent pumping of water supply wells in the Lake Wingra area could also induce significant short term fluctuations of the groundwater potentials. As will be discussed in the following section, the City of Madison has essentially not pumped any of its wells that are near Lake Wingra during the late spring and summer of 1981. The only sizable pumping that took place in the area during the study period, was from a well located in the Nakoma golf course. Ordinarily, this well is pumped at very modest rates ranging between 20 (28,800 gal/d) to 22 gal/min (31,680 gal/d), but occasionally rates of up to 800 gal/min (1.2 mgd) are reached during dry periods when irrigation is needed (Smith, personal communication, 1982). Since it is located both upflow and only a few hundred meters away from OS-6, it is possible that the effects of the higher pumping rates are felt in the piezometers at this Wingra marsh site. This could explain why the piezometers at OS-6 show the lowest r , but unfortunately the lack of data prevents a meaningful multiple correlation analysis which would separate the effects of pumping from those of rainfall on these

piezometers.

Causes for the Long Term Variations of the Groundwater Levels:

Large Scale Pumping of the City Wells

Long term variations of the groundwater levels are much more difficult to assess because of first, the lack of data, and secondly, the short duration of the present study. Earlier studies however have hinted or theoretically demonstrated that large scale city pumping of the sandstone aquifer influenced the groundwater levels in the shallow aquifer. Mcleod (1975b), for example, showed with the aid of a finite difference computer model that, in the long term, continued pumping of the city wells would not only lower the groundwater potentials in the sandstone aquifer, but would have a similar effect, although reduced, in the upper aquifer. Oakes et al. (1975) reported that after two days of pumping by a nearby city well (at a rate of about 60 l/s or 1.4 mgd), the total discharge at Sp-1a and Sp-1b declined from about 12 l/s (273,894 gal/d) to 9 l/s (205,420 gal/d).

One objective of the present study was to supply field evidence to evaluate the effects that large scale city pumping has on the upper aquifer and on the lake level. However, the difficulty encountered in coordinating efforts with the Madison Water Utility prevented the accomplishment of the proposed task. City well #1 (see Figure 5) has not been actively pumped since the publication of findings by Mcleod

(1975a,b) and Oakes et al. (1975), and could not be used for a pumping test. Pumping from the next closest city well to Lake Wingra, unit well #2 located about 1 km northeast of IL-2 and IL-3 (see Figure 5), basically ceased pumping at the onset of the late spring and summer study, although it had been pumped almost continuously in the last several years, and especially during the summers.

Nevertheless, the lack of pumping during the summer brought about an interesting observation: the water levels in the two deep in-lake piezometers at IL-2 and IL-3, were significantly lower in February, when city well unit #2 was pumped at a more or less constant rate of about 3 mgd, than during the summer (see Table 3). By contrast, the western in-lake piezometers about 2 km (1.24 mi) southwest of the pumping well on the whole did not show any significant difference between winter and summer groundwater potentials (see Table 3). Conclusions based on this observation should be made with caution because of the paucity of data.

Conclusions

The groundwater circulation in and around Lake Wingra is summarized in Figures 11 and 12. The groundwater system around the lake is permanently in a transient state; the water levels in the wells rise with a precipitation event, reach a peak and then fall until the next storm. However, the overall configuration of the groundwater system is relatively constant, although sometimes locally distorted or reversed

(e.g., northeast and southeast shore) by major storms or droughts. During extended dry spells, minor local groundwater flow reversals induced by small cones of depression caused by transpiration of the shoreline, may be superimposed over the main groundwater movement in some areas.

Short term fluctuations of the groundwater potentials in the upper aquifer appear to be controlled primarily by variations in precipitation, at least in the absence of large scale sporadic pumping. On the other hand, the cause of long term variations in the groundwater levels is more uncertain, but the continued large scale pumping of the City wells very likely contributes to these fluctuations.

There is a far greater potential for use of the antecedent precipitation index in groundwater studies, beyond the simple correlation of water levels with precipitation. For example, if the groundwater movement is mostly controlled by precipitation in a shallow aquifer (i.e., if the water level in wells regressed against the antecedent precipitation index show high correlation coefficients), than it should be possible to establish a relationship between the groundwater budget for an area and the antecedent precipitation index. A simple program as the one in Appendix F could compute the daily precipitation index, and the daily groundwater budget for the lake might be easily estimated from the established relation between the budget and the index. Likewise in studies of groundwater circulation and groundwater flow rates, the daily antecedent precipitation index might supply information on the daily groundwater circulation or flow rates, provided again that the relation between water levels in different

piezometers correlates with the index (i.e., the water level in wells can be estimated from the index), so that the hydraulic gradients can be evaluated. This aspect is especially attractive in studies of groundwater monitoring in hydrogeologically unstable areas and studies involving monitoring the migration of pollution plumes.

V. GROUNDWATER FLOW MODEL

Introduction

Computer modeling was undertaken in order to answer a few questions regarding the geologic configuration of the disturbed northwest portion of Gardner Marsh and to obtain information needed to calculate the groundwater budget for Lake Wingra. Specifically with regard to Gardner Marsh, logs from two piezometers suggested that although the unconsolidated geology in Vilas Park is represented essentially by one sedimentary layer, the northwest portion of Gardner Marsh consists of two layers, at least to a depth of about 35 feet. It was hoped that a computer groundwater flow model could provide important information, not only on the hydrogeology of the system, but also on the geology, or more specifically, on the location of simple geologic contacts. Extracting hydrogeologic information from groundwater flow models is now standard practice (e.g., see Freeze and Witherspoon, 1967; McBride and Pfannkuch, 1975; Winter, 1978; Munter, 1979; Remson et al., 1980; Munter and Anderson, 1981...). However, an application of these models to establish the shape and location of geologic contacts in simple geologic settings has not been published in the open literature accessible to the author. Remson et al. (1980), however did obtain some geologic

information for a fault in Utah, using a groundwater flow model.

Purpose and Scope

The purpose of this portion of the study is five fold: 1. to test the suspected two-layer marsh configuration, with a low hydraulic conductivity layer at the bottom. 2. to test the assumption that the observed increase in vertical gradients with depth in the eastern and possibly the western portions of the lake, could be caused by the presence of a low hydraulic conductivity layer at depth. 3. to estimate the difference in horizontal hydraulic conductivity between the upper and lower marsh layers in northwest Gardner Marsh. It is important to know relative hydraulic conductivities for the groundwater budget calculation. If the lower marsh layer is relatively impermeable compared to the upper marsh layer, the depth of interaction between the lake and the aquifer would be approximated by the thickness of the upper marsh layer (see Chapter VI). 4. to evaluate the influence of the configuration of geologic contacts on the equipotential lines or head distribution, if the ratio of horizontal hydraulic conductivity (K_h) between the high and low conductivity layers equals or exceeds 10. 5. to determine the average ratio of horizontal to vertical hydraulic conductivity in sediments around Lake Wingra. This is a key parameter in computing a groundwater budget for a lake (Munter and Anderson,

1981).

Model

The finite-difference two-dimensional groundwater flow model developed by Trescott et al. (1976) was used. The governing equation is the general two-dimensional partial differential equation describing groundwater flow. If the cartesian coordinates x and y are aligned with the principal components of the transmissivity tensor, T_{xx} and T_{yy} , this equation may be written:

$$\frac{\partial}{\partial x} \left(T_{xx} \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(T_{yy} \frac{\partial h}{\partial y} \right) = S \frac{\partial h}{\partial t} + W(x,y,t) \quad (5)$$

Where; T_{xx} and T_{yy} are the principal components of the transmissivity tensor

h is the hydraulic head

S is the storage coefficient

$W(x,y,t)$ is the volumetric flux of recharge or withdrawal per unit surface area of the aquifer

A finite-difference form of this equation is written for a set of nodal points, and the resulting set of linear algebraic equations is solved numerically for the head at each node in the grid designed for the model.

In this study, steady state simulations were done and both the storage coefficient (S) and the recharge-discharge (W) terms were assigned a zero value. All simulations used the SIP computation method with a β of 1 (see Trescott et al., 1976 for details).

Grid

Cross section B-B' was modeled for August 1981 (see Fig. 25). The cross section is 3888 feet in length and extends 35 feet below the water table. The grid designed to represent the cross section consists of 27 columns and 14 rows of nodes, with a variable horizontal nodal spacing ranging from 101 feet to 173 feet, and a fixed vertical nodal spacing of 2.5 feet (see Fig. 26). The cross section was assumed to have a unit thickness, so as to substitute directly hydraulic conductivity values (K) for the transmissivity tensor (T).

Boundary conditions

The cross section was surrounded by constant heads (bottom

Figure 25 : Groundwater circulation around Lake Wingra for August 1981, and location of cross section B-B'. Contour interval, 1 foot.

included) that were estimated from the August lake and pond elevations, and from the average August heads recorded by the 12 in-lake and onshore piezometers included in the cross section. Linear interpolation or extrapolation was generally done to estimate heads between control points (known heads), and between control points and the boundaries (bottom boundary included) of the grid.

Steady state calibration

To calibrate the model, three parameters were varied; the configuration of the geologic contacts between sedimentary facies, the horizontal hydraulic conductivity contrast between the various sedimentary facies and the ratio of horizontal to vertical hydraulic conductivity (K_h/K_v).

The geologic contact between the sediments in Vilas Park and the marsh sediments was varied from oblique, to straight and vertical, and finally was eliminated altogether by assuming the absence of marsh sediments. The contact between the upper and lower marsh sediments was changed from straight and horizontal, to dipping, and finally eliminated altogether.

The range in hydraulic conductivities used in the model (2×10^{-7} to 6×10^{-5} ft/s) was basically restricted to the range of values obtained from slug tests (2×10^{-7} to 9×10^{-5} ft/s). In a steady state simulation it is not the absolute values but rather the relative values of

hydraulic conductivity among the different sedimentary systems that determine the groundwater flow pattern (Munter, 1979). In this study the highest hydraulic conductivity value used was 6×10^{-5} ft/s, which corresponds to the average conductivity obtained for the northern, southern and parts of the eastern shores of Lake Wingra (see Chapter VI). All other conductivities used were scaled relative to 6×10^{-5} ft/s to create the desired contrasts. The tested horizontal hydraulic conductivity contrasts between the different geologic settings varied from:

1/200 to 250/1 for the contrast between Vilas and upper marsh sediments.

1/20 to 250/1 for the contrast between Vilas and lower marsh sediments.

1/20 to 200/1 for the contrast between upper marsh sediments and lower marsh sediments

Since neither the K -value for the bedrock (see Fig. 27), nor the exact configuration of the contact between the bedrock and the sediments in Vilas Park were known, 6×10^{-5} ft/s was assumed to represent K_h for the bedrock. This eliminated the need to determine the exact shape and location of the contact between the bedrock and the sediments in Vilas Park. The decision to assume a K_h of 6×10^{-5} ft/s for the bedrock was made because bedrock (sandstone) samples obtained from drilling at OS-15 (see Fig. 25) were highly weathered and showed a sand grain texture

similar in size to the sand found in Vilas Park. For convenience in this discussion, the Vilas sediments, whenever mentioned, will refer to both the unconsolidated and consolidated material found in Vilas Park. Finally, the ratio of horizontal to vertical hydraulic conductivity (K_h/K_v) for each sedimentary layer was varied from 1/1 (isotropic) to 1000/1 (greatly anisotropic).

Results and Discussion

The calibrated steady state model, as well as the results obtained from field data for August, are shown in Figure 27 and in Table 6, which compares the observed heads to those predicted by the model for the 12 piezometers in cross section B-B'. Table 7 lists the K_h and the anisotropy used for each sedimentary layer in the calibrated model. The model suggests that indeed the top 35 feet of marsh sediments are composed of two layers, the one at the bottom being a low conductivity layer relative to the one above. The increase in vertical hydraulic gradient with depth is therefore explained by the presence of the low conductivity layer or finer grained sediments (true marsh sediments) at depth.

If the horizontal hydraulic conductivity (K_h) for the sediments in Vilas Park is 6×10^{-5} ft/s, then the calibrated steady state model shows that the upper marsh layer must have about the same K_h -value (Table

Figure 26 : Finite-difference grid used for cross section B-B'. The black dots represent the locations of the piezometer screens.

Figure 27 : Comparison between the results of the computer simulation and field observations for August in cross section B-B'. The dashed lines are the equipotential lines generated by the computer and the thick solid lines are equipotential lines obtained from field data. The step-shape contact is the contact between the upper and lower marsh sediments. Contour interval, 0.5 and 1.0 foot.

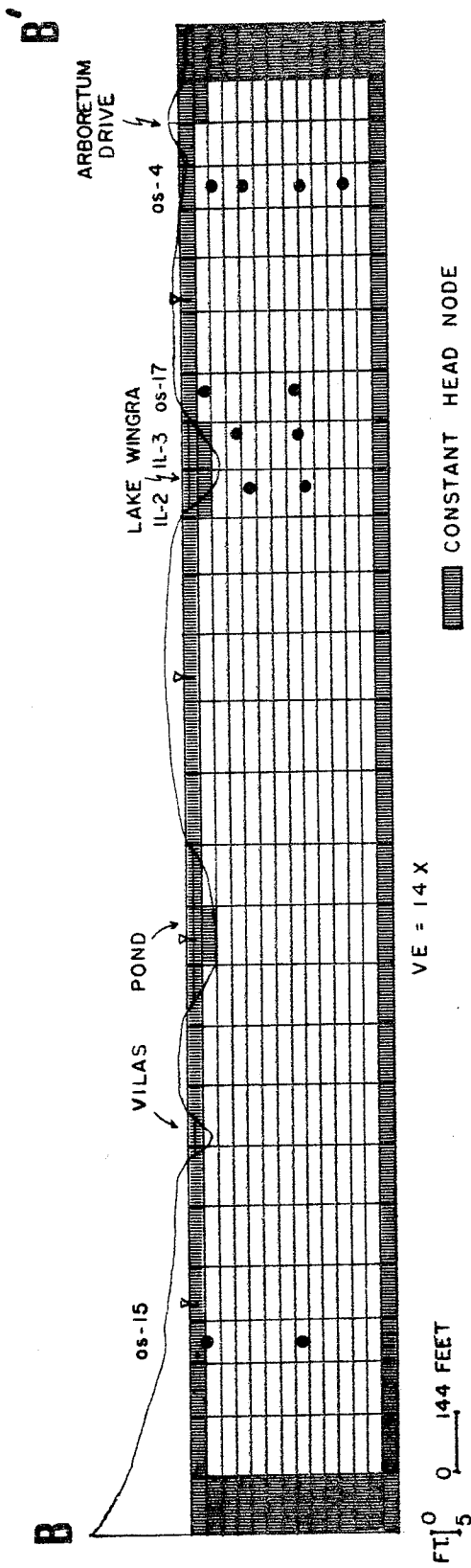


FIG. 26

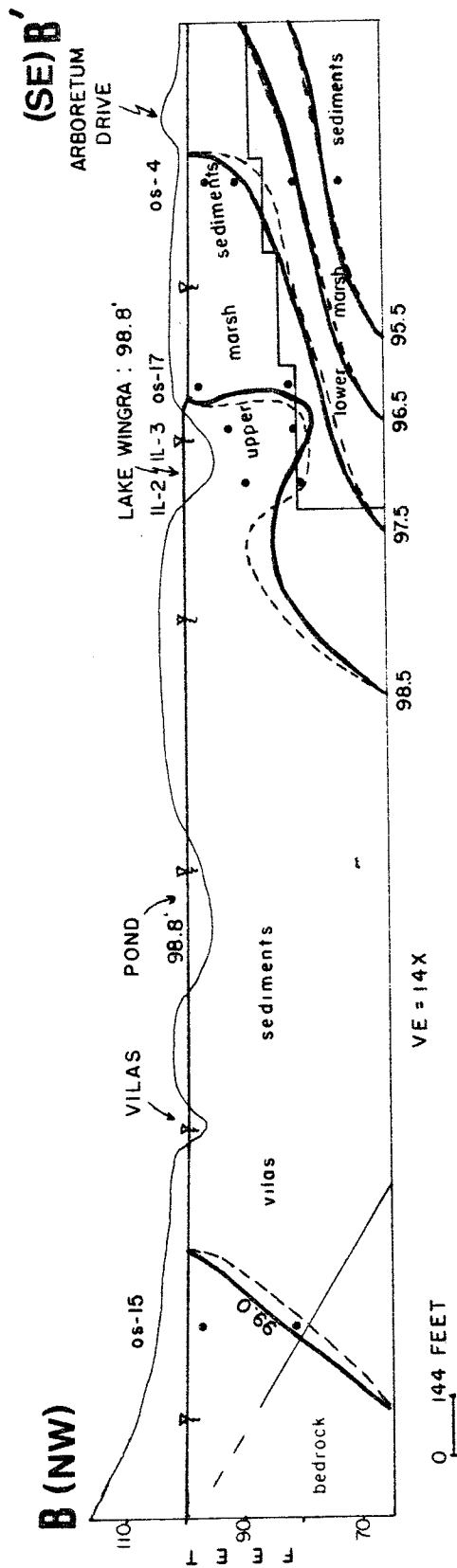


FIG. 27

TABLE 6 : Comparison between hydraulic heads predicted by the model for August 1981 and the August heads observed in the field for the 12 piezometers included in cross section B-B'.

PIEZOMETER	OBSERVED HYDRAULIC HEAD, h_o (feet)	PREDICTED HYDRAULIC HEAD, h_p (feet)	$h_p - h_o$ (feet)
OS-4Sh	97.8	97.9	+ .1
OS-4ShI	97.6	97.8	+ .2
OS-4DI	96.8	96.8	0
OS-4D	95.0	95.2	+ .2
OS-17Sh	98.3	98.3	0
Os-17D	98.5	98.3	- .2
OS-15Sh	99.1	99.1	0
OS-15D	99.0	99.0	0
IL-2Sh	98.8	98.8	0
IL-2D	98.4	98.6	+ .2
IL-3Sh	98.8	98.7	- .1
IL-3D	98.8	98.6	- .2

Table 7 : Horizontal hydraulic conductivity (K_h) and ratio of vertical to horizontal hydraulic conductivities ($K_v:K_h$) used for each geologic layer in the calibrated model.

GEOLOGIC LAYER	HORIZONTAL HYDRAULIC CONDUCTIVITY, K_h (FT/S)	VERTICAL HYDRAULIC CONDUCTIVITY, K_v (FT/S)	RATIO OF VERTICAL TO HORIZONTAL HYDRAULIC CONDUCTIVITY, $K_v:K_h$
Vilas sediments and bedrock	6×10^{-5}	6×10^{-6}	1:10
Upper marsh	6×10^{-5}	6×10^{-6}	1:10
Lower marsh	3×10^{-7}	3×10^{-8}	1:10

7). This argues that although somewhat different geologically (see Chapter I), hydrogeologically the Vilas and the upper marsh sediments are similar. On the other hand, the model estimates that the lower marsh sediments have a K_h -value of 3×10^{-7} ft/s. This value is very close to the value obtained from the only slug test conducted in the lower marsh sediments, at OS-4D (see Table G-1 in Appendix G). The horizontal hydraulic conductivity of the lower marsh sediments is therefore 1/200 of that of both the upper marsh layer and Vilas Park sediments. Freeze and Witherspoon (1967) reported that if an hydraulic conductivity contrast of 1 to 10 exists between two layers, the layer having the lower conductivity may be considered as impermeable relative to the higher conductivity layer. This means that the lower marsh sediments act as a relatively impermeable barrier to the groundwater flowing out of the lake, and so the depth of interaction between the lake and the aquifer can be assumed to be limited to the thickness of the upper marsh sediments, or about 15 feet.

The model proved to be very sensitive to the location and shape of the different geologic contacts between the hydrogeologically different layers (at least, if K_{h1}/K_{h2} is greater or equal to 10). To obtain a calibrated model, the contact between the Vilas and the lower marsh sediments must be nearly vertical and straight, and located close to the north shore of the northeast arm of the lake (see Fig. 25 and 27). The existence of a vertical contact is in fact likely, since Vilas Park is almost entirely composed of fill material that replaces old marsh sediments (Baumann et al. 1974), whereas south of the northeastern arm

of the lake the fill material is mixed with marsh sediments and does not extend to more than about 20 feet below the water table. On the other hand, the model suggests that the contact between the upper and lower marsh layers dips northwesterly (at least in cross section B-B'), and rises from about 20 feet below the water table at the north shore of the northeast arm of Lake Wingra, to about 10 feet below the water table at point B'.

Finally, the model suggests that the ratio of horizontal to vertical hydraulic conductivity or average anisotropy of most sediments around Lake Wingra, is about 1 to 10 (see Table 7).

With every computer model solution the questions of uniqueness and reliability arise; is the solution unique and does it simulate the real system? In many cases when field data are scarce, the hydrogeologic parameters can be combined in several different ways to produce a solution that would satisfy the few available control points (i.e., known hydraulic heads). However, in this study, because of the good knowledge for the range of possible hydraulic conductivities, combined with good control of the hydraulic heads at different depths throughout the sedimentary system, and good geologic logs, the groundwater flow model presented in Figure 27 and Table 7 most probably simulates the real system, and compares rather well with the groundwater flow determined from field data, even in the northwestern portion of cross section B-B' (see Figure 27) where several assumptions regarding the hydrogeologic nature of the bedrock had to be made (see Table 6). It is believed that a finer grid, especially under the lake area, would have

further improved the results.

Conclusion

The 2-dimensional model presented here simulates rather well the August groundwater flow in cross section B-B' determined from field data, especially in the southeast half of the cross section (see Figure 27). The model allowed to determine important hydrogeologic parameters (i.e., K_h , $K_v:K_h$, depth of interaction between the aquifer and the lake.) needed to calculate the groundwater budget for Lake Wingra, and thus again proves the importance of groundwater flow models as tools that can be used in conjunction with other calculations to help compute the groundwater component of a lake's budget.

In addition to gaining hydrogeologic information, this study suggests that by using groundwater flow models, it is possible to reconstitute the general subsurface geology and the configuration of geologic contacts in a fairly simple geologic setting, provided some information on the geology (e.g., a couple of well logs), reliable hydraulic head measurements for different depths into the aquifer system and a general knowledge of the horizontal hydraulic conductivities are available.

VI. GROUNDWATER BUDGET

Introduction

Two earlier studies (Oakes et al., 1975, and Novitzki and Holmstrom, 1979) have established a complete water budget for Lake Wingra. However, the groundwater component of that budget was neglected due to a lack of data. Because of this and because of uncertainties in estimating the water balance of lakes (Winter, 1981), efforts in this study concentrated on only establishing a groundwater budget for the lake.

In order to minimize the number of assumptions needed to determine the groundwater budget for the lake, an approach using data from both in-lake and onshore piezometers was used. Wherever the groundwater circulation through the shore is mostly horizontal (e.g., north and south shore and east shore through the Arboretum Drive) and/or if no or meager hydraulic gradient data existed for the sediments beneath the lake, the amount of groundwater seepage in or out of the lake was calculated using the onshore piezometers. If the groundwater movement is predominantly vertical through the shore (e.g., west and northeast shore), and if the vertical hydraulic gradient was well known for the sediments under the lake, the quantity of groundwater seepage was determined using in-lake piezometers and mini-piezometers.

Method

Depending on the approach used (in-lake or onshore), the lake and the shore affected by groundwater in or out-seepage, were first divided into several sub-areas and segments, respectively (see Fig. 28); this subdivision was based on Figure 12, on hydrogeologic (hydraulic gradient, hydraulic conductivity and piezometer location) and geologic criteria. Next, an average vertical hydraulic gradient and hydraulic conductivity were defined for each lake sub-area. Similarly, a typical or average horizontal hydraulic gradient and hydraulic conductivity were chosen for each shore segment. When all the lake sub-areas and shore segments were hydrogeologically defined, Darcy's law ($Q = KAh/\Delta l$, where Q is the groundwater volumetric flow, K , the hydraulic conductivity, A , the cross-sectional area through which the groundwater flows and $\Delta h/\Delta l$, the hydraulic gradient) was used to compute the groundwater seepage into or out of each lake sub-area, or to or from each shore segment.

Originally, the objective was to calculate a groundwater budget for each season of the year. However, the initial two year grant was changed to one year sometime after the study was underway. Furthermore, mechanical and logistic problems combined with winter weather delayed the installation of the groundwater monitoring network for several months. Because of this, the groundwater budget was calculated for the summer, 1981 (July through September), and for August, 1981, to emphasize the stability of the groundwater system around Lake Wingra.

The few data gathered during the winter and in the spring suggest nonetheless that the groundwater budget for the lake does not change very much from one season to the other. Figure 28 shows the set up used to calculate the groundwater budget for the lake, and the following discussion explains the details of the calculations.

Groundwater Inflow

Shore segments and lake sub-areas

There are ten lake sub-areas and shore segments through which groundwater seeps into the lake: I₁ to I₁₀. I₁ through I₆ and I₉ and I₁₀ represent shore segments through which groundwater flows nearly horizontally into the lake. I₇ and I₈ are areas of the lake affected by deeper upward groundwater seepage that originates mostly from the west and, to a lesser extent, from the northwest and southwest areas.

Areas of seepage

Wherever groundwater movement is mostly horizontal, a cross-sectional area (A) through which the aquifer-lake exchange takes place was determined. First, the length of each shore segment was measured on Figure 28 with a ruler. If the water table well(s) used to calculate the horizontal hydraulic gradient was (were) far from the shoreline (greater than 100 feet), the length of the shore segment was measured

Figure 28 : Map showing the lake sub-areas and shore segments used to calculate the groundwater budget for Lake Wingra, for August and for the period July through September, 1981. I_1 through I_{10} are the shore segments or lake sub-areas through which groundwater seeps into the lake. O_1 through O_5 are the shore segments or lake sub-areas through which groundwater seeps out of the lake.

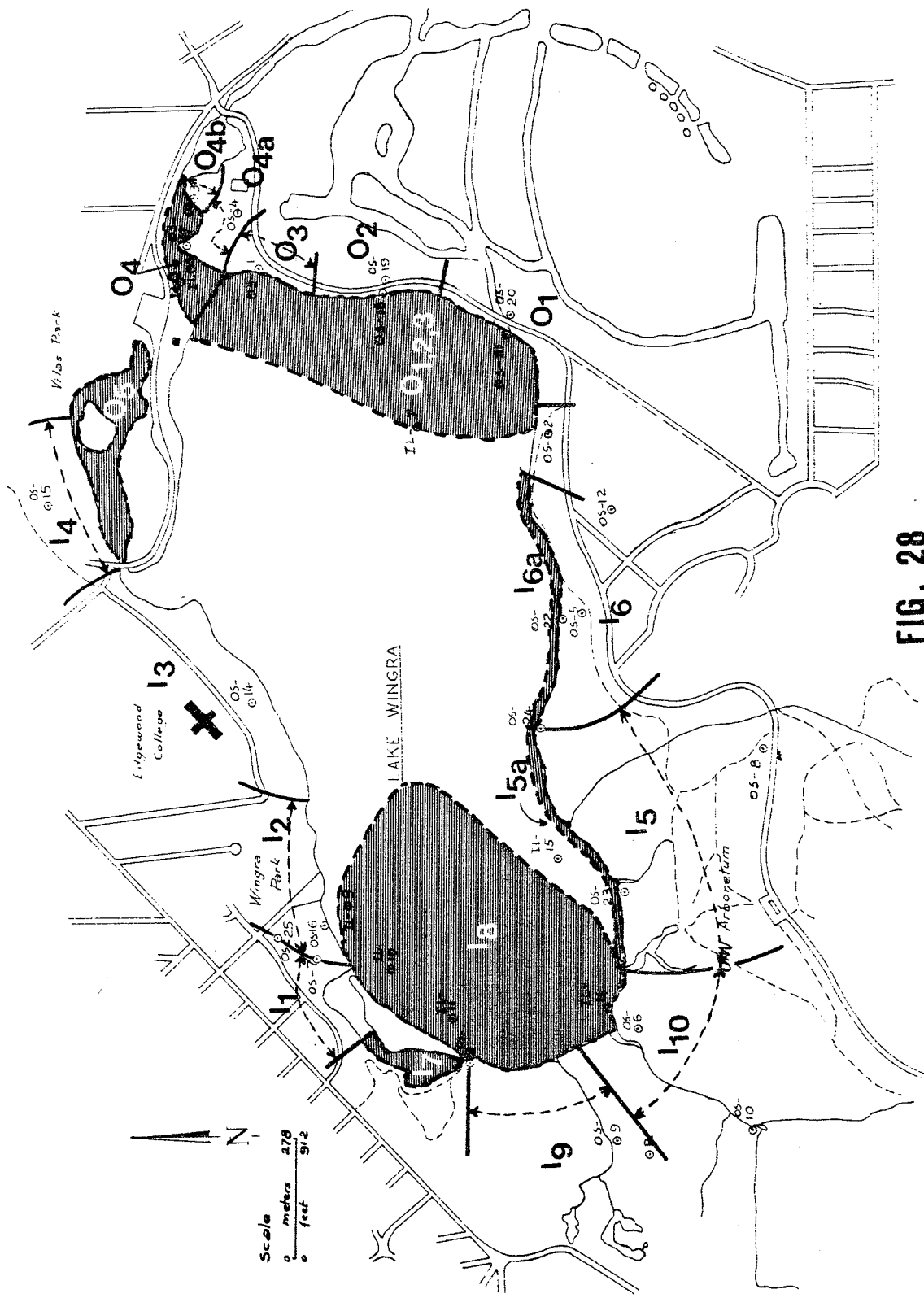


FIG. 28

along a line parallel to the shoreline and located half-way in between the well(s) and the lake (see the dashed arrow lines on Fig. 28). This technique was previously used by Rinaldo-Lee (1978), among others and, assumes a linear water table configuration. In the case of Lake Wingra, where most of the water table slopes almost linearly toward or away from the lake, this approach is reasonable. If the distance that separates the water table well from the lake is short (less than 100 feet), the shoreline itself, measured with a ruler on Figure 28, was used as the segment length.

When the length of the shore segment was defined, the depth of interaction between the lake and the groundwater system was estimated from cross-sections A-A', B-B', C-C' and D-D' (Fig. 13, 14, 15 and 16 in Chapter IV), and/or trends in the vertical hydraulic gradients. In many cases where a lateral groundwater flow dominates (as in most of the north and south shores - see Chapter IV), this depth was determined to equal roughly twice the average lake depth or pond depth or about 20 and 5 feet, respectively. This "two-lake-depth" rule is routinely used in many groundwater budget calculations for lakes (e.g., Rinaldo-Lee, 1978, and Hennings, 1978), and appears to hold in the absence of vertical gradients and lake sediments (Munter, 1979); except for a relatively thin layer of loose organic debris in Lake Wingra, the material under the lake is essentially similar to material composing the shores, and so no real lake sediments exist. The cross-sectional area affected by groundwater exchange is then the product of the length of the shore segment and 20 feet, for I_3 and I_6 , and the product of the segment

length and 5 feet for I_1 and I_4 .

Segments I_2 and I_5 are more difficult to analyze. In these segments, the vertical hydraulic gradient increase with depth (e.g., at OS-23 and at OS-16, see Table 2) and cross section C-C' (Fig. 15) suggest that a deeper groundwater flow from the northwest and southwest shore percolates upward into the lake, some distance away from shore, within sub-area I_8 (see also the discussion in Chapter IV). Therefore in order to separate the shallow groundwater flow from the deep groundwater flow which is already counted as part of the seepage in I_8 , it was assumed that the depth of interaction between the lake and the shallow groundwater system in shore segments I_2 and I_5 was about one average lake depth (estimated from cross section C-C' for July, August and September - see Fig. 15 for September). In I_2 and I_5 , the cross-sectional area through which groundwater flows from the aquifer to the lake is therefore obtained by multiplying the length of the shore segment by 10 feet.

In the west marsh, the groundwater movement is mostly upward until it reaches the upper couple of feet of disturbed sediments (e.g., root zone), when it becomes lateral and flows toward the lake. The groundwater movement in segments I_9 and I_{10} represents this thin lakeward flow that is sometimes increased by infiltrating storm runoff. Field observations suggested that the average depth of this shallow flow system is around 2 feet. The cross-sectional area for this shallow flow system is then equal to the length of the shore segments multiplied by 2 feet. The areas through which groundwater seeps upward

into the lake (I_7, I_8) were determined from a map by means of a planimeter.

Hydraulic gradients

The horizontal hydraulic gradients were calculated by subtracting the lake surface elevation from the elevation of the water table in a shallow onshore piezometer, and dividing this head difference (Δh) by the horizontal distance (Δl) which separates the piezometer from the shoreline. This distance was either determined in the field with a steel tape (short distances) or measured on a map, along flow lines (estimated from Figures 11 and 25) with a ruler graduated in millimeters. The vertical hydraulic gradients, on the other hand, were obtained directly by dividing the head difference between a shallow and a deep piezometer in a nest, by the vertical distance which separates the intakes of the piezometers. Since the groundwater budget was to be determined both for August and for the three-month period (July, August and September), the average groundwater water levels and lake stage for August and the three-month period were used, except where mentioned otherwise.

The horizontal hydraulic gradients in shore segments I_2, I_3 and I_4 were computed using either the only piezometer present or the piezometer farthest from the lake. In I_1, I_5, I_6 and I_{10} the horizontal gradient was determined from respectively, OS-25Sh, OS-8Sh, OS-5Sh and OS-10Sh, and the lake. In I_9 , the representative gradient was determined by

averaging the horizontal gradient between OS-9Sh and the lake, and the gradient between the water table well at the recorder site and the lake.

The vertical hydraulic gradient in I_7 was obtained by averaging the gradients recorded at OS-11Sh and MP-10a. Because MP-10a only began operating in the second half of August (and through September), only the data gathered in the second half of August entered the August groundwater budget. Likewise, only the vertical gradients observed in MP-10a in the last one and a half months of the three-month period (July, August and September) were used to calculate the three-month period average groundwater budget. The fact that the gradient at MP-10a is known for only one half of each period of interest should not introduce a significant error, since the vertical gradients observed in June at two experimental mini-piezometer stations installed in the vicinity of MP-10, were similar to those recorded in August and September.

The choice of representative vertical hydraulic gradient for sub-area I_8 met with more difficulties. First, because the piezometers IL-12 and IL-13 were destroyed by the rifting ice, no data existed for the central portion of I_8 during the summer. However, because of arguments made earlier showing the similarity between the winter and summer vertical hydraulic gradients in the western portion of the lake (see Table 3), it was decided to use the average gradients recorded in February 1981, for those two piezometers. Furthermore, the deep piezometer at IL-14 did not reach equilibrium until mid-August (after

the alcohol added to keep the water from freezing in winter was pumped out in early July) and so the gradients that typified the second half of August and the second half of the three-month period were assumed to represent the average gradients for August and for the three month period, respectively. The vertical gradient assigned for I_3 was the result of averaging all the gradients recorded or assumed for all the in-lake piezometers within the lake sub-area and for the onshore piezometers at OS-11 (see Fig 28).

Hydraulic conductivities

The choice of values for the hydraulic conductivity is based on the results from piezometer tests, a water level recession curve analysis and previously published data, all of which are summarized and discussed in Appendix G.

If the ratio of the vertical (K_v) to the horizontal hydraulic conductivity (K_h) is allowed to range from 1 (isotropic) to 1/1000 (greatly anisotropic) in the sediments beneath the lake, the K_h obtained for the lake sediments in this study, is about five to fifteen times greater than the K_v obtained by Oakes et al. (1975). Furthermore, the computer model of cross section B-B' showed that the three sedimentary layers around and below Lake Wingra were also best represented with an average anisotropy of 1:10 ($K_v:K_h$ - see Chapter V). For these two reasons all the sediments in the Lake Wingra area were assumed to have an anisotropy of 1:10. For the Edgewood College area, which is

dominated by bedrock (see Chapter I), this assumption may not be valid; however, for reasons discussed below the maximum error that could enter into the groundwater inflow calculations for this area, is minor.

Assuming a 1:10 anisotropy, all hydraulic conductivity values of interest yielded by slug and bail tests appear in column 3 of Table G-1.

From piezometer tests conducted at IL-9, MP-9a and MP-9b, the average K_h for the lake sediments was found to be about 6.8×10^{-6} ft/s. With the assumed anisotropy, K_v becomes 6.8×10^{-7} ft/s, close to the value reported by Oakes et al. (1975) which is 7.6×10^{-7} ft/s. For Ho-Nee-Um pond (I7), K_v was assumed to be 3.3×10^{-6} ft/s, or about 1/10 of the average value obtained for K_h at MP-10a (the reproducibility of the slug test at MP-10 was particularly good). From visual inspection of Table G-1 and a couple of simple arithmetic operations, it can be seen that the slug and bail tests performed on the north, south and east shore, between the lake and Gardner Marsh, generated very similar K_h values averaging 5.7×10^{-5} ft/s. This value was then assigned to all north and south shore sediments in the groundwater discharge area. Finally, the K_h in the intensively weathered upper portion of the west marsh was taken to be 2.1×10^{-4} ft/s, the value obtained from a water level recession curve analysis (see Appendix G). The reader is referred to Tables 8 or 9 for the details on the distribution of the K-values for all the shore segments and lake sub-areas.

Groundwater Outflow

Shore segments and lake sub-areas

The portion of the lake affected by groundwater outflow was divided into five segments and sub-areas (see Fig. 28). O_1 through O_3 represent the eastern shore, or the Arboretum drive, which separates the lake from Gardner Marsh. In these out-seepage segments groundwater flow is nearly horizontal. O_4 is the lake area closest to the USGS spilling dam where a significant down seepage occurs. O_5 occupies all of Vilas Pond. In O_4 and O_5 , most of the groundwater recharge to the aquifer is assumed to be vertical (see Fig. 14 and the discussion in Chapter IV).

Areas of seepage

The depth of major interaction between the lake and the aquifer in O_1 , O_2 and O_3 was estimated from computer modeling (see Chapter V) and field observations, to be about 25 feet. The cross-sectional areas through which exchange between the lake and the aquifer occurs are found by multiplying 25 feet by the length of each shore segment. The horizontal areas affected by vertical seepage in O_4 and O_5 were determined with a planimeter.

Hydraulic gradients

The hydraulic gradients characteristic of the groundwater outflow areas were obtained in much the same manner as those computed for the groundwater discharge portions of the lake and shore. The lake elevation and the water elevation in the shallow piezometers located on the east side of the Arboretum drive determined the horizontal hydraulic gradient in O_1 and O_2 . However, it should be noted that this is essentially the horizontal gradient value at the top of the aquifer. At a depth of roughly 25 feet the horizontal gradient appears to become negligible. Therefore, the horizontal gradient was vertically averaged over the first 25 feet below the water table. The average of the vertical hydraulic gradients recorded in the deep piezometers at IL-2 and IL-3 produced the vertical gradient used in O_4 ; for the August groundwater budget the average of all the gradients recorded during the month was computed. However, because the vertical gradient was not known for the first half of July, the average gradient used in the three-month groundwater budget was determined from all gradients recorded during a period extending from the second half of July through September. Finally, the average vertical gradient recorded at OS-15Sh was assumed to be representative of the gradient that existed in O_5 (Vilas Pond).

Hydraulic conductivities

Following the arguments developed earlier in this section, the horizontal hydraulic conductivity chosen for O_1 , O_2 and O_3 was 5.7×10^{-5} ft/s. In sub-area O_4 , the vertical hydraulic conductivity for the lake (6.8×10^{-7} ft/s) was assumed, whereas the K_v -value determined for I_7 (3.3×10^{-6} ft/s) was assigned to Vilas Pond (O_5).

Results and Discussion

The results show that the groundwater budget calculated for August is very similar to the one computed for the three-month period (see Tables 8 and 9). This is expected in view of the general overall stability of the groundwater system described in Chapter IV. In both cases, for August and for the three-month period groundwater seeps into the lake at an average rate of +.27 and .28 ft^3/s , respectively. Groundwater out-seepage varies somewhat more; in August groundwater outseepage averaged $-.075 \text{ ft}^3/\text{s}$, whereas the mean groundwater outflow for the three-month period was about $-.089 \text{ ft}^3/\text{s}$. These results indicate that roughly three to three and a half times more groundwater seeps into the lake than exits it. More than 80% of the groundwater inflow to the lake comes as deep flow from the western portion of the area. About 15% of the groundwater seeps into the lake from the south (7.5%) and the north (7.5%) central hills.

Several major assumptions that could lead to potential errors in

Table 8 : Parameters and gradients used to calculate the groundwater budget for August 1981, and results.

SHORE SEGMENT OR LAKE SUB-AREA	HYDRAULIC GRADIENT (ft/ft)	AREA PER- PENDICULAR TO THE FLOW LINES (ft ²)	HYDRAULIC CONDUCTIVITY, K (ft/s)	VOLUMETRIC FLOW RATE (ft ³ /s)
I ₁	+ 1.83x10 ⁻³	4,650	5.7x10 ⁻⁵	+ .00049
I ₂	+ 2.31x10 ⁻³	12,500	5.7x10 ⁻⁵	+ .00165
I ₃	+ 6.84x10 ⁻³	51,700	5.7x10 ⁻⁵	+ .02016
I ₄	+ 7.43x10 ⁻⁴	6,500	5.7x10 ⁻⁵	+ .00028
I ₅	+ 3.57x10 ⁻³	22,000	5.7x10 ⁻⁵	+ .00448
I ₆	+ 9.82x10 ⁻³	40,200	5.7x10 ⁻⁵	+ .02250
I ₇	+ 2.06x10 ^{-1†}	117,710	3.3x10 ^{-6*}	+ .08002
I ₈	+ 6.56x10 ^{-2†}	3,173,620	6.8x10 ^{-7*}	+ .14157
I ₉	+ 2.47x10 ⁻³	2,360	2.1x10 ⁻⁴	+ .00122
I ₁₀	+ 2.08x10 ⁻³	2,880	2.1x10 ⁻⁴	+ .00126
TOTAL GROUNDWATER INFLOW :				+ .27363 or about .27
O ₁	- 1.99x10 ⁻²	31,500	5.7x10 ⁻⁵	- .03573
O ₂	- 1.63x10 ⁻²	25,125	5.7x10 ⁻⁵	- .02334
O ₃	- 4.84x10 ⁻³	17,000	5.7x10 ⁻⁵	- .00469
O ₄	- 1.64x10 ^{-2†}	219,270	6.8x10 ^{-7*}	- .00245
O ₅	- 6.70x10 ^{-3†}	378,527	3.3x10 ^{-6*}	- .00837
TOTAL GROUNDWATER OUTFLOW :				- .07458 or about - .075

† represents vertical gradients. All other values are horizontal gradients.

* represents K_v. All other values are K_h.

Table 9 : Parameters and gradients used to calculate the groundwater budget for the three-month period extending from July through September 1981, and results.

SHORE SEGMENT OR LAKE SUB-AREA	HYDRAULIC GRADIENT (ft/ft)	AREA PER- PENDICULAR TO THE FLOW LINES (ft ²)	HYDRAULIC CONDUCTIVITY, K (ft/s)	VOLUMETRIC FLOW RATE (ft ³ /s)
I ₁	+ 2.26x10 ⁻³	4,650	5.7x10 ⁻⁵	+ .00060
I ₂	+ 2.86x10 ⁻³	12,500	5.7x10 ⁻⁵	+ .00204
I ₃	+ 7.43x10 ⁻³	51,700	5.7x10 ⁻⁵	+ .02190
I ₄	+ 1.53x10 ⁻³	6,500	5.7x10 ⁻⁵	+ .00057
I ₅	+ 3.72x10 ⁻³	22,000	5.7x10 ⁻⁵	+ .00466
I ₆	+ 1.04x10 ⁻²	40,200	5.7x10 ⁻⁵	+ .02383
I ₇	+ 2.02x10 ^{-1†}	117,710	3.3x10 ^{-6*}	+ .07847
I ₈	+ 6.53x10 ^{-2†}	3,173,620	6.8x10 ^{-7*}	+ .14092
I ₉	+ 2.50x10 ⁻³	2,360	2.1x10 ⁻⁴	+ .00124
I ₁₀	+ 2.14x10 ⁻³	2,880	2.1x10 ⁻⁴	+ .00129
TOTAL GROUNDWATER INFLOW :				+ .27552 or about + .28
O ₁	-1.90x10 ⁻²	31,500	5.7x10 ⁻⁵	- .03411
O ₂	-1.60x10 ⁻²	25,125	5.7x10 ⁻⁵	- .02291
O ₃	-4.66x10 ⁻³	17,000	5.7x10 ⁻⁵	- .00452
O ₄	-1.58x10 ^{-2†}	219,270	6.8x10 ^{-7*}	- .00236
O ₅	-1.99x10 ^{-2†}	378,527	3.3x10 ^{-7*}	- .02486
TOTAL GROUNDWATER OUTFLOW :				- .08876 or about : - .089

† represents vertical gradients. All other values are horizontal gradients.

* represents K_v . All other values are K_h .

the groundwater budget were made to arrive at these results. These are discussed below.

1. The horizontal hydraulic conductivity (K_h) of all the sediments around and below the lake was assumed to be ten times greater than the vertical hydraulic conductivity (K_v). Since K_h and K_v calculated from the results of slug/bail tests depend on the assumed anisotropy of the sediments (Hvorslev, 1951), an inaccurate anisotropy value could generate incorrect hydraulic conductivity values. However, Table G-1 in Appendix G shows that if the anisotropy varies from 1:1 to 1:1000, K_h only doubles whereas K_v remains essentially unchanged. An anisotropy greater than 1:100 is unlikely in most aquifer systems, and so the errors in K_h that could arise from a wrong choice of anisotropy are probably minor especially because reasons given earlier argue for an anisotropy in the neighborhood of 1:10.

2. Another source of error could arise from the assumed hydraulic gradients. In the case of sub-areas, the chosen vertical gradients are probably fairly representative of the areas, in most cases. Possible exceptions are O_4 , I_{5a} and I_{6a} where the vertical gradients are probably underestimated. In O_4 , the gradient near the dam is not known, but is expected to be four to six times greater than that observed at IL-2. This is based on the two feet drop in hydraulic head at the dam and on vertical gradients observed in the lakeward side along the Arboretum drive. In the total groundwater budget however, this possible underestimation of the vertical gradient leads to only a minor error, as shown when the groundwater seepage out of O_4 is recalculated using horizontal gradients and an onshore approach. Most of groundwater out-

seepage from O_4 occurs along its southern shore and through its bottom, as suggested by cross section B-B' (see Fig. 14 in Chapter IV), and so the southern shore was subdivided into two shore segments, O_{4a} and O_{4b} (see Fig. 28). The depth of interaction between the lake and the aquifer was assumed to be 15 feet, based on the computer model of cross section B-B' (see Chapter V) and on field data indicating the presence of a low hydraulic conductivity layer at a depth of roughly 15 feet below the water table. The lengths of the shore segments O_{4a} and O_{4b} were determined from the dashed arrow line shown in Figure 28. K_h was taken to be 5.7×10^{-5} ft/s. The total average groundwater seepage out of O_{4a} and O_{4b} was about .0026 ft³/s for August and .0027 ft³/s for the three-month period. This compares rather well with the outseepage obtained by using in-lake piezometers (.0025 ft³/s for August and .0024 ft³/s for the three-month period).

August groundwater inflow rates from I_5 and I_6 were re-computed using data obtained from the mini-piezometers, assuming that most of the seepage into the adjacent lake sub-areas I_{5a} and I_{6a} (see Fig. 28) was vertical. The in-seepage rates obtained were .00035 and .00336 ft³/s for I_{5a} and I_{6a} respectively, or about 1/10 of the rates found with the onshore approach for I_5 and I_6 . Several reasons may lead to this underestimation of the inflow rate; a. There may be an important horizontal component to the groundwater flow in the narrow lake band adjacent to the southern shoreline. b. The actual average vertical gradient in the lake band may be greater than that recorded by the mini-piezometers if there is an increase in vertical gradient with depth

below the sediment-water interface. c. The vertical gradient obtained from the mini-piezometers may not have been representative for the narrow band. It is believed that there is an important horizontal groundwater component of the flow in this area, based on the generally lateral nature of the flow in the southern shore, and so the first is probably the major cause that led to the underestimation of the groundwater inflow using the mini-piezometer data.

In O₅, no in pond gradients were recorded, but the assumption that the vertical gradient in the pond is similar to that found at OS-15 is probably reasonable, in view of the proximity of OS-15 to the pond and the homogeneous sediments in the area.

In some cases, the horizontal hydraulic gradient chosen to represent shore segments may not be very representative of the entire segment; this is particularly true in I₁, I₂ and I₉. These three shore segments are fortunately minor contributors to the groundwater inflow into the lake, and would not significantly change the groundwater inflow even if the gradients were either doubled or halved.

3. the third major assumption and the one most difficult to verify is the depth of interaction between the lake and the aquifer system. This problem is circumvented by using an in-lake piezometers approach in the western portion of the study area, where more than 80% of the groundwater inflow to the lake takes place.

In the east, the depth of interaction between the lake and aquifer system is better known as it is determined and verified by computer modeling, cross sections (e.g., Figures 13 and 14) and field

observations. The confidence is even further reinforced when groundwater outflow is computed independently using in-lake piezometers and mini-piezometers (IL-4, IL-5, IL-6, IL-7, OS-21 and MP-7a). Using a K_v of 6.8×10^{-7} ft/s and average winter levels for all in-lake piezometers, and August levels for OS-21 and MP-7a, groundwater outseepage from the lake sub-area $O_{1,2,3}$ (see Fig. 28) amounts to .085 ft^3/s . This is slightly higher than the total groundwater outflow calculated from segments O_1 , O_2 and O_3 combined (.064 ft^3/s for August and .062 ft^3/s for the three-month period), but if the vertical gradient in the eastern portion of the lake was indeed greater in winter than during the summer (see Chapter 4), then the discrepancy between the results obtained from the two techniques would be significantly reduced. In addition, the lake bottom area ($O_{1,2,3}$) affected by groundwater outflow may be slightly overestimated.

In view of all the arguments presented above, the author believes that the computed groundwater budgets for the lake for August and for the three-month period, are fairly accurate. A comparison between groundwater seepage rates calculated in this study with rates calculated earlier by Oakes et al. (1975) and Novitzki and Holmstrom (1979), reveals that groundwater inflow and outflow rates obtained here are significantly lower than the rates determined by these earlier researchers. Oakes et al. (1975) estimated that the average groundwater inflow for 1972 was .52 ft^3/s or roughly twice the average in-seepage rate computed in this study for August or the three-month period. They obtained their result assuming a groundwater inflow area (260,000 m^2)

roughly comparable to the one assumed in this study ($320,000 \text{ m}^2$), but a hydraulic conductivity ($7.6 \times 10^{-7} \text{ ft/s}$) about five times smaller than the weighted average hydraulic conductivity ($3.6 \times 10^{-6} \text{ ft/s}$) used here. Nevertheless, the discrepancy between the estimates of groundwater inflow arises mostly from the much greater hydraulic gradient assumed by Oakes et al. (1975) ($.24 \text{ m/m}$) relative to the average gradient used here ($.03 \text{ m/m}$).

Novitzki and Holmstrom (1979) did not measure groundwater inflow directly, but they assumed that groundwater varied with spring discharge. Assuming that groundwater inflow rate was about 42% of the average spring discharge for the period 1972 to 1977, they estimated groundwater inflow rates of $.96 \text{ ft}^3/\text{s}$ for August and $.94 \text{ ft}^3/\text{s}$ for the three-month period. These rates are about 3.5 times superior to those calculated in this study.

If spring discharge is added to the groundwater seepage, the total groundwater inflow rate found in this study is $3.25 \text{ ft}^3/\text{s}$ ($2.98 \text{ ft}^3/\text{s} + .27 \text{ ft}^3/\text{s}$) for August and $3.26 \text{ ft}^3/\text{s}$ ($2.98 \text{ ft}^3/\text{s} + .28 \text{ ft}^3/\text{s}$) for the three-month period. Novitzki and Holmstrom (1979) found that the August total groundwater inflow rates for a six year period (1972-1977) ranged from $1.79 \text{ ft}^3/\text{s}$ to $2.57 \text{ ft}^3/\text{s}$, and that the total groundwater inflow rates for the three-month period of July through September varied from $1.74 \text{ ft}^3/\text{s}$ to $2.60 \text{ ft}^3/\text{s}$ (see Table 10). The combined spring discharge and groundwater in-seepage found in this study is therefore only about 87% more than the lowest and 25% more than the highest combined spring and groundwater in-seepage rates reported by Novitzki and Holmstrom

Table 10 : Comparison between the groundwater seepage rates in and out of Lake Wingra obtained in this study and in two previous studies.

	STUDY	YEAR(S)	AUGUST FLOW RATE	COMBINED JULY, AUGUST AND SEPTEMBER FLOW RATE	ONE YEAR AVERAGE FLOW RATE
Combined spring discharge and groundwater in-seepage rate (ft ³ /s)	Oakes et al. (1975)	1972			+ 2.38
	Novitzki and Holmstrom (1979)	HIGH : 1975	+ 2.57	+ 2.60	
		LOW : 1972	+ 1.79	+ 1.74	
		AVERAGE : 1972-77	+ 2.28	+ 2.24	
	This study	1981	+ 3.25	+ 3.26	
Groundwater out-seepage rate (ft ³ /s)	Oakes et al. (1975)	1972			- 0.66
	Novitzki and Holmstrom (1979)	HIGH : 1975	- 0.38	- 0.38	
		LOW : 1972	- 0.26	- 0.27	
		AVERAGE : 1972-77	- 0.32	- 0.32	
	This study	1981	- 0.075	- 0.089	

(1979) for August and the three-month period.

Groundwater outflow rates found in this study disagree with those reported by Novitzki and Holmstrom (1979), and even more so with those published by Oakes et al. (1975); the outseepage rates estimated by Oakes et al. (1975) for 1972 ($-.66 \text{ ft}^3/\text{s}$) is roughly between seven and a half and nine times greater than the groundwater outflow rates computed here for the three-month period ($-.089 \text{ ft}^3/\text{s}$) and for August ($-.075 \text{ ft}^3/\text{s}$), respectively. Novitzki and Holmstrom (1979) reported groundwater outflow rates varying from $.26$ to $.38 \text{ ft}^3/\text{s}$ for August, or about three and a half to five times greater than those found in this study for August, and outseepage rates for the three-month period ranging between $.27$ and $.38 \text{ ft}^3/\text{s}$, or about three to four and a half times higher than the groundwater outflow rates calculated here for the same period. The discrepancies seen in the reported outflow rates result from the nature of the technique used and from the assumptions made by each researcher. For the details of the methods used by Oakes et al. (1975) and by Novitzki and Holmstrom (1979), the reader is referred to Chapter I and to the reports by Oakes et al. (1975) and Novitzki and Holmstrom (1979).

Oakes et al. (1975) and Novitzki and Holmstrom (1979) were forced to make questionable assumptions and use approximating techniques because of the paucity of field data. In light of the new field data collected for this study, the author feels that the results presented here are more accurate than those published earlier, even though several assumptions (discussed above) had to be made. These assumptions, even

if erroneous, would however not affect significantly the results obtained in this study. The major finding about the groundwater budget for Lake Wingra, is that both groundwater inflow and outflow to the lake are much less than previously thought. The total groundwater inflow (spring discharge + groundwater in-seepage), on the other hand, is roughly similar to that estimated in earlier studies. However, the ratio of groundwater in-seepage to spring discharge obtained for August in this study (.09) is much less than the ratio assumed for every month of the study by Novitzki and Holmstrom (1979) (.42), or estimated by Oakes et al. (1975) (.28) for 1972.

Conclusion

A detailed groundwater budget was determined for Lake Wingra for August, 1981 and for the period July through September, 1981, using in-lake and onshore piezometers and Darcy's Law. Groundwater seeps into the lake, mostly from the west, at rates of about .27 and .28 ft³/s during August and the three-month period, respectively, and flows out of the lake at rates of roughly .075 ft³/s for August, and .089 ft³/s for the period July through September. Compared with previous estimates of groundwater in- and out-seepage rates for Lake Wingra, the new in- and out-seepage rates are significantly lower. If however the spring discharge is added to the groundwater in-seepage rate, then the total

groundwater inflow calculated here roughly compares with previous total groundwater inflow estimates.

VII. CONCLUSIONS AND RECOMMENDATIONS

Conclusions were given at the ends of the last three chapters and will be summarized below. The limitations of this particular study will also be discussed and possible improvements that could benefit future studies will be suggested.

The first three conclusions pertain specifically to Lake Wingra.

1. Groundwater circulation and overall stability of the system:

Groundwater seeps into Lake Wingra from the north, west and south shores, and seeps out of the lake from the east shore. The general groundwater movement is nearly horizontal in the north, south and in the east under the Arboretum drive, but is mostly vertical in the east and west marshes and in the adjacent portions of the lake. The overall configuration of the groundwater circulation in the Lake Wingra area is relatively stable, at least in the absence of large scale pumping.

2. Fluctuations of the groundwater levels:

Short term fluctuations of the groundwater levels are controlled by precipitation, whereas long term fluctuations may be induced by pumping of the sandstone aquifer.

3. Groundwater budget:

Calculations of the groundwater budget for Lake Wingra revealed that most groundwater in-seepage occurs from the west (more than 80%), but a significant groundwater in-seepage component (15%) comes from the

north and south central hills. Most of the out-seepage occurs through the Arboretum drive. The rates of groundwater in- and out- seepage calculated in this study are significantly lower than previously published rates (Oakes et al., 1975, and Novitzki and Holmstrom, 1979). The new rates, however, are believed to be more accurate.

The remaining conclusions are more general and include recommendations for similar future studies.

4. Piezometer placement:

The horizontal and vertical distribution of the piezometer points is crucial in a lake study. Horizontally, piezometers should be present both near and far away from the lake, in order to define adequately the groundwater movement and the average horizontal gradient. For Lake Wingra, there was a lack of water table piezometers far from shore (500-1000 feet), especially in the north and east shores. The vertical gradients are even more important. Vertical gradients around a shallow lake often vary with depth; a nest consisting of only two piezometers is inadequate to define this variation. At least three and preferably more piezometers should be emplaced in some of the strategic piezometer nests around the lake. This is especially important to determine the depth of interaction between the lake and the aquifer. In addition, nests should have deep piezometers (greater than two lake depths), not only to aid in determining the depth of interaction between the aquifer and the lake, but also to provide the needed field data to calibrate a three-dimensional groundwater flow model. This study lacked these deep piezometers.

Along the same lines, in-lake or mini-piezometers should be used generously. Piezometer nests in the lakebed itself are also essential. In-lake piezometers or mini-piezometers should be installed parallel to groundwater flow lines, at different distances from shore, and until the water level in the piezometers is at essentially the same elevation as the lake surface. In-lake piezometers are much more easily installed in winter, when the lake is covered by ice. Alternatively a flat bottom barge could be used in the summer in order to provide the solid substrate needed for the emplacement of piezometers in the lakebed. The unavailability of a flat bottom barge combined with drifting ice late in the winter, seriously reduced and limited the in-lake piezometer network in this study.

5. Pressure transducers and seepage meters:

Pressure transducers, if available, probably are valuable tools for groundwater circulation and groundwater budget studies, especially in shallow mucky bottom lakes. On the other hand, seepage meters are not recommended for mucky bottom, highly eutrophic lakes.

6. Groundwater flow models:

Groundwater flow computer models are particularly attractive to compute groundwater budgets for lakes (Munter, 1979, Munter and Anderson, 1981), but care should be exercised in determining how and when to use these models. For example, an areal two-dimensional computer model for Lake Wingra is inappropriate, because of the importance of the vertical groundwater component in the eastern and western areas. Two-dimensional cross section models are possible, but

many are needed to obtain a reasonable average groundwater budget for the entire lake, because of the diverse hydrogeologic conditions. Three-dimensional computer models are the best choice, but reliable field data at many points in the system, including information to define the boundaries of the three-dimensional grid, are needed to allow calibration of the model. Although there was a lot of field data available for the Lake Wingra area, the data distribution was uneven, and, for example, there was essentially no data available below a depth of 30 feet beneath the ground surface.

Even if there are not enough data to calibrate the model for the purpose of computing the groundwater component of the lake's water budget, groundwater flow models can still provide useful information. For example, a simple two-dimensional cross section model can help estimate the average hydraulic conductivity or average anisotropy of sedimentary layers. Furthermore, as was shown in Chapter V, a cross section groundwater flow model can help locate geologic contacts and aid in determining the depth of interaction between the aquifer and the lake.

Groundwater flow computer models are also valuable design tools. For example, a model could be developed at the onset of a study, so as to determine where information is needed, and thus where piezometers should be emplaced. This probably would have been beneficial for the Lake Wingra study, and would be beneficial for many similar studies.

7. The Darcy's law approach:

As an alternative to the use of a three-dimensional flow model, a

standard Darcy's law approach using results from both in-lake and on-shore piezometers (see Chapter VI) provides satisfactory results for a groundwater circulation and budget study of a shallow lake around and below which the groundwater circulation is complex and has in some places a significant vertical component.

8. Chemical speciation models:

Other techniques that could be useful in determining areas of groundwater seepage into a lake are chemical speciation models, such as WATEQ (Truesdell and Jones, 1974) or PHREEQE (Parkhurst et al., 1980). If the lake water is chemically different from the groundwater (as it often is), then samples of pore water taken from near the water-sediment interface could reveal whether groundwater flows out of the lake (the chemical composition of the pore water is essentially the same as the lake water's) or if groundwater flows into the lake (the chemical composition of the pore water resembles that of the groundwater). Computer speciation models could easily compute saturation indices for various chemical species in many samples of pore water, lake water and groundwater. The main obstacle in using this technique is to gather reliable samples and make certain that biological activity has not significantly altered the chemistry of the pore water.

In the Lake Wingra study, preliminary WATEQ runs based on one set of chemical data obtained from Oakes et al. (1975), showed that the lake water is undersaturated with respect to CaCO_3 and $\text{CaMg}(\text{CO}_3)_2$, whereas the groundwater is saturated with both. If the problem of getting reliable pore water samples can be overcome, Lake Wingra would be a

potential candidate for using the chemical technique. Additional efforts to study Lake Wingra could certainly focus on gathering good chemical samples of the lake, pore and ground waters. Presently, there is a lack of chemical data for the Lake Wingra area; limited chemical data can be found in the reports of Oakes et al. (1975) and Kluesener (1972), among others.

9. Temperature probes:

Temperature probes could also be useful to study groundwater in-seepage to a lake. Indeed, the temperature difference between the groundwater and the lake water could be exploited so as to delineate zones of groundwater inflow. The best period to try this technique is early in the spring as soon as the ice breaks out, when the groundwater should be warmer than the lake water, or in mid-summer when the groundwater is colder than the lake water.

Beside efforts to gather chemical data for the entire Lake Wingra area, more research is needed in the northern portion of the lake in order to delineate more precisely the extent of the groundwater seepage zone, although this probably would not significantly affect the groundwater budget calculated in this study. In addition, groundwater budgets should be computed for many different days (or months), so that a relationship between the groundwater budget and the antecedent precipitation index can be established for Lake Wingra. This would allow estimations of the daily (or monthly) groundwater budget from the

value of the index (see Chapter IV). Finally, the long term fluctuations of the groundwater potentials should be monitored to determine whether the water table in the Lake Wingra area is declining as feared.

REFERENCES CITED

- Barwell, V.K. and D.R. Lee. 1981. Determination of horizontal-to-vertical hydraulic conductivity from seepage measurement on lakebeds. *WRR*, 17 (3), pp. 565-570.
- Baumann, P.C., J.F. Kitchell, J.J. Magnuson, and T.B. Kayes. 1974. Lake Wingra, 1837-1973: a case history of human impact. *Transactions of the Wisconsin Academy of Sciences, Arts and Letters*, Vol. LXII, pp. 57-94.
- Born, S.M., S.A. Smith, and D.A. Stephenson. 1974. The hydrogeologic regime of glacial-terrain lakes, with management and planning applications. *Inland Lake Renewal and Management Demonstration Project*. p. 73.
- Bradbury, K.R. 1982. Hydrologic relationships between Green Bay and onshore aquifers in Door County, Wisconsin. Ph.D. Thesis, Univ. Wi., Madison. Not completed yet.
- Brown, R.H. 1963. Ground-water movement in a rectangular aquifer bounded by four canals. U.S.G.S. Water-Supply Paper 1545c, pp. c86-c100.
- Erickson, D.R. 1981. A study of littoral groundwater seepage at Williams Lake, Minnesota using seepage meters and wells. M.S. Thesis, Univ. of Minnesota. p. 135.
- Freeze, R.A. and P.A. Witherspoon. 1967. Theoretical analysis of regional groundwater flow. 2. Effect of water-table configuration and subsurface permeability variation. *WRR*, 3 (2), pp. 623-634.
- Hasler, A.D. 1975. Coupling of land and water systems. edit by A.D. Hasler, Springer-Verlag, Inc. p. 309.
- Hennings, R.G. 1978. The hydrogeology of a sand plain seepage lake, Portage County, Wisconsin. M.S. Thesis, Univ. Wi., Madison. p. 70.
- Huff, D.D. and H.L. Young. 1980. The effect of a marsh on runoff: I. A water-budget model. *J. Envir. Qual.*, 9 (4), pp. 633-640.
- Hvorslev, M.J. 1951. Time lag and soil permeability in ground-water observations. U.S. Army Corps Engrs. Waterways Exp. Station. Bull. 36, Vicksburg, Miss. p. 50.

- Ingersoll, L.R. and O.J. ZoBel. 1948. Heat conduction. New York, McGraw-Hill Book Co.
- Kluesener, J.W. 1972. Nutrient transport and transformation in Lake Wingra, Wisconsin. Univ. of Wisc., Eastern Deciduous Forest Biome Memo, Rept. 72-42. p. 180.
- Lee, D.R. 1977. A device for measuring seepage flux in lakes and estuaries. *Limnol. Oceanogr.*, 22 (1), pp. 140-147.
- Lee, D.R. and J.A. Cherry. 1978. A field exercise on groundwater flow using seepage meters and mini-piezometers. *J. Geol. Education*, 27, pp. 6-10.
- Linsley, R.K., M.A. Kohler, and J.L.H. Paulhus. 1975. Hydrology for engineers, 2nd ed. McGraw-Hill, Inc. p. 482.
- McBride, M.S. and H.O. Pfannkuch. 1975. The distribution of seepage within lakebeds. *J. Research, U.S.G.S.*, 3 (5), pp. 505-512.
- McLeod, R.S. 1975a. A digital-computer model for estimating drawdowns in the sandstone aquifer in Dane County, Wisconsin. *Wis. Geol. and Nat. Hist. Surv. Inf. Circ.* 28. p. 91.
- McLeod, R.S. 1975b. A digital-computer model for estimating hydrologic changes in the aquifer system in Dane County, Wisconsin. *Wisc. Geol. and Nat. Hist. Surv. Inf. Circ.* 30. p. 40.
- Meyboom, P. 1966. Unsteady groundwater flow near a willow ring in hummocky moraine. *J. Hydrology*, 4, pp. 38-62.
- Meyboom, P. 1967. Mass-transfer studies to determine the groundwater regime of permanent lakes in hummocky moraine of western Canada. *J. Hydrology*, 5, pp. 117-142.
- Munter, J.A. 1979. Groundwater modeling of three lake/aquifer systems in Wisconsin. M.S. Thesis, Univ. Wi., Madison. p. 100.
- Munter, J.A. and M.P. Anderson. 1981. The use of ground-water flow models for estimating lake seepage rates. *GW*, 19 (6), pp. 608-616.
- Novitzki, R.P. and B.K. Holmstrom. 1979. Monthly and annual water budgets of Lake Wingra, Madison, Wisconsin, 1972-77. U.S.G.S. Water-Res. Investig., 79-100, Madison, Wi. p. 31

- Oakes, E.L., G.E. Hendrickson, and E.E. Zuehls. 1975. Hydrology of the Lake Wingra basin, Dane County, Wisconsin. U.S.G.S. Water-Res. Investig., 17-75, Madison, Wi. p. 31.
- Parkhurst, D.L., D.C. Thorstenson, and L.N. Plummer. 1980. PHREEQE-- A computer program for geochemical calculations, U.S.G.S. Water-Res. Investig., 80-96. p. 210.
- Remson, I., S.M. Gorelick, and J.F. Fliegner. 1980. Computer models in ground-water exploration. GW, 18 (5), pp. 447-451.
- Rinaldo-Lee, M.B. 1978. Hydrogeology and computer model of the Bass Lake area, St. Croix County, Wisconsin. M.S. Thesis, Univ. Wi., Madison. p. 95.
- Sale, R.D. and R. Helgeland. 1961. Hydrographic map of Madison area lakes, Published by Wisc. Geol. and Nat. Hist. Surv., Univ. Wi.
- Theis, C.V. 1935. The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using ground-water storage. Am. Geophys. Union Trans., pt. 2, pp. 519-524.
- Trescott, P.C., G.F. Pinder, and S.P. Larson. 1976. Finite-difference model for aquifer simulation in two dimensions with results of numerical experiments. TWRI of the U.S.G.S., Book 7, Chapter C1. p. 116.
- Truesdell, A.H. and B.F. Jones. 1974. WATEQ, a computer program for calculating chemical equilibria of natural waters. J. Research, U.S.G.S., 2 (2), pp. 233-248.
- United States Department of the Interior, Bureau of Reclamation. 1967. Water measurement manual. 2nd ed., Denver, Colorado, pp. 162-164.
- University of Waterloo field notes, 1981. Field methods in contaminant hydrogeology. Department of Earth Sciences, Univ. Waterloo, Waterloo-Ontario, unpublished.
- Viessman, W. Jr., J.W. Knapp, G.L. Lewis, and T.E. Harbaugh. 1977. Introduction to hydrology, 2nd ed., IEP-Dun-Donnelley Harper and Row, publishers, Inc.
- Weeks, E.P. 1964. Use of water-level recession curves to determine the hydraulic properties of glacial outwash in Portage county, Wisconsin. U.S.G.S. Prof. Paper, 501-B, pp. B181-B184.

- Winter, T.C. 1978. Numerical simulation of steady state three-dimensional groundwater flow near lakes. WRR, 14 (2), pp. 245-254.
- Winter, T.C. 1981. Uncertainties in estimating the water balance of lakes. Wat. Res. Bul., 17(1), pp 82-115.

APPENDIX A :

Construction Details for the Piezometers and Water Table Wells Composing
the Main Groundwater Monitoring Network in and around Lake Wingra.

EXPLANATION OF THE SYMBOLS

Sh = Shallow piezometer

I = Intermediate piezometer

D = Deep piezometer

EC = Galvanized steel electrical conduit

J = Jetting with a 140 GPM Homelite jetting pump

HA = 3" diameter hand auger

P = Pipe pounder

SD = Mobil minute man drill with 2 3/4" diameter augers

D = Truck mounted drill

TABLE A-1: CONSTRUCTION DETAILS OF THE ONSHORE PIEZOMETERS AND WATER
TABLE WELLS

STATION NUMBER	WELL	PIPE ID DIAMETER (INCHES)/MATERIAL	SCREEN* ID DIAMETER (INCHES) X LENGTH (FEET) MATERIAL ¹	COLLAR ELEVATION (FEET)
1	Sh	3/4 / EC	3/4 X 1.0, PVC	100.38
	D	3/4 / EC	-	100.06
2	Sh	1 1/4 / PVC	1 1/4 X 1.2, PVC	101.25
	I	3/4 / EC	-	100.77
	D	3/4 / EC	-	99.62
3	D	1 1/4 / Steel	1 1/4 X 1.0, PVC	101.68
	Sh	1 1/4 / PVC	1 1/4 X 1.0, PVC	101.80
	D	1 1/4 / PVC	1 1/4 X 1.0, PVC	-
4	SH	3/4 / EC	3/4 X 1.0, PVC	100.00
	SH-I	1 1/4 / PVC	1 1/4 X 1.0, PVC	101.75
	D-I	3/4 / EC	3/4 X 1.0, PVC	101.77
	D	1 1/4 / PVC	1 1/4 X 1.0, PVC	103.12
5	Sh	1 1/4 / PVC	1 1/4 X 1.0, PVC	104.56
	D	1 1/4 / Steel	1 1/4 X 1.5, PVC	104.55
6	Sh	1 1/4 / PVC	1 1/4 X 1.0, PVC	101.10
	I	1 1/4 / PVC	1 1/4 X 1.0, PVC	101.75
	D	1 1/4 / PVC	1 1/4 X 1.0, PVC	102.26
7	Sh	1 1/4 / PVC	1 1/4 X 1.0, PVC	102.66
	D	1 1/4 / PVC	1 1/4 X 1.0, PVC	102.37
8	Sh	3/4 / EC	3/4 X 1.0, PVC	111.42
	I	3/4 / EC	3/4 X 1.0, PVC	111.15
	D	3/4 / EC	3/4 X 1.0, PVC	110.63
9	Sh	1 1/4 / Steel	1 1/4 X 1.5, Steel	103.17
	D	1 1/4 / Steel	1 1/4 X 1.5, Steel	103.44
10	Sh	1 1/4 / PVC	1 1/4 X 1.0, PVC	108.33
	D	1 1/4 / Steel	1 1/4 X 2.0, Steel	110.31
11	Sh	3/4 / EC	3/4 X 0.5, PVC	100.98
	D	3/4 / EC	3/4 X 0.5, PVC	101.87
12	Sh	3/4 / EC	3/4 X 1.0, PVC	104.70
	D	3/4 / EC	3/4 X 1.0, PVC	107.86
14	Sh	3/4 / EC	3/4 X 1.0, PVC	102.76
	Sh	3/4 / EC	3/4 X 1.0, PVC	108.06
15	I	3/4 / EC	3/4 X 1.0, PVC	106.02
	D	3/4 / EC	3/4 X 1.0, PVC	106.63
	Sh	3/4 / EC	3/4 X 1.0, PVC	102.11
16	I	3/4 / EC	-	102.69
	D	1 1/4 / Steel	1 1/4 X 2.2, Steel	104.90
	Sh	3/4 / EC	3/4 X 1.0, PVC	101.67
17	D	3/4 / EC	-	101.47
	Sh	3/4 / EC	3/4 X 1.0, PVC	100.45
18	D	3/4 / EC	-	99.75
	Sh	3/4 / EC	3/4 X 1.0, PVC	99.12
19	D	3/4 / EC	3/4 X 1.0, PVC	99.32
	Sh	3/4 / EC	3/4 X 1.0, PVC	97.80
20	D	3/4 / EC	-	99.10

STATION NUMBER	WELL	PIPE ID DIAMETER (INCHES)/MATERIAL	SCREEN* ID DIAMETER (INCHES) X LENGTH (FEET) MATERIAL ¹	COLLAR ELEVATION (FEET)
21	Sh	3/4 / EC	3/4 X 1.0, PVC	100.01
	D	3/4 / EC	3/4 X 1.0, PVC	100.39
22	Sh	3/4 / EC	3/4 X 1.0, PVC	101.33
	D	3/4 / EC	3/4 X 1.0, PVC	102.72
23	Sh	3/4 / EC	-	99.90
	I	3/4 / EC	-	100.14
	D	3/4 / EC	-	100.48
24	Sh	3/4 / EC	-	99.76
	I	3/4 / EC	-	100.20
	D	3/4 / EC	-	102.01
25	Sh	3/4 / EC	-	107.70
	D	1 1/4 / PVC	3/4 X 1.0, PVC	106.84
	RECORDER WATER TABLE WELL	6/Steel	-	103.59

*Slot size = .008 inch

TABLE A-1: CONSTRUCTION DETAILS OF THE ONSHORE PIEZOMETERS AND WATER TABLE WELLS.

STATION NUMBER	WELL	TOTAL LENGTH OF PIEZOMETER OR PIEZOMETER PLUS POINT/APPROX. ELEVATION (FEET) ABOVE GROUND SURFACE	MODE OF EMPLACEMENT	REMARKS
1	Sh	6.0/0.5	J	-
	D	20.0/0.2	J	-
2	Sh	9.2/2.0	HA	-
	I	20.0/1.60	J	-
	D	30.0/0.5	J	-
	D	31.0/2.3	SD + P	Plugged
3	Sh	6.3/0.55	SD	-
	D	12.9/1.9	SD	Destroyed
4	Sh	6.1/.25	HA	Was pushed down less than 0.1 foot some-time during the study
	Sh-I	12.5/2.10	SD	-
5	D-I	21.0/0.5	HA + P	Partially plugged
	D	31.0/3.3	SD	-
	Sh	7.1/2.45	SD	-
	D	20.75/2.25	SD + P	-
6	Sh	6.32/1.75	SD	-
	I	20.65/2.40	SD	-
	D	35.50/2.90	SD	-
7	Sh	7.17/1.2	SD	-
	D	24.1/0.9	SD	Plugged
8	Sh	6.0/1.1	SD	Most of the time a dry well
	I	11.0/0.75	HA	-
	D	26.0/0.35	SD	Partially plugged since mid-June in 1981, perhaps earlier
9	Sh	12.3/1.45	HA + ?	Emplaced by the USGS
	D	70.8/1.85	HA + ?	Emplaced by the USGS
10	Sh	9.65/3.85	HA	-
	D	61.0/6.0	?	Emplaced by the USGS
11	Sh	5.50/0.45	P	-
	D	21.70/1.35	HA + P	-
12	Sh	6.1/0.85	HA	-
	D	16.1/4.0	HA + P	Plugged
14	Sh	6.0/2.05	SD	-

STATION NUMBER	WELL	TOTAL LENGTH OF PIEZOMETER OR PIEZOMETER PLUS POINT/APPROX. ELEVATION (FEET) ABOVE GROUND SURFACE	MODE OF EMPLACEMENT	REMARKS
15	Sh	11.0/3.40	SD + P	-
	I	16.0/1.40	SD + P	Plugged
	D	26.0/1.95	SD + P	-
16	Sh	6.0/.30	J	-
	I	20.0/.80	J	-
	D	53.0/2.85	D	-
17	Sh	6.0/1.20	J	-
	D	20.0/0.95	J	-
18	Sh	6.0/1.20	J	Pushed down and corrected several times
	D	20.0/0.50	J	-
19	Sh	6.0/2.3	HA	-
	D	21.0/1.5	HA + P	Partially plugged
20	Sh	6.0/1.15	J	-
	D	20.0/2.50	J	-
21	Sh	6.0/1.65	J	-
	D	21.0/1.30	J	Plugged
22	Sh	6.0/1.9	J	-
	D	21.0/3.1	J	-
23	Sh	5.0/.35	J	-
	I	20.0/.50	J	-
	D	40.0/.90	J	-
24	Sh	5.40/.60	J	-
	I	15.0/.90	J	-
	D	30.0/2.65	J	-
25	Sh	10.0/2.4	HA + P	-
	D	37.7/1.3	D	-
RECORDER WATER TABLE WELL		6.5/~.5	HA	Emplaced by the USGS

TABLE A-1: CONSTRUCTION DETAILS OF THE ONSHORE PIEZOMETERS AND WATER TABLE WELLS.

TABLE A-2: Construction details of the in-lake piezometers. See front page of Table A-1 for the explanation of the symbols used.

STATION #	PIEZOMETER IF MORE THAN ONE PER STATION	(FEET) TOTAL PIEZOMETER LENGTH (PIPE & POINT)	(FEET) DEPTH INTO LAKE BOTTOM SED- IMENTS (CORRECTED FOR THE POINT)*	(FEET) WATER DEPTH AT THE STATION	REMARKS
1	-	14.6	8.9	4.3	Broken in January. Water level in the well seemed to be slightly lower than lake level
2	Sh D	11.0 21.1	5.5 15.0	4.2 4.2	
3	Sh D	11.2 21.1	4.5 15.0	2.8 2.8	
4	-	21.6	12.1	7.8	Destroyed by the drifting ice
5	-	11.0	5.8	3.7	Destroyed by the drifting ice
6	-	12.6	6.6	4.7	Destroyed by the drifting ice
7	-	21.0	11.0	8.6	Partially plugged during the summer
8	-	21.1	10.4	9.3	Destroyed by the drifting ice
9	-	21.0	14.1	5.5	
10	-	21.1	12.3	7.6	
11	-	21.0	13.5	6.2	
12	-	24.2	11.7	11.2	Destroyed by the drifting ice

STATION #	PIEZOMETER IF MORE THAN ONE PER STATION	(FEET) TOTAL PIEZOMETER LENGTH (PIPE & POINT)	(FEET) DEPTH INTO LAKE BOTTOM SED- IMENTS (CORRECTED FOR THE POINT)*	(FEET) WATER DEPTH AT THE STATION	REMARKS
13	-	25.4	15.7	8.3	Destroyed by the drifting ice
14	Sh D	6.1 21.2	3.3 17.3	~1.5 1.5	
15	-	21.0	10.5	9.0	

*Length of the piezometer pipe located below the water-sediment interface + 1/2 the length of the sand point.

TABLE A-2: CONSTRUCTION DETAILS OF THE IN-LAKE PIEZOMETERS.

APPENDIX B :

Empirical Determination of the Relationship between the Conductivity and the Concentration of A.C.S. (Crystals) NaCl Solutions (NaCl in Deionized Water).

- Reagent tested : A.C.S. (crystals) NaCl

- Conductivity meter used : 1) a YSI model 33 for salt concentrations of less than 40.00 gm/L
- 2) a Beckman RB3 solu-bridge with a VH20 conductivity cell for salt concentrations greater than 40.00 gm/L

Temperature of the experiment : 19°C

Results : The relationship between the conductivity and the concentration of the salt solutions is summarized in Figure B-1. It is basically linear until about 40.00 gm/l, as indicated by the relatively high correlation coefficient, r (.9999), that characterizes the relation.

Note : Figure B-1 was only used for the increased specific conductance above background level due to the sole injection of the salt tracer, and so background conductivity in the field was assumed to be zero for each spring.

SPECIFIC CONDUCTANCE

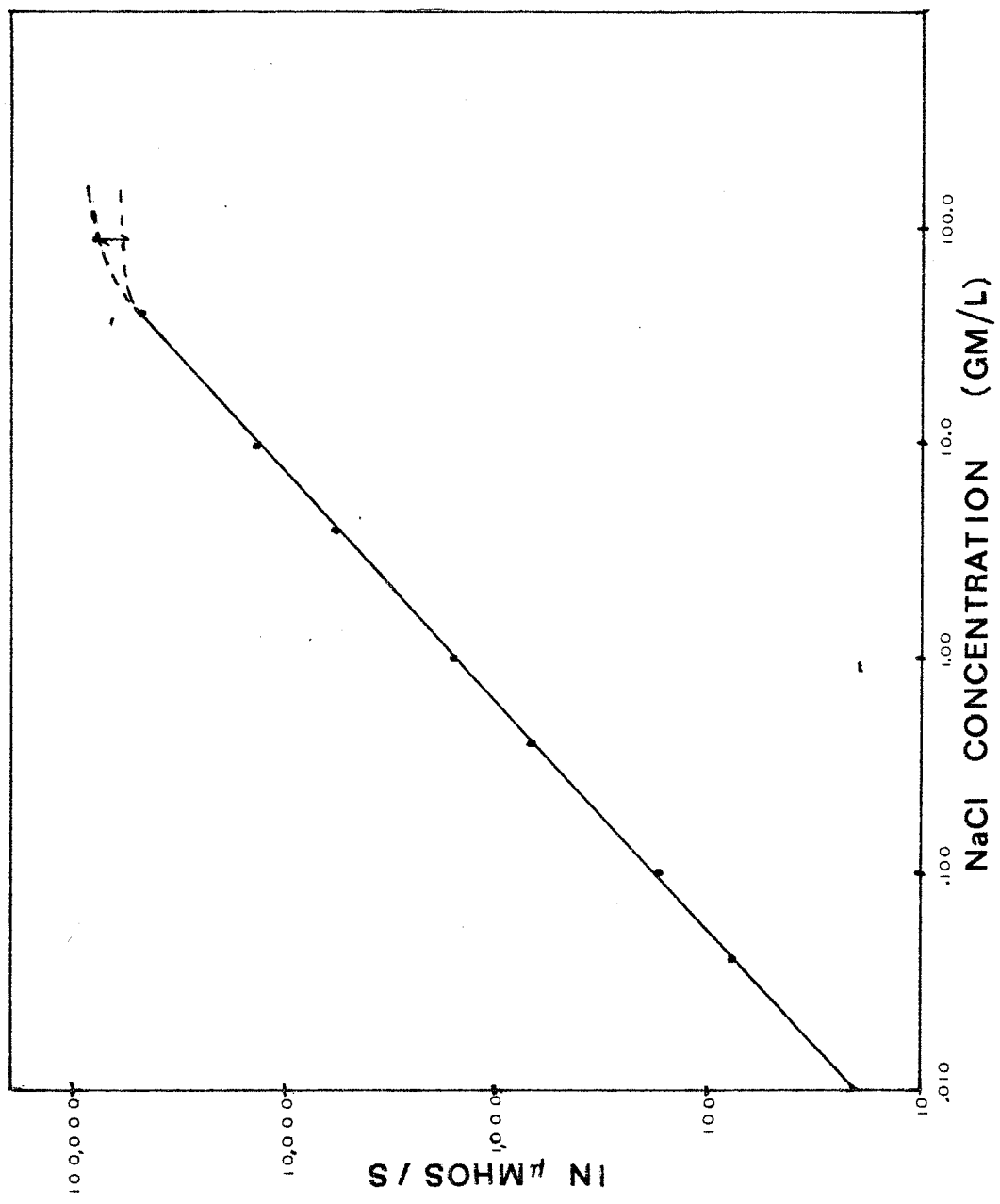


Fig. B-1 : Empirically determined relation between concentration and conductivity for the NaCl solution.

APPENDIX C : Expansion, Evaporation and Diffusion Columns.

Evaluation of the Effects of: 1. Temperature on the Alcohol Column, 2. of the Evaporation of Alcohol from In-Lake Piezometers and 3. of its Diffusion through the Water Column, and their Significance on the Calculated Water Levels, if Neglected in the Correction Applied for the Difference in Density between Water and Rubbing Alcohol.

A. Expansion column experiment (2/3/81 to 3/26/81).

(i) Purpose : To determine the significance of temperature on the alcohol column in the piezometers.

(ii) Material and method : A solution of 70% rubbing alcohol was poured in a $1\frac{1}{4}$ inch diameter transparent plastic pipe (same diameter as the inlake piezometers). Next a small thermometer was placed inside the pipe which was sealed, and then placed vertically and attached to a mobile support.

(iii) Location of the experiment : Outside on the fourth floor balcony of Weeks Hall, at the UW-Madison.

(iv) Results : Figure C-1 shows the thermal expansion curve obtained for a .74 pint (.012 cubic foot) of 70% solution rubbing alcohol contained in a $1\frac{1}{4}$ inch diameter pipe. The slope of the curve is .025 cm/°C. For one pint of alcohol the slope becomes .033 cm/°C (.025 cm/°C x 1/.74), which means that for each °C added or subtracted, the height of the water column changes by .033 cm or .001 foot.

(v) Discussion of the results : The experiment confirms that for the range of temperatures affecting in-lake piezometers (0° to 5°C), the height of the alcohol column in these piezometers does not vary significantly : indeed, for a 5°C increase in temperature, the alcohol column expands by only .0054 foot in height (.033 cm/°C x 5°C), well under the .01 foot accuracy of the water level measurement.

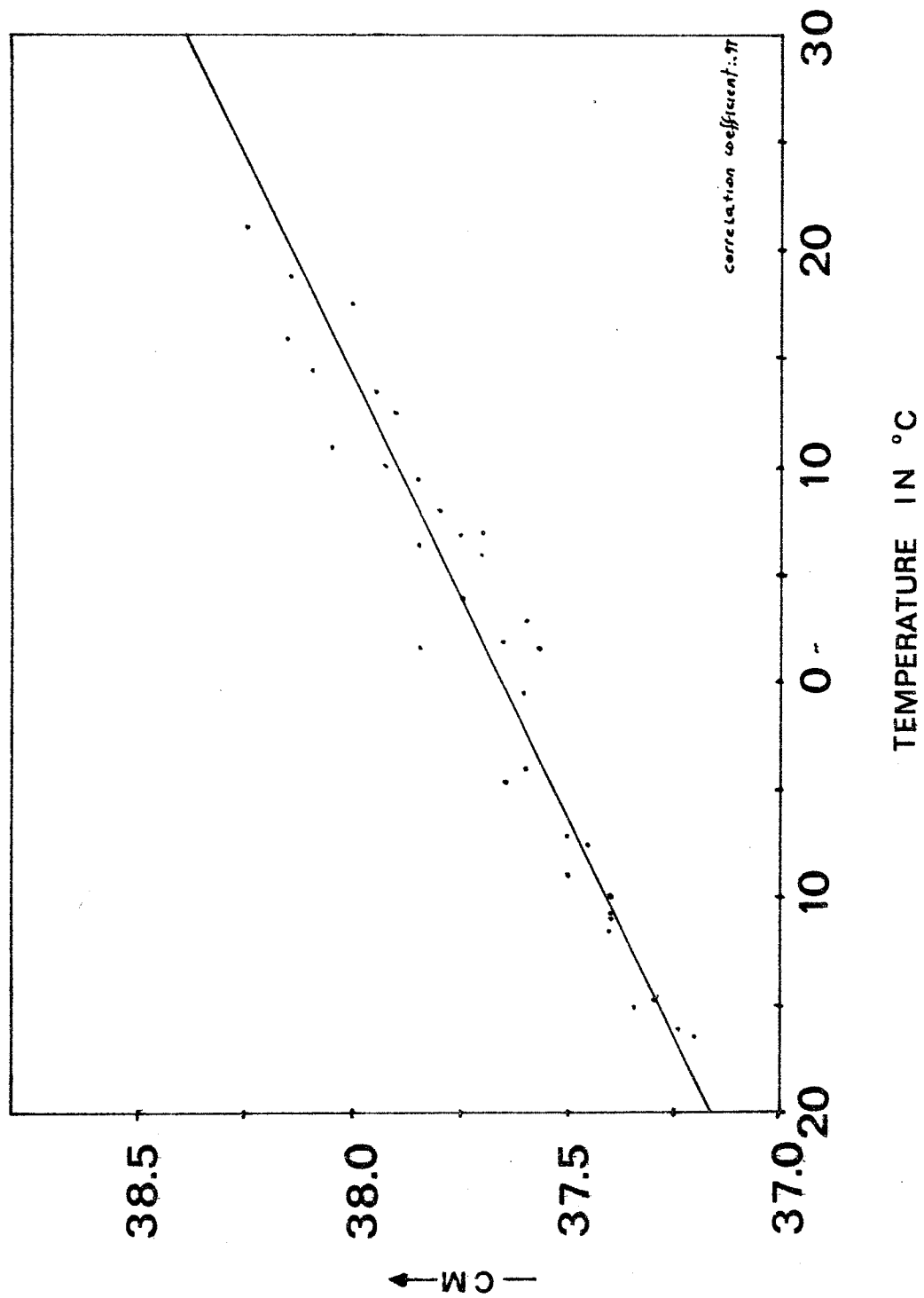


Fig. C-I : Thermal expansion curve for 0.74 pint of rubbing alcohol (see text for the details).

B. Evaporation column experiment (from 1/29/81 to 3/26/81).

(i) Purpose: To determine whether the evaporation of alcohol through the small air vent located in the piezometer cap could be neglected.

(ii) Material and method: One pint of 70% rubbing alcohol was poured over a small amount of water in a 1 $\frac{1}{4}$ inch diameter transparent plastic pipe. A PVC glue-on male treaded adaptor and a pierced cap closed the top of the pipe, as is the case for the in-lake piezometers. The tube was placed vertically and attached to a mobile support.

(iii) Location of the experiment: Outside on the fourth floor balcony of Weeks Halls, at the UW-Madison.

(iv) Results: See Table C-1.

(v) Discussion of the results: Some of the small discrepancies visible in the progression with time of the alcohol levels corrected to 0°C, could be explained by two reasons: first, the assumption that the temperature in the all-alcohol expansion column may not have been exactly the same as in the evaporation column with a water-alcohol mixture. Secondly, the readings were taken by two different people, which certainly could introduce a significant source of error. On the whole, however there is a gradual lowering of the alcohol level, which suggests that the effects of evaporation are noticeable. Nevertheless, from the time when the alcohol was poured into the piezometers (1/9/81) to the time when the last water level reading was made (2/21/81), less than 1 cm of alcohol did evaporate. Assuming that 1 cm of alcohol evaporation took place by February 21st, the error that would result from only applying a straight density correction based on the initial

TABLE C-1: Results of the evaporation column experiment (see text).

DATE OF MEASUREMENT	MEASURED ALCOHOL HEIGHT IN THE COLUMN (CM)	TEMPERATURE OF ALCOHOL ($^{\circ}\text{C}$) ¹	ALCOHOL HEIGHT AT 0°C (CM) ²
01/29	59.90	+20.0	59.24
02/03	58.35	-15.0	58.84
02/04	58.20	-16.5	58.74
02/05	58.15	-16.0	58.68
02/06	58.60	-4.0	58.68
02/08	58.40	-9.0	58.70
02/09	58.30	-10.9	58.66
02/11	58.25	-11.5	58.63
02/12	58.25	-11.5	58.63
02/13	58.70	+0.5	58.70
02/14	58.80	+10.5	58.45
02/15	58.80	+8.0	58.54
02/16	59.20	+18.9	58.58
02/17	59.00	+12.5	58.59
02/18	59.00	+12.5	58.59
02/20	58.65	+7.0	58.42
02/21	58.65	+8.0	58.39
02/23	58.35	+2.0	58.28
02/25	58.15	-0.5	58.15
02/26	58.15	0.0	58.15
03/01	58.15	+3.0	58.05
03/04	58.35	+9.5	58.04
03/07	57.65	-4.5	57.80
03/10	58.05	+6.5	57.84
03/14	58.15	+16.0	57.62
03/26	57.20	+6.0	57.00

1.- The temperature was obtained from the thermometer in the expansion column.

2.- The alcohol height is corrected to 0°C using the temperature correction factor ($.033 \text{ cm}/^{\circ}\text{C}$) obtained in the expansion column experiment.

alcohol volume added in the piezometer, would be about .12 cm (.0039 foot). In other words, the calculated water levels would be .0039 foot lower than what they really are, clearly insignificant in this case.

C. Diffusion.

A diffusion experiment was set up in the laboratory by simply letting dried alcohol float over a column of water and observe what happened through a short time. The diffusion was judged insignificant at least in a laboratory setting. The presence of a considerable quantity of alcohol retrieved from deep in-lake piezometers in early July (one month after the caps were broken off from the wells) seems to reinforce this belief. Finally, a diffusion column set up in the laboratory confirms the validity of this assumption for deep piezometers, but perhaps not for the shallow wells 11 feet or less in total length.

Diffusion column experiment (1/23/81 to 3/1/81).

(i) Purpose: To determine the significance of diffusion of alcohol through the water column.

(ii) Material and method: A solution of 70 % rubbing alcohol was poured into one side of a vertically standing open ended U-shape tube partially filled with water. The initial difference of the fluid level in both sides of the tube was recorded and monitored through time.

(iii) Location: In the hydrogeology laboratory at the UW-Madison.

(iv) Results: See Table C-2.

TABLE C-2 : Results of the diffusion column experiment (see text).

DATE OF MEASUREMENT	TIME OF MEASUREMENT	HEIGHT OF THE ALCOHOL AND WATER COLUMN (CM)	HEIGHT OF THE WATER COLUMN (CM)
01/23	12:15 pm	70.20	67.30
	06:15 pm	70.15	67.25
01/24	09:10 pm	69.90	67.25
01/25	05:30 pm	69.15	66.55
01/26	07:00 pm	68.67	66.10
01/27	07:00 pm	68.20	65.70
01/28	07:30 pm	67.85	65.35
01/29	01:25 pm	67.60	65.15
02/02	06:15 pm	66.30	64.00
02/03	05:30 pm	66.05	63.75
02/04	06:20 pm	65.75	63.50
02/06	07:20 pm	65.25	62.95
02/08	06:00 pm	64.70	62.50
02/12	12:00 pm	63.75	61.20
02/16	01:35 pm	62.70	60.70
02/20	04:45 pm	61.75	59.70
02/25	09:00 pm	60.60	58.70
03/01	02:30 pm	59.65	57.90

(v) Discussion of the results: The evaporation is high in both sides of the U-shape tube since both are open-ended. However, only the relative difference in fluid height between the two sides is important. Within a month's time this difference in height decreased from about 3 cm (1/23/81) to about 2 cm (2/20/81). In other words, about 1/6 of the alcohol had migrated to the other side, or at least 2 feet, which is the approximate length of the column. This migration alone seems too slow to affect the calculated water levels in the in-lake piezometers. However, it could be greatly enhanced as rapid mixing occurs in the upper water column when the alcohol is poured from the piezometer collar onto the water standing 1 or 2 feet below. If the piezometer is not very long (10 feet or so), enough alcohol could diffuse out through the screen, and the water levels corrected for the original volume of added alcohol would result in being lower than they should be. This may explain the discrepancy observed in most shallow wells between the water levels obtained when the alcohol was in the well and the levels measured before the alcohol was poured in and after it was pumped out. Table D-1 shows that most shallow piezometers have calculated February water levels (corrected for density difference between alcohol and water using the initial volume of alcohol added to piezometers) which stand .05 to .08 foot below the lake level, whereas earlier and later measurements indicate that they normally are within .01 to .02 foot of the lake surface.

In this brief discussion, it was assumed that the reduction with time of the difference in the fluid height between the two sides of the

U-shape tube, was solely due to diffusion of the alcohol. One could argue that differential evaporation between alcohol and water could also account for the 1 cm decrease: if this is involved, then the diffusion process could be less important than presented above. Nevertheless, a laboratory experiment whereby two small beakers were filled, one with 50 ml of alcohol and the other with 50 ml of water, suggested that the evaporation rates of both the water and the 70% rubbing alcohol are roughly the same.

APPENDIX D :FIELD DATA.

Tables D-1, D-2A, D-2B	Water level measurements in in-lake and onshore piezometers
Figure D-1 - D-14	Fluctuations in water levels
Table D-3	Precipitation data
Table D-4	Seepage meter data
Table D-5	Water levels in mini-piezometers
Table D-6	Spring discharges

(See the symbol index at the beginning of the thesis for the symbols used in the following tables.)

Table D-1 : Water levels in in-lake piezometers for January and February 1981. Locations of piezometers are shown in Figure 5. See the symbol index at the beginning of the thesis for the symbols used in the following table.

DIFFERENCE (IN FEET) BETWEEN THE WATER LEVEL IN THE PIEZOMETER
AND THE LAKE SURFACE, Δh*

SITE # AND
PIEZOMETER

JANUARY, 1981 (NO ALCOHOL) FEBRUARY, 1981
(CORRECTED FOR ALCOHOL)

REMARKS

	20	21	22	24	25	26	28	8	21	
IL-2:SH					0.00			-.07	-.05	**
:D									-.63	
IL-3:SH					+.01	+.01		-.11	-.05	**
:D								-.17	-.18	
IL-4								-.02	-.05	
IL-5		-.84	-.81	-.79	-.70	-.85	-.84	-.07	-.08	**
IL-6		-.01	-.02		+.01	+.01		-.02	-.05	**
IL-7								-.02	-.05	
IL-8								-.02	-.03	
IL-9								+.06	+.08	
IL-10								+.13	+.10	
IL-11					+.53	+.53		+.45	+.42	
IL-12								+.19	+.17	
IL-13					+.27	+.32		+.32	+.30	
IL-14:SH									-.06	**
:D								+.85	+.81	
IL-15								+.04	+.01	

* - To convert these Δh values to heads, add or subtract to/from the lake levels (Ho-Nee-Um gauge) given in Table D-2B.

** - February water levels may be too low due to alcohol diffusion (see Appendix C).

TABLE D-1: WATER LEVELS IN IN-LAKE PIEZOMETERS FOR JANUARY AND FEBRUARY, 1981.

Table D-2A: Water levels in onshore and in-lake piezometers, lake stage and stage-discharge gauge from May 15 to September 28, 1981 - Main data collection. Locations of piezometers are shown in Figure 5. See the symbol index at the beginning of the thesis for the symbols used in the following table.

TABLE D-2A: WATER LEVELS IN ONSHORE PIEZOMETERS FROM MAY 15 TO
SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	OS-1			OS-2		OS-3
	SH	D	Sh	I	D	Sh
15/5	97.24	97.85	98.59	98.72	98.60	97.62
22/5	96.87	97.71	98.51	98.65	98.67	97.40
30/5	96.70	97.57	98.51	98.59	98.63	97.49
6/6	96.23	97.46	98.16	98.52	98.62	96.93
13/6	96.44	97.59	98.83	98.63	98.64	97.77
15/6	98.61	98.15	-	-	-	-
16/6	98.36	98.37	99.25	98.99	98.86	98.70
18/6	98.07	98.13	99.03	98.95	98.78	98.47
25/6	97.74	97.79	98.90	98.86	98.71	98.20
30/6	97.14	97.74	98.76	98.75	98.66	97.99
7/7	96.58	97.60	98.40	98.58	98.58	97.34
13/7	97.30	97.77	98.93	98.76	98.68	98.28
21/7	97.53	97.90	98.91	98.81	98.70	98.31
28/7	97.46	97.87	98.82	98.72	98.64	98.09
4/8	97.07	97.86	98.70	98.73	98.67	97.92
16/8	97.63	97.94	98.95	98.79	98.68	98.39
21/8	97.14	97.85	98.67	98.69	98.61	97.97
25/8	97.11	97.86	98.56	98.65	98.61	97.97
4/9	98.43	98.30	99.03	98.98	98.78	98.52
11/9	97.71	98.28	98.85	98.93	98.79	98.20
28/9	97.46	97.94	98.82	98.76	98.66	98.14

TABLE AD-2A: WATER LEVELS IN ONSHORE PIEZOMETERS FROM MAY 15 TO SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	OS-4				OS-5	
	Sh	Sh-I	D-I	D	Sh	D
15/5	96.65	96.65	96.21	93.74	101.01	101.22
22/5	96.44	96.45	96.13	93.82	100.90	100.94
30/5	96.24	96.25	96.12	93.94	100.82	100.91
6/6	96.05	96.06	96.01	93.96	100.63	100.83
13/6	95.92	95.94	95.89	94.04	100.84	100.80
15/6	-	-	-	-	101.30	101.63
16/6	97.92	96.98	95.92	94.56	101.25	101.48
18/6	98.26	97.93	96.35	94.62	101.10	101.42
25/6	97.71	97.61	96.58	94.62	101.11	101.28
30/6	97.20	97.16	96.54	94.62	100.99	101.25
7/7	96.80	96.81	96.74	94.57	100.71	100.92
13/7	97.09	96.84	96.57	94.53	101.04	101.29
21/7	97.65	97.46	96.62	94.72	101.00	101.11
28/7	97.58	97.39	96.61	94.74	100.96	100.94
4/8	97.36	97.34	96.71	94.83	100.81	100.90
16/8	98.14	97.70	96.78	95.11	100.95	101.12
21/8	97.93	97.79	96.86	95.07	100.74	100.81
25/8	97.61	97.57	96.92	95.03	100.67	100.75
4/9	98.52	98.33	97.35	95.55	101.12	101.31
11/9	98.01	98.08	97.50	95.55	101.07	101.21
28/9	97.50	97.46	97.21	95.54	101.05	100.95

TABLE D-2A: WATER LEVELS IN ONSHORE PIEZOMETERS FROM MAY 15 TO
SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	OS-6			OS-7		OS-8	
	Sh	I	D	Sh	Sh	D	
15/5	99.20	99.63	101.33	99.22	105.23	105.05	
22/5	99.04	99.52	101.24	98.97	105.06	104.92	
30/5	98.93	99.40	101.23	98.81	104.83	104.91	
6/6	98.53	99.25	101.17	98.54	104.50	104.90	
13/6	98.86	99.19	101.19	98.88	104.44	104.47	
15/6	-	-	-	100.58	107.26	105.26	
16/6	99.22	99.36	101.34	100.73	106.91	105.31	
18/6	99.15	99.34	101.29	100.36	105.35	105.31	
25/6	99.18	99.32	101.28	100.06	105.21	105.08	
30/6	99.14	99.27	101.25	99.61	104.84	105.07	
7/7	98.76	99.22	101.18	98.92	104.37	105.07	
13/7	99.10	00.24	101.22	99.44	104.59	105.05	
21/7	99.14	99.31	101.27	99.87	104.98	105.03	
28/7	99.17	99.36	101.24	99.43	104.64	105.03	
4/8	99.09	99.40	101.24	99.08	104.47	105.00	
16/8	99.14	99.35	101.25	99.88	104.73	104.97	
21/8	99.05	99.33	101.21	99.27	104.36	104.97	
25/8	98.99	99.34	101.19	99.02	104.35	104.95	
4/9	99.20	99.45	101.29	100.33	105.43	104.99	
11/9	99.20	99.51	101.31	99.89	105.14	104.98	
28/9	99.22	99.45	101.27	99.46	104.87	104.96	

TABLE D-2A: WATER LEVELS IN ONSHORE PIEZOMETERS FROM MAY 15 TO SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	OS-9		OS-10		OS-11		OS-12
	Sh	D	Sh	D	Sh	D	Sh
15/5	99.74	109.48	-	-	98.83	104.54	100.46
22/5	99.51	108.93	-	-	98.52	104.13	100.21
30/5	99.45	109.10	-	-	98.53	104.20	99.98
6/6	99.15	108.61	-	-	98.22	103.76	99.76
13/6	99.77	108.88	-	-	99.30	104.12	99.77
15/6	-	-	-	-	-	-	-
16/6	100.63	109.54	-	-	99.52	104.53	101.59
18/6	100.17	109.37	-	-	99.17	104.40	101.23
25/6	100.18	109.38	-	-	99.18	104.53	101.26
30/6	99.99	109.06	-	-	99.10	104.26	100.85
7/7	99.51	108.54	101.10	109.69	98.38	103.78	100.25
13/7	100.45	109.05	101.66	110.33	99.39	104.17	100.68
21/7	100.26	109.14	101.65	110.61	99.27	104.31	101.07
28/7	100.07	109.18	101.39	110.76	99.35	104.33	100.68
4/8	99.78	109.08	101.23	110.53	99.11	104.30	100.35
16/8	100.25	109.24	101.65	110.69	99.24	104.43	100.98
21/8	99.78	108.69	101.17	109.96	98.65	103.90	100.39
25/8	99.74	108.81	101.11	110.27	98.62	103.85	100.18
4/9	100.15	109.40	101.47	111.11	99.23	104.32	101.29
11/9	100.00	109.18	101.36	110.84	98.99	103.90	101.06
28/9	100.07	109.53	101.41	111.24	99.19	104.44	100.97

TABLE D-2A: WATER LEVELS IN ONSHORE PIEZOMETERS FROM MAY 15 TO
SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	OS-14		OS-15		OS-16			OS-17	
	Sh	Sh	D	Sh	I	D	Sh	D	
15/5	99.63	-	-	98.66	98.86	-	97.53	97.74	
22/5	99.44	-	-	98.51	98.69	-	97.26	97.71	
30/5	99.38	-	-	98.47	98.67	-	97.19	97.69	
6/6	99.09	98.35	-	98.33	98.53	100.42	96.85	97.49	
13/6	99.72	99.36	-	98.55	98.79	100.63	97.25	97.73	
15/6	-	100.28	-	99.39	99.64	100.98	-	-	
16/6	100.04	99.80	-	99.50	99.62	100.97	99.10	98.79	
18/6	99.71	99.34	-	99.47	99.41	100.87	98.94	98.86	
25/6	99.82	99.44	98.95	99.27	99.19	101.00	98.20	98.36	
30/6	99.71	99.07	98.83	99.05	99.03	100.78	97.71	98.05	
7/7	99.35	98.53	98.67	98.42	98.68	100.54	97.21	97.88	
13/7	100.13	99.81	98.76	98.70	99.01	100.69	98.03	98.15	
21/7	99.91	99.57	98.91	98.99	99.11	100.81	98.27	98.43	
28/7	100.02	99.38	98.95	98.70	98.97	100.90	98.17	98.41	
4/8	99.73	98.88	98.89	98.50	98.80	100.79	97.83	98.33	
16/8	99.96	99.66	99.08	98.96	99.20	101.01	98.80	98.48	
21/8	99.71	98.99	99.02	98.58	98.89	100.74	98.46	98.56	
25/8	99.74	98.85	98.94	98.51	98.79	100.68	98.11	98.45	
4/9	100.07	100.04	99.44	99.27	99.33	101.19	98.97	98.86	
11/9	100.01	99.78	99.54	99.01	99.11	101.20	98.41	98.79	
28/9	100.04	99.80	99.53	98.70	98.97	101.23	97.96	98.29	

TABLE D-2A: WATER LEVELS IN ONSHORE PIEZOMETERS FROM MAY 15 TO SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	OS-18		OS-19		OS-20		OS-21
	Sh	D	Sh	Sh	D	Sh	
15/5	97.23	97.38	96.33	96.30	-	98.56	
22/5	97.12	97.31	96.21	96.13	-	98.53	
30/5	97.05	97.30	96.18	96.10	-	98.56	
6/6	96.76	97.27	95.82	95.69	98.87	98.51	
13/6	96.97	97.37	96.31	96.33	99.06	98.65	
15/6	-	97.64	97.23	97.49	99.38	-	
16/6	99.27	97.87	97.34	97.46	99.50	99.15	
18/6	98.98	97.73	96.97	97.05	99.24	98.90	
25/6	98.79	97.47	96.53	96.69	99.10	98.74	
30/6	97.77	97.35	96.31	96.46	99.10	98.64	
7/7	96.98	97.23	95.87	95.94	99.12	98.54	
13/7	99.07	97.41	96.51	96.70	99.10	98.77	
21/7	98.86	97.47	96.56	96.70	99.23	98.75	
28/7	98.22	97.37	96.41	96.54	99.09	98.65	
4/8	97.29	97.38	96.22	96.32	99.10	98.60	
16/8	98.87	97.45	96.73	96.82	99.10	98.81	
21/8	97.22	97.37	96.33	96.34	99.25	98.62	
25/8	97.03	97.34	96.26	96.29	99.08	98.57	
4/9	98.84	97.75	97.05	97.16	99.29	98.86	
11/9	97.81	97.66	96.54	96.73	99.11	98.67	
28/9	98.18	97.33	96.43	96.62	99.10	98.62	

TABLE D-2A: WATER LEVELS IN ONSHORE PIEZOMETERS FROM MAY 15 TO SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	OS-22		OS-23		
	Sh	D	Sh	I	D
15/5	99.15	99.51	-	-	-
22/5	99.12	99.45	-	-	-
30/5	99.11	99.45	98.65	98.69	-
6/6	98.92	99.25	98.49	98.59	-
13/6	99.28	99.61	98.81	98.82	-
15/6	99.44	99.90	-	-	-
16/6	99.40	99.86	99.21	99.12	99.37
18/6	99.28	99.69	98.99	99.02	99.31
25/6	99.29	99.68	98.86	98.90	99.18
30/6	99.23	99.59	98.76	98.81	99.31
7/7	98.99	99.33	98.53	98.66	99.42
13/7	99.37	99.74	98.99	98.91	99.41
21/7	99.29	99.67	98.94	98.91	99.49
28/7	99.32	99.68	98.91	98.89	99.57
4/8	99.21	99.54	98.74	98.79	99.62
16/8	99.30	99.69	98.97	98.93	99.65
21/8	99.14	99.47	98.74	98.81	99.59
25/8	99.14	99.46	98.72	98.79	99.53
4/9	99.33	99.74	99.07	99.05	99.56
11/9	99.27	99.66	98.85	98.94	99.66
28/9	99.30	99.68	98.85	98.87	99.67

TABLE D-2A: WATER LEVELS IN ONSHORE AND IN-LAKE PIEZOMETERS FROM
MAY 15 TO SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	OS-24			OS-25		IL-2	
	Sh	I	D	Sh	D	Sh	D
15/5	-	-	-	-	-	-	-
22/5	-	-	-	-	-	-	-
30/5	98.72	98.74	98.70	-	100.50	-	-
6/6	98.66	98.61	98.62	-	100.24	-	-
13/6	98.82	98.82	98.88	-	100.44	-	-
15/6	-	-	-	-	-	-	-
16/6	99.23	99.24	99.26	-	100.85	-	-
18/6	99.00	99.01	99.05	-	100.73	-	-
25/6	98.90	98.91	98.97	-	100.85	-	-
30/6	98.82	98.84	98.90	-	100.65	-	-
7/7	98.69	98.67	98.69	99.75	100.40	98.72	-
13/7	98.92	98.94	99.01	99.65	100.54	98.90	-
21/7	98.91	98.93	99.01	100.26	100.70	98.93	98.43
28/7	98.85	98.88	98.97	99.87	100.75	98.78	98.31
4/8	98.79	98.81	98.87	99.59	100.63	98.78	98.41
16/8	98.97	98.99	99.05	100.11	100.87	98.93	98.49
21/8	98.81	98.82	98.86	99.73	100.62	98.78	98.33
25/8	98.76	98.78	98.82	99.52	100.55	98.77	98.38
4/9	99.00	99.03	99.08	100.78	101.13	99.01	98.60
11/9	98.87	98.90	98.97	100.51	101.12	98.86	98.61
28/9	98.86	98.91	98.97	99.93	101.10	98.78	98.47

TABLE D-2A: WATER LEVELS IN IN-LAKE PIEZOMETERS FROM MAY 15 TO SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	IL-3		IL-7	IL-9	IL-10	IL-11	IL-14	
	Sh	D	D	D	D	D	Sh	D
15/5	-	-	-	98.76	-	-	-	-
22/5	-	-	-	-	-	-	-	-
30/5	-	-	-	98.78	-	-	-	-
6/6	-	-	-	98.76	98.67	-	-	-
13/6	-	-	-	98.84	98.76	-	-	-
15/6	-	-	-	-	-	-	-	-
16/6	-	-	-	99.11	99.19	-	-	-
18/6	-	-	-	99.07	99.01	99.46	-	-
25/6	-	-	-	98.92	98.91	99.53	-	-
30/6	-	-	-	98.86	98.77	99.41	-	-
7/7	98.71	-	-	98.81	98.73	99.41	98.72	-
13/7	98.89	98.77	-	98.99	98.90	99.49	98.92	-
21/7	98.89	98.82	-	98.95	98.89	99.44	98.89	-
28/7	98.79	98.68	-	98.87	98.81	99.33	98.80	-
4/8	98.76	98.75	98.71*	98.88	98.79	99.26	98.77	-
16/8	98.97	98.87	-	99.01	98.99	99.52	98.97	99.55
21/8	98.79	98.70	98.80*	98.86	98.83	99.35	98.80	99.48
25/8	98.75	98.72	98.76*	98.86	98.78	99.32	98.77	99.45
4/9	99.00	98.92	99.05*	99.08	99.05	99.51	99.02	99.73
11/9	98.84	98.93	99.01*	99.01	98.89	99.40	98.86	99.59
28/9	98.80	98.78	98.96*	98.87	98.84	99.31	98.81	99.54

* These water levels are probably erroneous in view of the extremely slow response of the piezometer.

TABLE D-2A: WATER LEVELS IN ONE IN-LAKE PIEZOMETER, LAKE STAGE AND STAGE-DISCHARGE GAUGE FROM MAY 15 TO SEPTEMBER 28, 1981.

DATES OF MEASUREMENT	IL-15 D	HO-NEE-UM STAFF GAUGE: LAKE STAGE	USGS STAFF GAUGE: STAGE- DISCHARGE GAUGE FOR THE WEIR AT THE USGS DAM
15/5	-	98.72	-
22/5	-	98.69	-
30/5	-	98.71	-
6/6	-	98.68	-
13/6	-	98.79	-
15/6	-	99.30	98.96
16/6	-	99.23	98.89
18/6	-	99.01	98.58
25/6	-	98.87	98.28
30/6	-	98.79	98.05
7/7	-	98.72	97.82
13/7	-	98.92	98.34
21/7	-	98.89	98.31
28/7	-	98.80	98.04
4/8	-	98.78	98.00
16/8	-	98.97	98.47
21/8	98.80	98.80	98.06
25/8	98.76	98.76	97.92
4/9	99.01	99.01	98.57
11/9	98.84	98.84	98.18
28/9	98.81	98.81	98.09

Figure D-1 to D-14 :

Figure D-1 is the explanation for Figures D-2 to D-14. Figures D-2 to D-14 represent the fluctuations of the water level in in-lake and onshore piezometers during the late spring and summer 1981. Figures D-2 to D-14 are graphical representations of Table D-2. See the symbol index at the beginning of the thesis for the symbols used in the following figures.

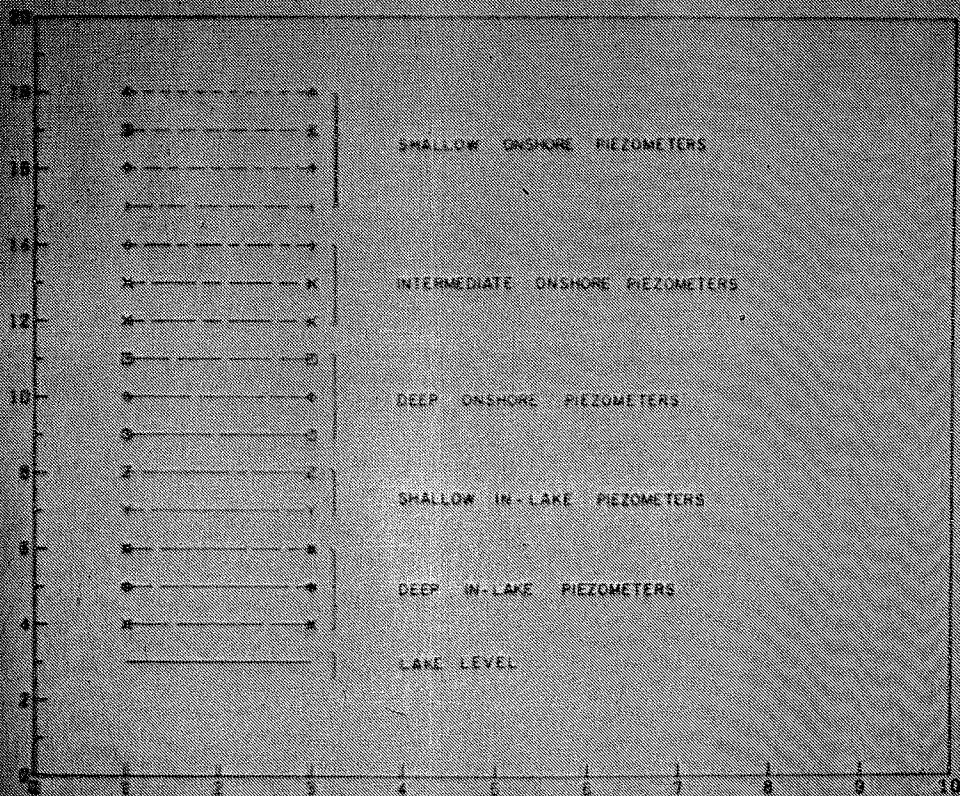


FIG. D-1

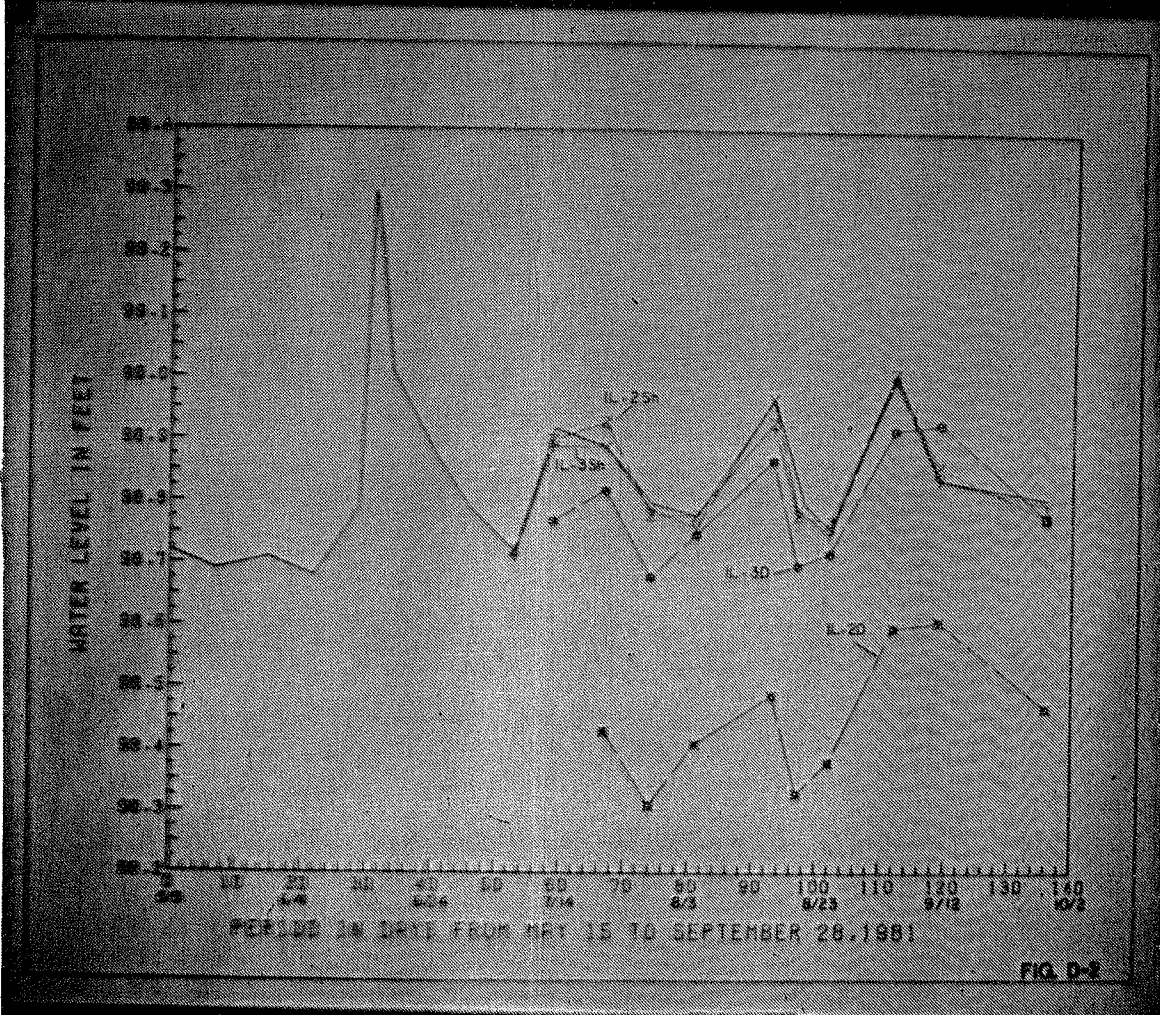


FIG. D-2

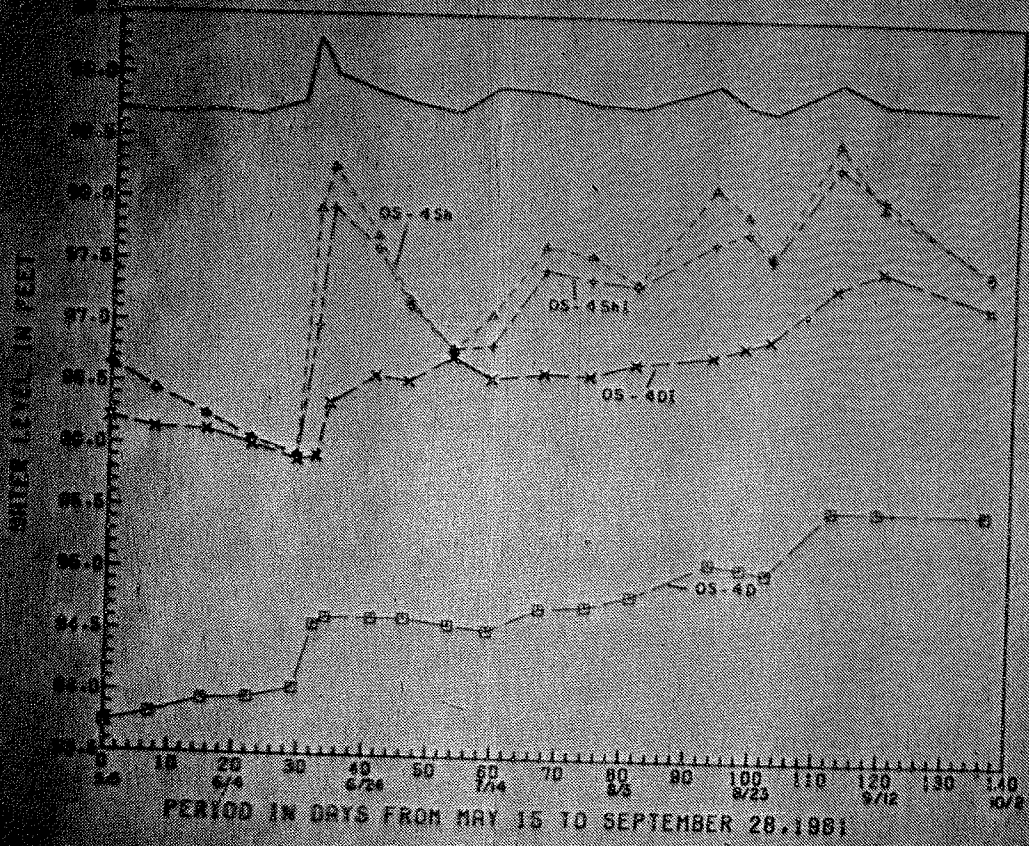


FIG. D-4

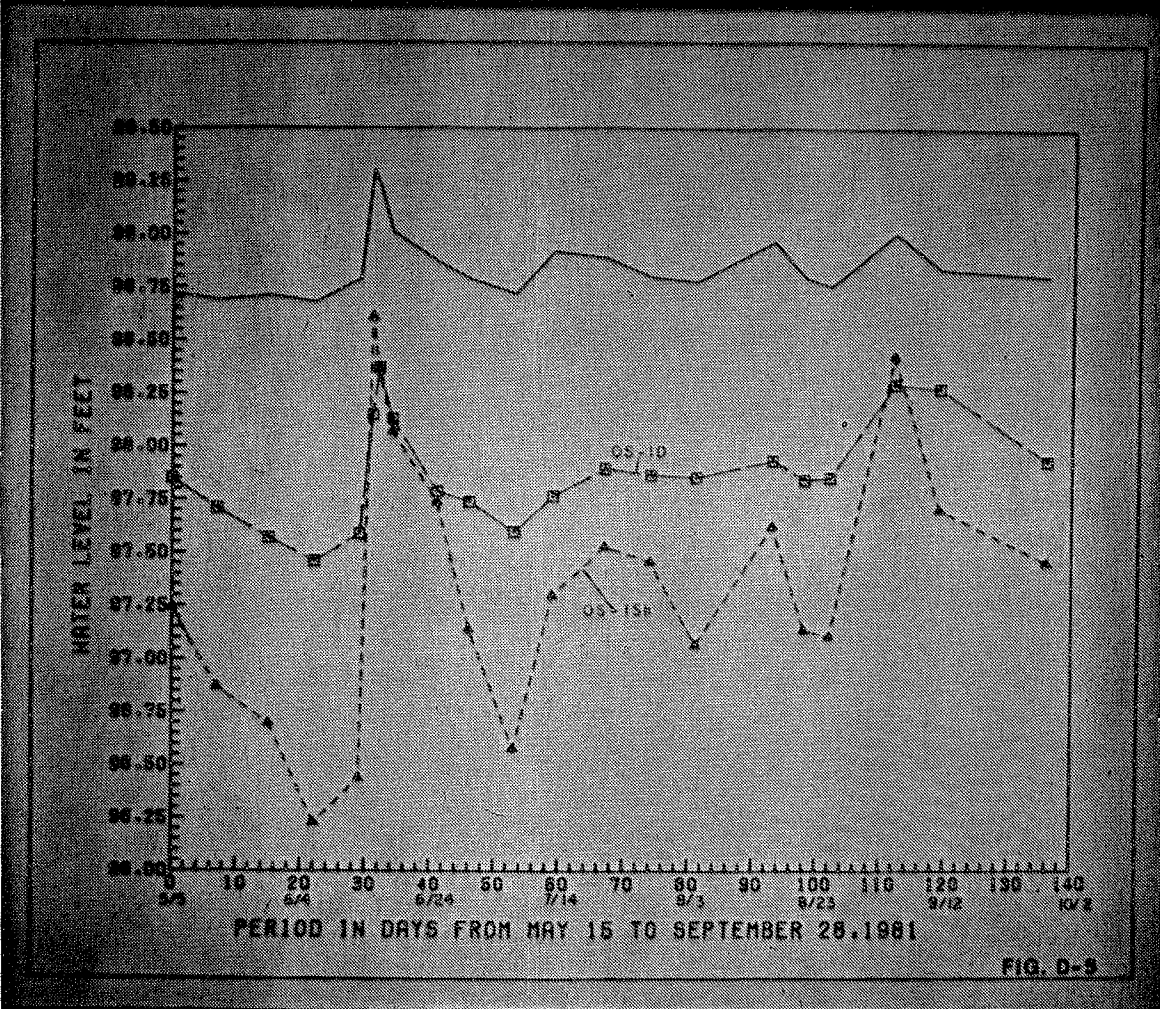


FIG. D-3

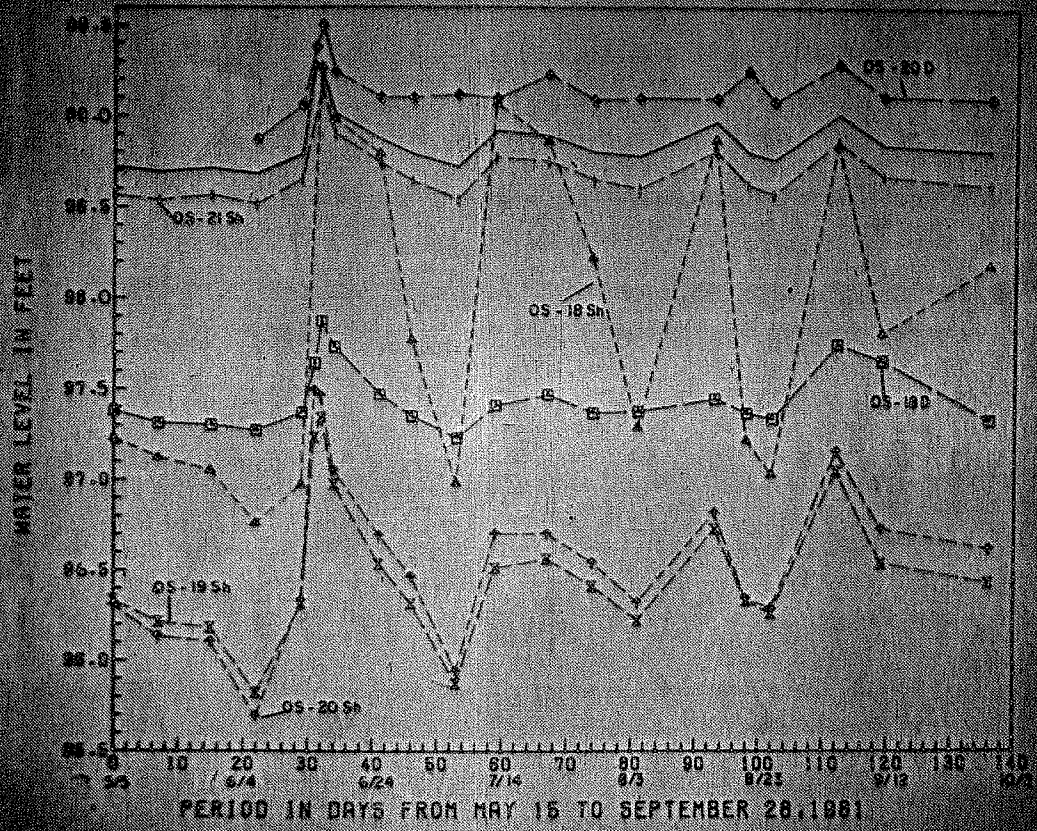


FIG. 9-5

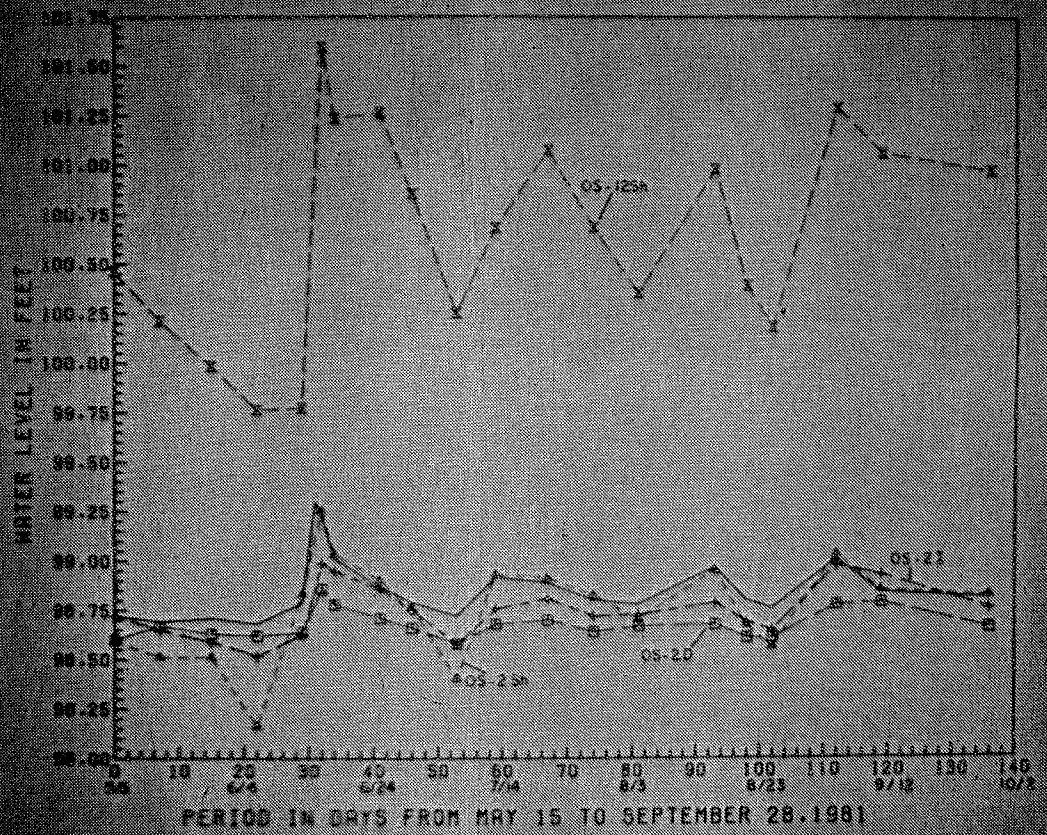


FIG. D-7

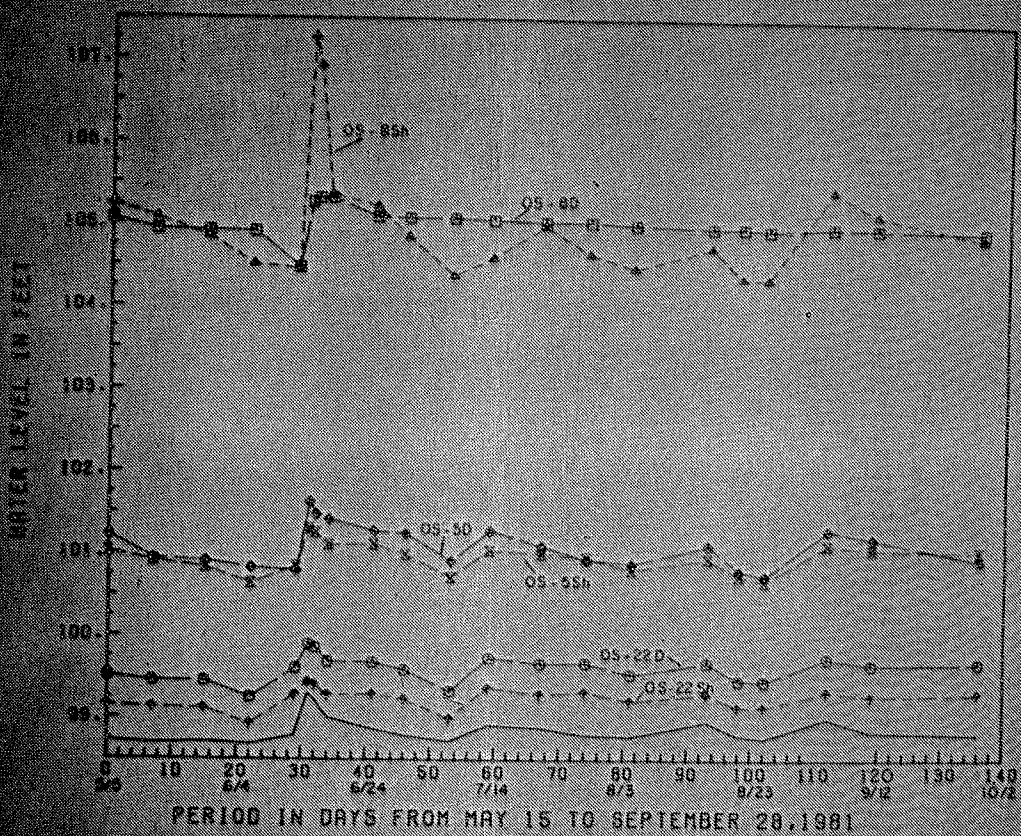


FIG. D-6

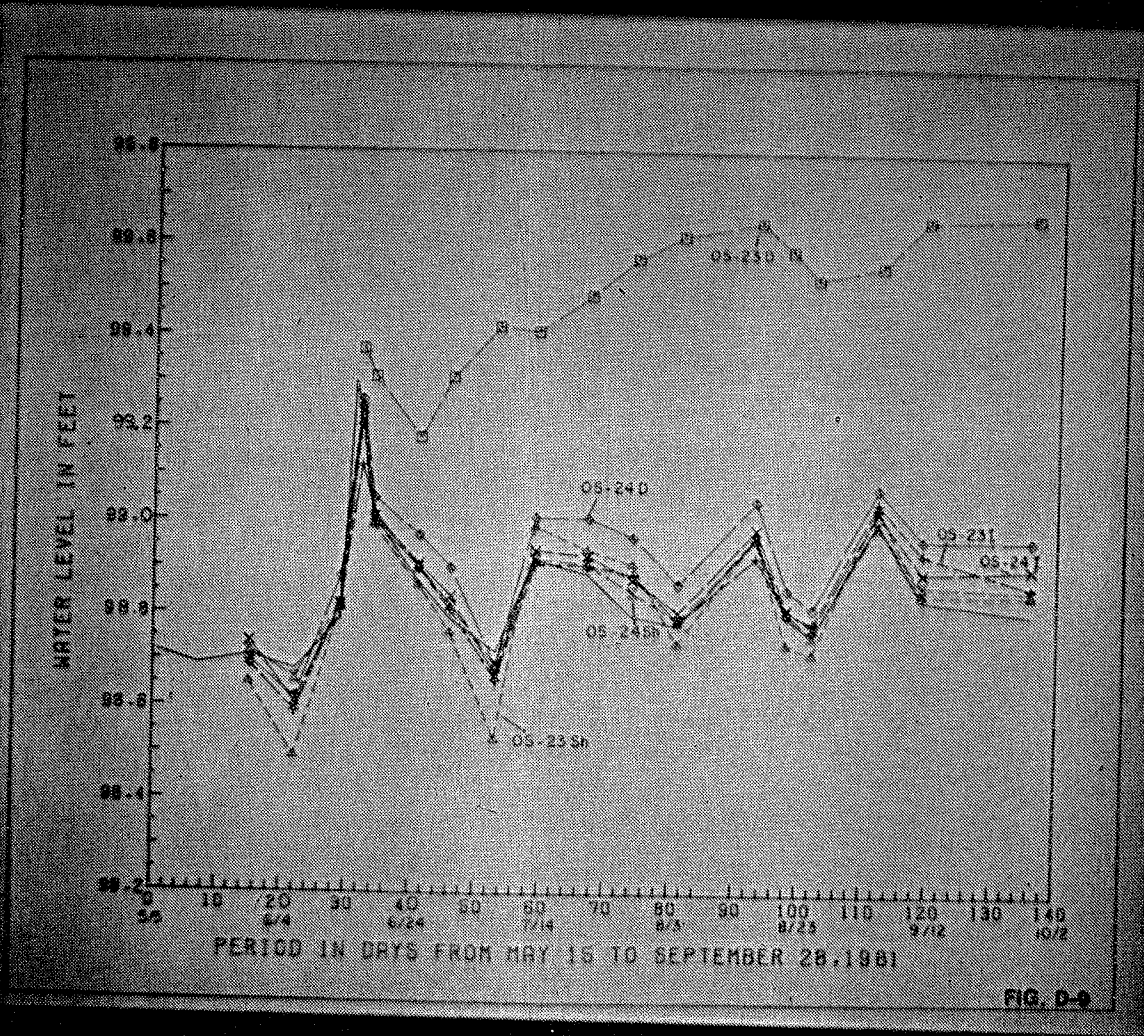


FIG. D-9

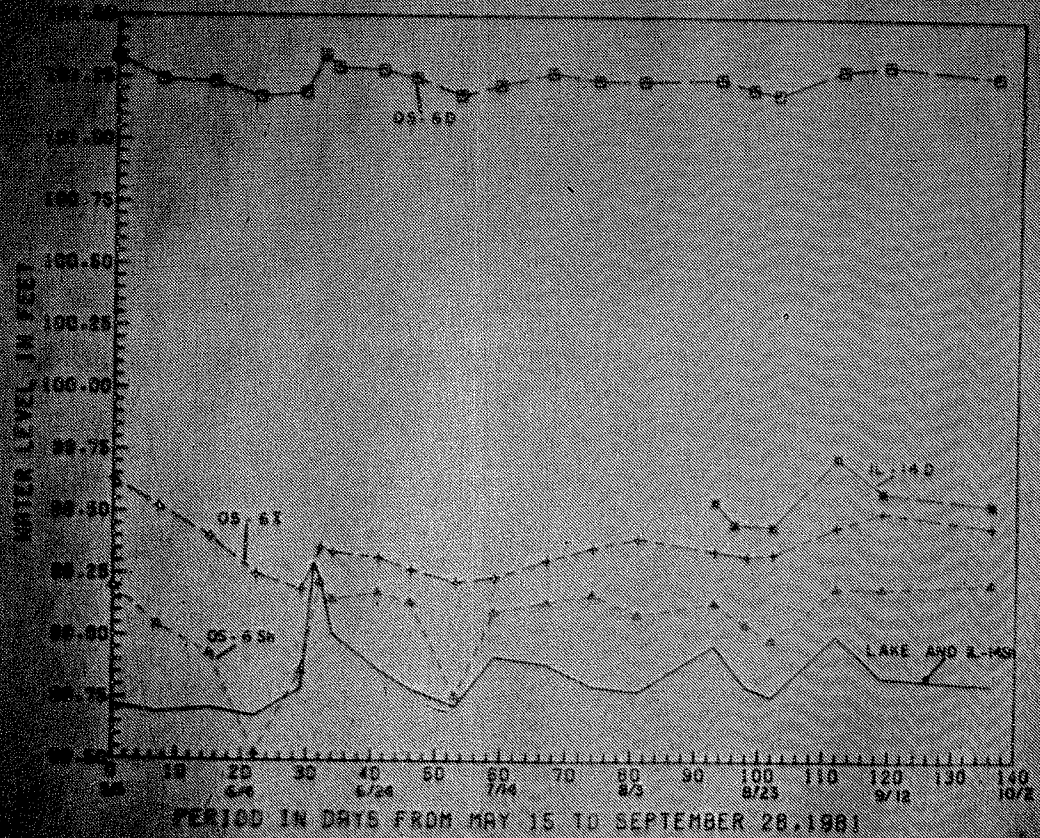


FIG. D-10

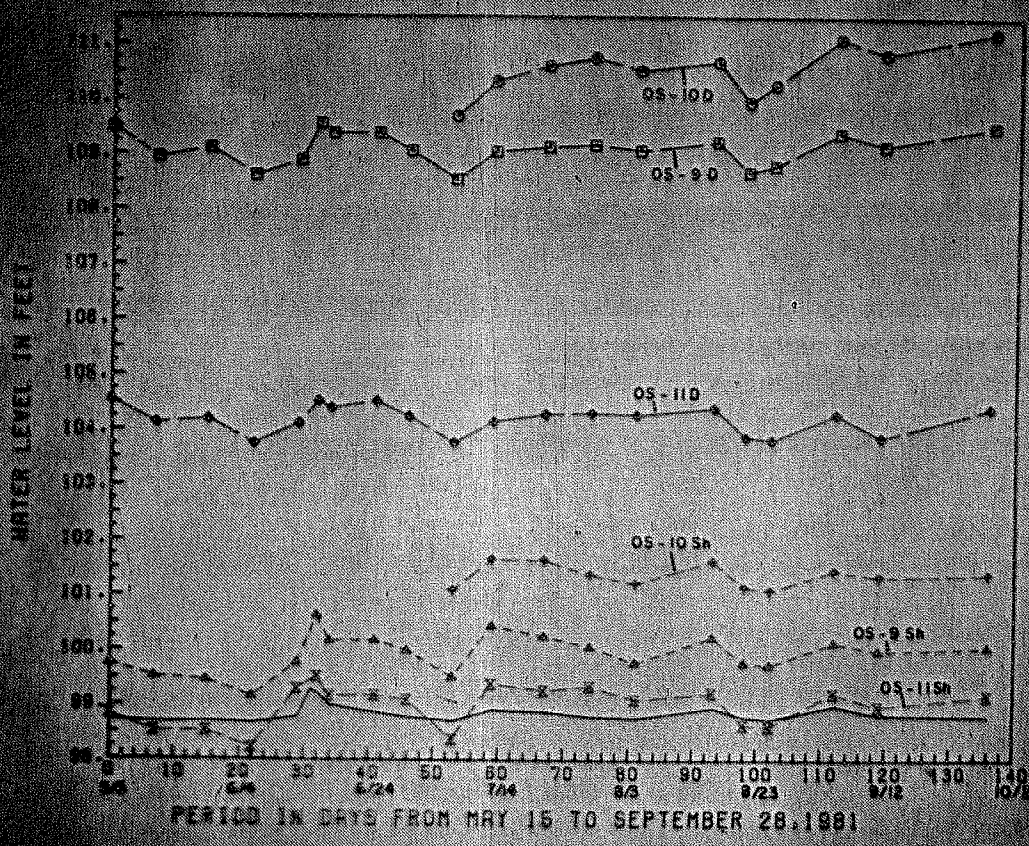
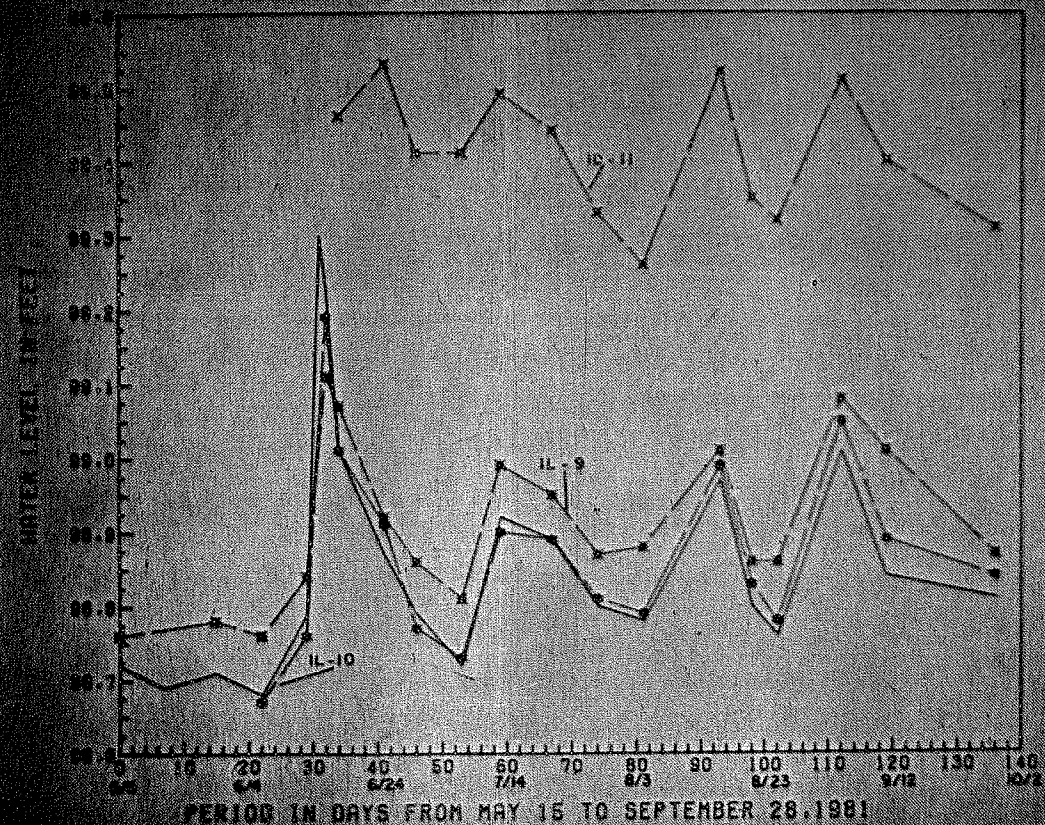


FIG. D-11



PERIOD IN DAYS FROM MAY 15 TO SEPTEMBER 28, 1981

FIG. D-12

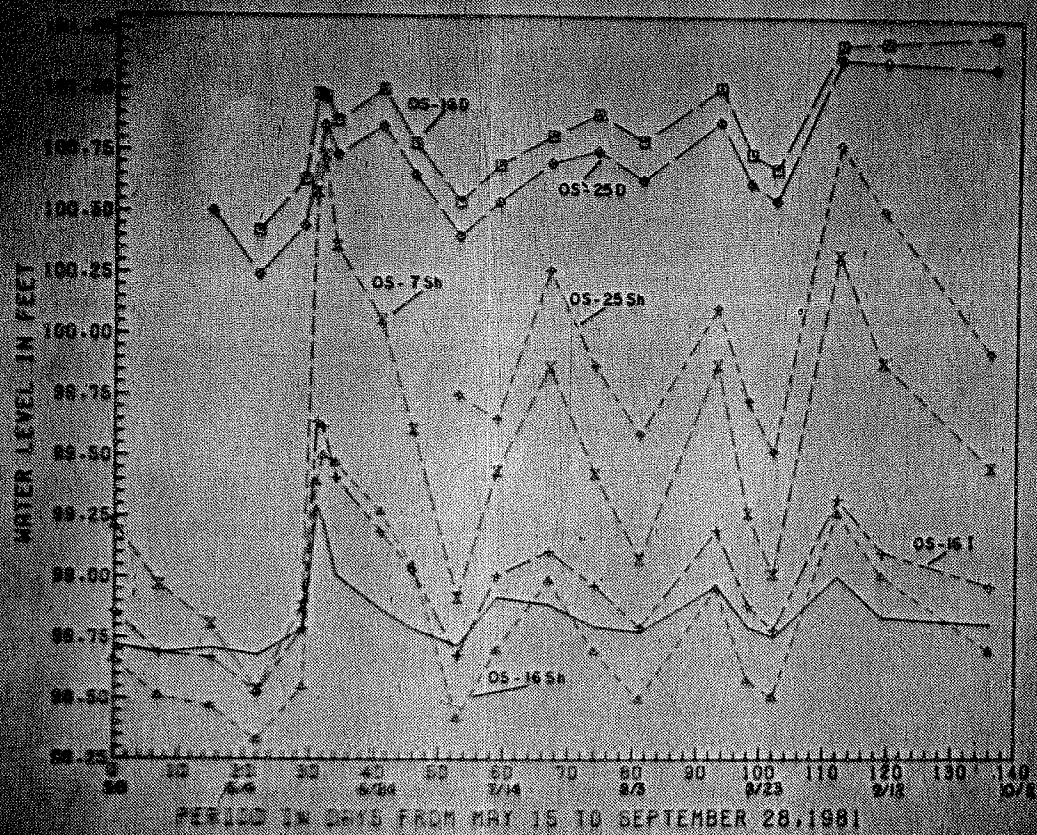


FIG. D-13

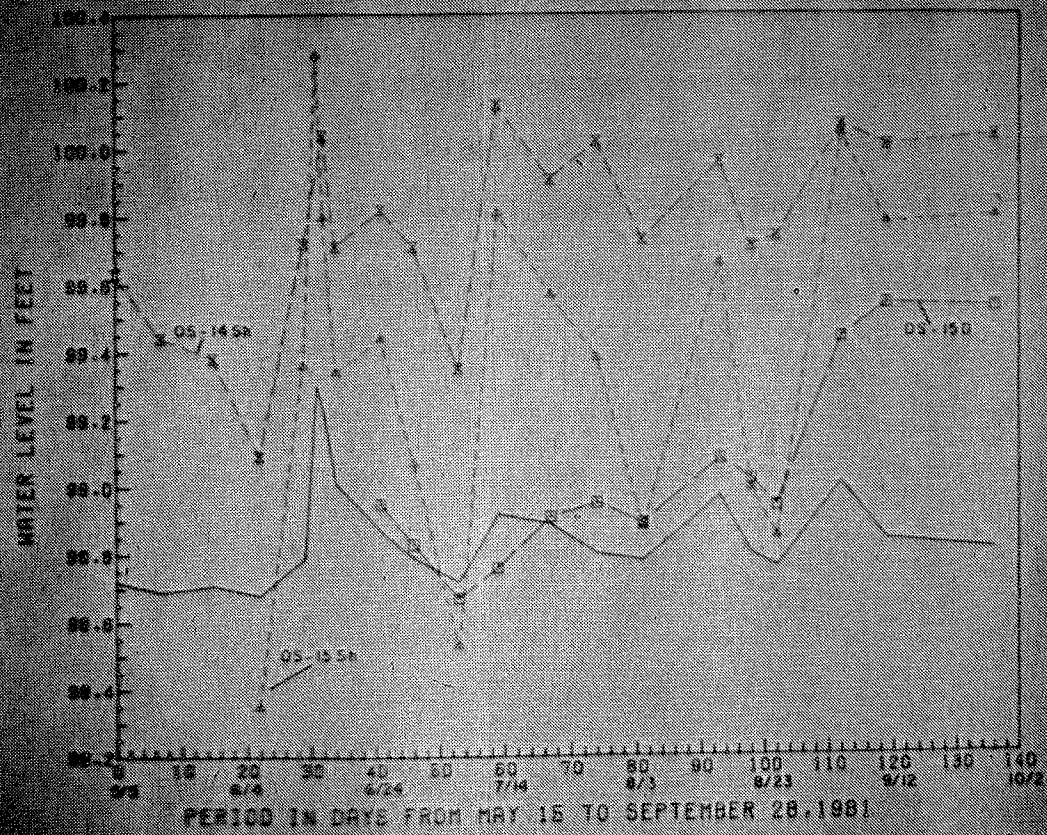


FIG. D-21

Table D-2B: Water levels in onshore piezometers, lake stage and surface-discharge gauge - Additional data for 1981. Piezometer locations are shown in Figure 5. See the symbol index at the beginning of the thesis for the symbols used in the following table.

SITE #	PIEZOMETER	1/20	1/26	2/18	2/21	2/28	3/7	3/14	5/16	6/7	8/13	8/14
OS-1	SH											
	D											
OS-3	SH				96.36	97.17	96.86	96.67				
OS-4	SH-I	96.66	96.56		96.42	94.43	96.86	96.85				
	D	92.56	92.59		92.53	92.93	92.70	92.64				
OS-5	SH		101.15		101.30		101.38	101.40	100.98			
	D		101.25		101.11		101.40	101.19	101.01			
OS-7	SH				99.83	100.11	100.03	100.00				
OS-8	D						105.25	105.15				
OS-9	SH											
OS-16	SH					100.41						
	D								98.65			
OS-18	SH								98.98			
	D								97.14	96.73		
OS-19	SH								97.28	97.34		
OS-23	SH											
	I											
	D											
OS-24	SH										98.65	98.75
	I										98.63	98.73
	D										98.63	98.79
Ho-Nee-Um Staff Gauge Lake Stage					98.78	98.85						

USGS Staff Gauge:
 Stage Discharge Gauge for the Weir at the USGS Dam

TABLE D-2B WATER LEVELS IN ONSHORE PIEZOMETERS, LAKE STAGE AND SURFACE-DISCHARGE GAUGE - ADDITIONAL DATA FOR 1981.

SITE #	PIEZOMETER	8/15	8/19	8/20	8/24
OS-18	SH			97.26	
	D			97.42	
OS-19					
OS-23	SH		98.83		
	I		98.86		
	D		99.67		
OS-24	SH	99.07	98.87		
	I	99.07	98.88		
	D	99.13	98.95		
Ho-Nee-Um Staff Gauge Lake Stage				98.82	98.76

USGS Staff Gauge:
Stage Discharge Gauge for the
Weir at the USGS Dam

TABLE D-2B WATER LEVELS IN ONSHORE PIEZOMETERS, LAKE STAGE AND SURFACE-DISCHARGE GAUGE -
ADDITIONAL DATA FOR 1981.

SITE #	PIEZOMETER	9/21		9/22				
		11:30am	12:30pm	7:00pm	9:00am	12:45pm	4:15pm	10:00pm
OS-1	SH	97.62	97.62	97.75	97.90	97.84	97.74	97.70
	D	97.97	97.97	97.97	97.85	97.84	97.87	97.81
	USGS Staff Gauge: Stage Discharge Gauge for the Weir at the USGS Dam	97.79			97.91	97.91	97.90	97.89

SITE #	PIEZOMETER	9/23		9/24	
		9:45pm	7:20pm	9:15am	
OS-1	SH	97.83	97.63	97.72	
	D	97.79	97.93	97.84	
	USGS Staff Gauge: Stage Discharge Gauge for the Weir at the USGS Dam	97.88	97.87	97.88	

TABLE D-2B WATER LEVELS IN ONSHORE PIEZOMETERS, LAKE STAGE AND SURFACE-DISCHARGE GAUGE-
ADDITIONAL DATA FOR 1981.

Table D-3 : Precipitation recorded at the University of Wisconsin-Madison Arboretum rain gauge during the period extending from May 15 to September 30, 1981. The location of the rain gauge is shown on Figure 5.

MONTH	DATE	PPT(IN) ¹	MONTH	DATE	PPT(IN) ¹	MONTH	DATE	PPT(IN) ¹	MONTH	DATE	PPT(IN) ¹
May	15	0	June	17	0	July	20	1.94	August	22	0
	16	0		18	0		21	T		23	0
	17	0		19	0		22	0		24	0
	18	0		20	0		23	T		25	.07
	19	0		21	.53		24	0		26	.19
	20	0		22	.48		25	0		27	1.00
	21	0		23	0		26	.26		28	.39
	22	0		24	.61		27	T		29	.16
	23	0		25	0		28	.65		30	.99
	24	.18		26	.12		29	.10		31	.08
	25	0		27	0		30	0		1	2.43
	26	0		28	.37		31	0		2	T
	27	0		29	T		1	0		3	0
	28	0		30	0		2	.34		4	0
	29	0		1	0		3	.13		5	0
	30	.21		2	0		4	.02		6	0
31	0	3	0	5	0	7	0				
June	1	0	July	4	0	August	8	.94			
	2	0		5	0		9	0			
	3	.02		6	0		10	0			
	4	0		7	0		11	0			
	5	0		8	0		12	0			
	6	0		9	0		13	0			
	7	0		10	0		14	0			
	8	0		11	0		15	0			
	9	.74		12	0		16	0			
	10	.15		13	1.02		17	.12			
	11	0		14	1.21		18	0			
	12	0		15	0		19	0			
	13	.65		16	.20		20	0			
	14	0		17	T		21	0			
	15	0		18	0		22	.43			
	16	2.06		19	0		23	0			

MONTH	DATE	PPT(IN) ¹
Sept. (cont.)	24	0
	25	.74
	26	.46
	27	.08
	28	0
	29	.13
	30	T

¹ All readings done at 8:00 am

TABLE D-3: PRECIPITATION RECORDED AT THE UNIVERSITY OF WISCONSIN-MADISON ARBORETUM RAIN GAUGE DURING THE PERIOD EXTENDING FROM MAY 15 TO SEPTEMBER 30, 1981.

Table D-4 : Seepage rates obtained from the seepage meters in August 1981. The locations of the seepage meter stations are shown in Figure 6. See the symbol index at the beginning of the thesis for the symbols used in the following table.

The following table shows seepage rates obtained at four seepage meter stations. However most of these rates are not believed to represent absolute seepage rates (see the discussion on seepage meters in Chapter III). The reader is referred to Figure 6 for the details on the location of the stations and seepage meters. The letter subscripts (a and b) are used when two seepage meters are placed at the same location.

Most plastic bags were prewetted* except at S0. The initial volume of water placed in the bags of the seepage meters of S1 and S2 was 500 ml. At S3, the initial volume of water poured into the bags was 500 ml on August 13, and 1000 ml on August 14 and 15.

* rinsed with lake water inside and outside, and then partially filled with lake water before the measurement started.

SEEPAGE RATE IN ML/SEC OBTAINED FROM SEEPAGE METERS IN AUGUST, 1981

SEEPAGE METER STATION #	DATES OF MEASUREMENT	SEEPAGE METERS INVOLVED ¹ /SEEPAGE RATES IN ML/SEC							
S0	6/15	1	2						
		+ .051	+ .10						
S1	8/13	1	2	3	4	5	6	7	8
		+ .0083	+ .024	+ .019	+ .00091	+ .0025	+ .011	+ .018	+ .17
S2	8/13	1	2						
		+ .0016	+ .00066						
S3	8/13	1	2	3	4	5a	5b	6	7
		+ .00029	+ .00076	+ .0037	- .011	+ .013	- .016	+ .018	- .0015
	8/14	1	2	3	4	5a	5b	6a	6b
		- .000095	- .0015	- .00096	+ .00094	- .000094	+ .0019	+ .0054	+ .0025
	8/15	1a	1b	2	3	4	5a	5b	6a
		- .000092	+ .0032	+ .033	- .00043	+ .013	0.0	+ .000067	+ .0030
									+ .0048

¹See Plate 4 for the location of the seepage meter stations and for the seepage meters.

TABLE D-4: SEEPAGE RATES OBTAINED FROM THE SEEPAGE METERS IN AUGUST, 1981.

Table D-5 : Water levels and installation details for the mini-piezometers used in Lake Wingra during the summer 1981. The locations of the mini-piezometer stations are shown in Figure 6. See the symbol index at the beginning of the thesis for the symbols used in the following table.

MINI-PIEZOMETER	DISTANCE AWAY FROM SHORE (FT)	SCREEN DEPTH	DIFFERENCE IN MILLIMETERS BETWEEN THE WATER LEVEL IN THE MINI-PIEZOMETERS AND THE LAKE SURFACE, Δh ₂							EQUILIBRATION TIME ⁴ AND REMARKS		
			19	20	21	22	24	25	4		11	28
#1	a	16	3.1									6 days
	b	30	3.0									
	c	43	2.9									
#2	a	18	3.8									3 hours
	b	36	3.3									
	c	54	3.45									
#3	a	18	2.85									4 hours *affected by gases
	b	36	2.85									
	c	54	2.75									
	d	72	4.0									
#4	a	18	2.9									2.5 hours
	b	36	4.0									
	c	54	3.4									
#5	a	18	3.8					+6.0				overnight
	b	36	4.1					+11.0				
	c	54	4.0					+17.0				
#6	a	18	3.1					-4.0				7 hours *probably affected by gases
	b	36	3.0					+1.0*				
	c	54	2.8					+3.0*				
#7	a	18	2.5					-135.0				5 hours for a,b; overnight for c *affected by gases
	b	36	2.8					+6.5*				
	c	54	3.6					+1.0*				
#7 bis a	50	4.2									+9.5* +7.0* +8.0* -2.0* +10.0* +4mm* overnight. Partially clogged-response too slow to be of any use.	

EQUILIBRATION TIME⁴ AND REMARKS

DIFFERENCE IN MILLIMETERS BETWEEN THE WATER LEVEL IN THE MINI-PIEZOMETERS AND THE LAKE SURFACE, Δh^3

MINI-PIEZOMETER¹ DISTANCE SCREEN AWAY FROM DEPTH² SHORE (FT)

	AUGUST 1981				SEPTEMBER 1981				
	19	20	21	22	24	25	4	11	28
#8 a				0.0					overnight
b				0.0					
c				0.0					
#9 a			+70.0	+71.0	+74.0	+73.0	+81.0	+87.0	+93.0
b				+42.0	+42.0	+42.0	+49.0	+50.0	+57.0
c				+28.0	+28.5	+28.0	+30.0	+33.0	+36.0
#10 a				+87.5	+88.0	+93.0	+97.0	+102.0	2 days

¹ Mini-piezometer station/mini-piezometer.

² Depth of the screen below the sediment-lake interface (ft).

³ To convert these Δh values to heads, add or subtract to/from the lake levels (Ho-Nee-Um gauge) given in Table D-2 in Appendix D.

⁴ Time elapsed between the emplacement of the mini-piezometer and the first reading of its water level.

TABLE D-5: WATER LEVELS AND INSTALLATION DETAILS FOR THE MINI-PIEZOMETERS USED IN LAKE WINGRA DURING THE SUMMER 1981.

Table D-6 : Spring discharge into Lake Wingra (ft³/s). Spring locations are shown in Figure 6.

SPRING	DATE(S) OF MEASUREMENT	VELOCITY AREA METHOD	TRACER DILUTION METHOD	FLOATER TECHNIQUE	OTHER
1a	8/14/81	.11	.093	-	-
	8/17/81	.22	.25	-	-
	3/5/82	.17	-	-	-
1b	3/5/82	.068	-	-	-
2a	8/17/81	.68	.51	-	-
2b	8/17/81	.14	.17	-	-
3	8/17/81	1.15	.92	-	-
4	8/17/81	-	-	.053	-
5	8/17/81	-	-	-	minor
6	8/17/81	-	-	.003	-
7	8/17/81	.46	-	-	-
8	8/17/81	.38	-	-	-
9	-	-	-	-	*see below

* Not discharging into Lake Wingra

TABLE D-6: SPRING DISCHARGE INTO LAKE WINGRA (FT³/SEC.).

APPENDIX E :

Comparison of the Monthly Precipitation that Fell over the Lake Wingra Area in 1981, with the Ten Year Average Monthly Precipitation for the Same Area and the 30 Year Average Monthly Precipitation for the Madison Area.

Compared to the late spring and summer precipitation that fell from 1971 to 1980, over the Arboretum precipitation gauge, or to the average late spring and summer precipitation recorded for the Madison area over a 30 year period (1941 - 1970), at the Dane County Regional Airport, the late spring and summer of 1981, should be considered as significantly wetter than normal (see Table E-1 and also see precipitation data in Table D-3).

The total precipitation recorded at the Arboretum station for June, July, August and September 1981, exceeds by about 38% the average precipitation for these four months measured at that station during a period starting in 1971 and ending in 1980, and by about 55% the long-term average precipitation expected over the Madison area, based on a 30 year period beginning in 1941. The reverse is true for the first five months of 1981: about 43% and 22% less precipitation fell over Lake Wingra relative to the Arboretum ten year average and to the 30 year average precipitation for the Madison area, respectively. On the whole, the Arboretum gauge, which began operating in 1971, catches more rain or snow than the airport gauge, and so the long term average precipitation for the Lake Wingra area would probably exceed that for the Madison area, predicted by the airport gauge. This is illustrated in Figure E-1 which shows the average precipitation recorded at the two gauges, based on a seven year period.

TABLE E-1 : Monthly precipitation that fell over the Lake Wingra area during the first nine months of 1981, compared with the ten year average (1971-1980) monthly precipitation that fell over the same area and with the 30 year average (1941-1970) monthly precipitation computed for the madison area, for the first nine months of each year. The measurements are in inches.

MONTH	ARBORETUM GAUGE 1981 PRECIPITATION	ARBORETUM GAUGE 1971 - 1980 AVERAGE PRECIPITATION	DANE COUNTY REGIONAL AIRPORT (MADISON) 1941 - 1970 AVERAGE PRECIPITATION
Jan.	0.58	1.39	1.25
Feb.	2.78	1.48	0.95
Mar.	0.42	2.94	1.93
Apr.	3.55	4.42	2.66
May	0.67	3.69	3.41
Jun.	5.73	4.19	4.33
Jul.	5.38	4.04	3.81
Aug.	6.09	4.63	3.05
Sep.	5.33	3.45	3.36

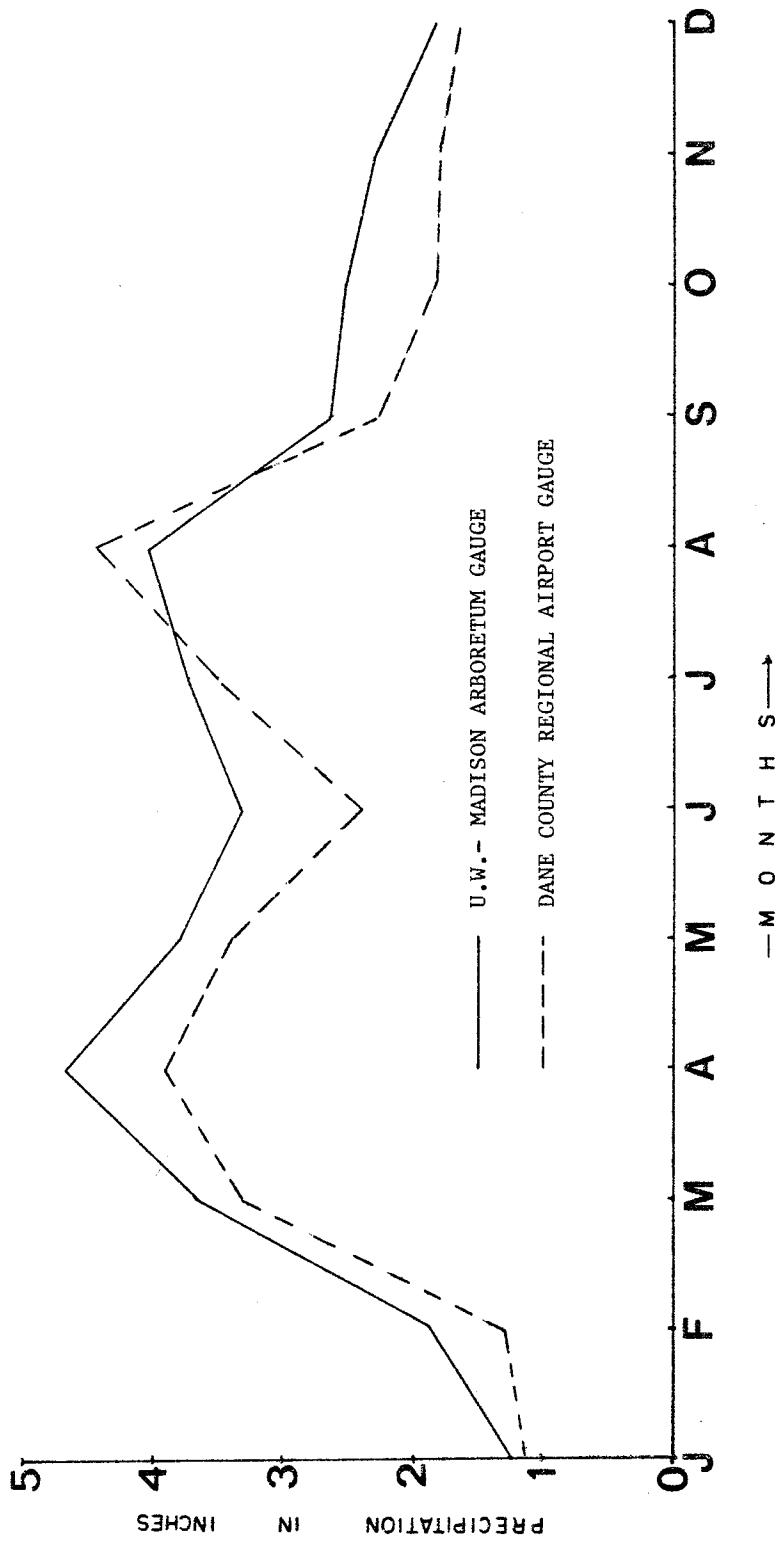


FIGURE E-1 : Average monthly precipitation recorded at the University of Wisconsin - Madison Arboretum gauge and at the Dane County regional airport gauge, for a period of seven years (1971 - 1977).

APPENDIX F :

Computer Program Used in the Correlation (Precipitation with Groundwater Levels) Analysis (See Text for the Details).

```

BRUN PENNEQUIN,7972,3917203782,31.50
#FDR,18
FORTRAN=MACC 1.178-04/14/82-23155114 NAMES
00101 1. DIMENSION A(270),P(270),AM(25),WATLEV(25)
00102 2. COMMON AM,WATLEV,N,JUNE15
00103 3. INTEGER JUNE15
00104 4. READ(5,5)A(1),M,ALPHA,N
00112 5. 5 FORMAT(F5.2,I5,F5.2,I5)
00113 6. PRINT 10
00115 7. PRINT 15 A(1),M,ALPHA,N
00123 8. 10 FORMAT(3X,'INITIAL PPT INDEX',3X,'#DAYS FOR ANT,PPT',15X,'ALPHA',3
    9. 1X,'WATER LEV. READ. ')
00124 10. 15 FORMAT(F20.2,I20,F20.2,I20)
00125 11. READ(5,20)(P(I),I=2,M)
00133 12. 20 FORMAT(16F5.2)
00134 13. WRITE(6,20)(P(I),I=2,M)
00142 14. PRINT 22
00144 15. 22 FORMAT(1X,'ANTECEDENT MOISTURE',5X,'RAIN',5X,'ALPHA')
00145 16. DO 30 I=2,M
00150 17. A(I)=ALPHA+A(I-1)+P(I)
00151 18. PRINT 25 I,A(I),P(I),ALPHA
00157 19. 25 FORMAT(6X,'A(',I3,')=',1X,F5.2,6X,F4.2,6X,F4.2)
00160 20. 30 CONTINUE
00162 21. AM(1)=A(129)
00163 22. AM(2)=A(136)
00164 23. AM(3)=A(144)
00165 24. AM(4)=A(151)
00166 25. AM(5)=A(158)
00167 26. AM(6)=A(160)
00170 27. AM(7)=A(161)
00171 28. AM(8)=A(163)
00172 29. AM(9)=A(170)
00173 30. AM(10)=A(175)
00174 31. AM(11)=A(182)
00175 32. AM(12)=A(188)
00176 33. AM(13)=A(196)
00177 34. AM(14)=A(203)
00200 35. AM(15)=A(210)
00201 36. AM(16)=A(222)
00202 37. AM(17)=A(227)
00203 38. AM(18)=A(231)
00204 39. AM(19)=A(241)
00205 40. AM(20)=A(248)
00206 41. AM(21)=A(265)
00207 42. READ(5,100)(WATLEV(K),K=1,N)
00215 43. 100 FORMAT(5X,F5.2)
00216 44. PRINT 105
00220 45. 105 FORMAT(1X,'OS=1SH')
00221 46. JUNE15=1
00222 47. CALL CORCDF
00223 48. READ(5,110)(WATLEV(K),K=1,N)
00231 49. 110 FORMAT(10X,F5.2)
00232 50. PRINT 115

```

00234	51.	115	FORMAT(1X,'08-1D')
00235	52.		JUNE15=1
00236	53.		CALL CORCOF
00237	54.		READ(5,120)(WATLEV(K),K=1,N)
00245	55.	120	FORMAT(35X,F5.2)
00246	56.		PRINT 125
00250	57.	125	FORMAT(1X,'08-5SH')
00251	58.		JUNE15=1
00252	59.		CALL CORCOF
00253	60.		READ(5,130)(WATLEV(K),K=1,N)
00261	61.	130	FORMAT(40X,F5.2)
00262	62.		PRINT 135
00264	63.	135	FORMAT(1X,'08-5D')
00265	64.		JUNE15=1
00266	65.		CALL CORCOF
00267	66.		READ(5,140)(WATLEV(K),K=1,N)
00275	67.	140	FORMAT(5X,F5.2)
00276	68.		PRINT 145
00300	69.	145	FORMAT(1X,'08-7SH')
00301	70.		JUNE15=1
00302	71.		CALL CORCOF
00303	72.		READ(5,150)(WATLEV(K),K=1,N)
00311	73.	150	FORMAT(20X,F5.2)
00312	74.		PRINT 155
00314	75.	155	FORMAT(1X,'08-8SH')
00315	76.		JUNE15=1
00316	77.		CALL CORCOF
00317	78.		READ(5,160)(WATLEV(K),K=1,N)
00325	79.	160	FORMAT(10X,F5.2)
00326	80.		PRINT 165
00330	81.	165	FORMAT(1X,'08-16SH')
00331	82.		JUNE15=1
00332	83.		CALL CORCOF
00333	84.		READ(5,170)(WATLEV(K),K=1,N)
00341	85.	170	FORMAT(15X,F5.2)
00342	86.		PRINT 175
00344	87.	175	FORMAT(1X,'08-16D')
00345	88.		JUNE15=1
00346	89.		CALL CORCOF
00347	90.		READ(5,180)(WATLEV(K),K=1,N)
00355	91.	180	FORMAT(10X,F5.2)
00356	92.		PRINT 185
00360	93.	185	FORMAT(1X,'08-18D')
00361	94.		JUNE15=1
00362	95.		CALL CORCOF
00363	96.		READ(5,190)(WATLEV(K),K=1,N)
00371	97.	190	FORMAT(15X,F5.2)
00372	98.		PRINT 195
00374	99.	195	FORMAT(1X,'08-19SH')
00375	100.		JUNE15=1
00376	101.		CALL CORCOF
00377	102.		READ(5,200)(WATLEV(K),K=1,N)
00405	103.	200	FORMAT(20X,F5.2)
00406	104.		PRINT 205
00410	105.	205	FORMAT(1X,'08-20SH')
00411	106.		JUNE15=1
00412	107.		CALL CORCOF

00413	108.	READ(5,210)(WATLEV(K),K=1,N)
00421	109.	210 FORMAT(45X,F5.2)
00422	110.	PRINT 215
00424	111.	215 FORMAT(1X,'05-228H')
00425	112.	JUNE15=1
00426	113.	CALL CORCOF
00427	114.	READ(5,220)(WATLEV(K),K=1,N)
00435	115.	220 FORMAT(50X,F5.2)
00436	116.	PRINT 225
00440	117.	225 FORMAT(1X,'05-22D')
00441	118.	JUNE15=1
00442	119.	CALL CORCOF
00443	120.	READ(5,230)(WATLEV(K),K=1,N)
00451	121.	230 FORMAT(5X,F5.2)
00452	122.	PRINT 235
00454	123.	235 FORMAT(1X,'05-28H')
00455	124.	JUNE15=2
00456	125.	CALL CORCOF
00457	126.	READ(5,240)(WATLEV(K),K=1,N)
00465	127.	240 FORMAT(10X,F5.2)
00466	128.	PRINT 245
00470	129.	245 FORMAT(1X,'05-2I')
00471	130.	JUNE15=2
00472	131.	CALL CORCOF
00473	132.	READ(5,250)(WATLEV(K),K=1,N)
00501	133.	250 FORMAT(15X,F5.2)
00502	134.	PRINT 255
00504	135.	255 FORMAT(1X,'05-2D')
00505	136.	JUNE15=2
00506	137.	CALL CORCOF
00507	138.	READ(5,260)(WATLEV(K),K=1,N)
00515	139.	260 FORMAT(15X,F5.2)
00516	140.	PRINT 265
00520	141.	265 FORMAT(1X,'05-3SH')
00521	142.	JUNE15=2
00522	143.	CALL CORCOF
00523	144.	READ(5,270)(WATLEV(K),K=1,N)
00531	145.	270 FORMAT(20X,F5.2)
00532	146.	PRINT 275
00534	147.	275 FORMAT(1X,'05-4SH')
00535	148.	JUNE15=2
00536	149.	CALL CORCOF
00537	150.	READ(5,280)(WATLEV(K),K=1,N)
00545	151.	280 FORMAT(25X,F5.2)
00546	152.	PRINT 285
00550	153.	285 FORMAT(1X,'05-4SHI')
00551	154.	JUNE15=2
00552	155.	CALL CORCOF
00553	156.	READ(5,290)(WATLEV(K),K=1,N)
00561	157.	290 FORMAT(30X,F5.2)
00562	158.	PRINT 295
00564	159.	295 FORMAT(1X,'05-4DI')
00565	160.	JUNE15=2
00566	161.	CALL CORCOF
00567	162.	READ(5,300)(WATLEV(K),K=1,N)
00575	163.	300 FORMAT(35X,F5.2)
00576	164.	PRINT 305

00600	165.	305	FORMAT(1X,'08=4D')
00601	166.		JUNE15=2
00602	167.		CALL CORCOF
00603	168.		READ(5,310)(WATLEV(K),K=1,N)
00611	169.	310	FORMAT(5X,F5.2)
00612	170.		PRINT 315
00614	171.	315	FORMAT(1X,'08=68H')
00615	172.		JUNE15=2
00616	173.		CALL CORCOF
00617	174.		READ(5,320)(WATLEV(K),K=1,N)
00625	175.	320	FORMAT(15X,F5.2)
00626	176.		PRINT 325
00630	177.	325	FORMAT(1X,'08=6D')
00631	178.		JUNE15=2
00632	179.		CALL CORCOF
00633	180.		READ(5,330)(WATLEV(K),K=1,N)
00641	181.	330	FORMAT(20X,F5.2)
00642	182.		PRINT 335
00644	183.	335	FORMAT(1X,'08=98H')
00645	184.		JUNE15=2
00646	185.		CALL CORCOF
00647	186.		READ(5,340)(WATLEV(K),K=1,N)
00655	187.	340	FORMAT(25X,F5.2)
00656	188.		PRINT 345
00660	189.	345	FORMAT(1X,'08=9D')
00661	190.		JUNE15=2
00662	191.		CALL CORCOF
00663	192.		READ(5,350)(WATLEV(K),K=1,N)
00671	193.	350	FORMAT(40X,F5.2)
00672	194.		PRINT 355
00674	195.	355	FORMAT(1X,'08=118H')
00675	196.		JUNE15=2
00676	197.		CALL CORCOF
00677	198.		READ(5,360)(WATLEV(K),K=1,N)
00705	199.	360	FORMAT(45X,F5.2)
00706	200.		PRINT 365
00710	201.	365	FORMAT(1X,'08=11D')
00711	202.		JUNE15=2
00712	203.		CALL CORCOF
00713	204.		READ(5,370)(WATLEV(K),K=1,N)
00721	205.	370	FORMAT(35X,F5.2)
00722	206.		PRINT 375
00724	207.	375	FORMAT(1X,'08=148H')
00725	208.		JUNE15=2
00726	209.		CALL CORCOF
00727	210.		READ(5,380)(WATLEV(K),K=1,N)
00735	211.	380	FORMAT(40X,F5.2)
00736	212.		PRINT 385
00740	213.	385	FORMAT(1X,'08=178H')
00741	214.		JUNE15=2
00742	215.		CALL CORCOF
00743	216.		READ(5,390)(WATLEV(K),K=1,N)
00751	217.	390	FORMAT(45X,F5.2)
00752	218.		PRINT 395
00754	219.	395	FORMAT(1X,'08=17D')
00755	220.		JUNE15=2
00756	221.		CALL CORCOF

```

00757 222.      READ(5,400)(MATLEV(K),K=1,N)
00765 223.      400 FORMAT(5X,F5.2)
00766 224.      PRINT 405
00770 225.      405 FORMAT(1X,'OS=18SH')
00771 226.      JUNE15=2
00772 227.      CALL CORCOF
00773 228.      READ(5,410)(MATLEV(K),K=1,N)
01001 229.      410 FORMAT(30X,F5.2)
01002 230.      PRINT 415
01004 231.      415 FORMAT(1X,'OS=21SH')
01005 232.      JUNE15=2
01006 233.      CALL CORCOF
01007 234.      STOP
01010 235.      END

```

END OF COMPILATION; NO DIAGNOSTICS.

```

#FOR,18 CORCOF
FORTRAN=MACC 1.178-04/14/82=23:55:19 (.0) CORCOF
00101 1.      SUBROUTINE CORCOF
00102 2.      COMMON AM,MATLEV,N,JUNE15
00103 3.      DIMENSION AM(25),MATLEV(25)
00104 4.      50 SUMX=0.0
00105 5.      SUMY=0.0
00106 6.      DO 60 K=1,N
00111 7.      IF(JUNE15.EQ.1)GOTO 55
00113 8.      IF(K.EQ.6)GOTO 60
00115 9.      55 SUMX=SUMX+AM(K)
00116 10.     SUMY=SUMY+MATLEV(K)
00117 11.     60 CONTINUE
00121 12.     RN=N
00122 13.     IF(JUNE15.EQ.1)GOTO 65
00124 14.     RN=N-1
00125 15.     65 AVGM=SUMX/RN
00126 16.     AVGL=SUMY/RN
00127 17.     RNUM=0.0
00130 18.     DENX=0.0
00131 19.     DENY=0.0
00132 20.     DO 75 K=1,N
00135 21.     IF(JUNE15.EQ.1)GOTO 70
00137 22.     IF(K.EQ.6)GOTO 75
00141 23.     70 RNUM=RNUM+((AM(K)-AVGM)*(MATLEV(K)-AVGL))
00142 24.     DENX=DENX+((AM(K)-AVGM)**2.)
00143 25.     DENY=DENY+((MATLEV(K)-AVGL)**2.)
00144 26.     75 CONTINUE
00146 27.     R=RNUM/SQRT(DENX*DENY)
00147 28.     WRITE(6,80)R
00152 29.     80 FORMAT(1X,'PEARSON PRODUCT MOMENT CORRELATION COEFFICIENT,R=',F6.4)
00152 30.     1//)
00153 31.     RETURN
00154 32.     END

```

END OF COMPILATION; NO DIAGNOSTICS.

```

#XGT
MAP 30R1U06 SL156 04/14/82 23:55:23
AFCH STATUS OF OUTPUT ELEMENT=UNKNOWN
QUARTER/THIRD WORD INSENSITIVE

```

Appendix G :

Determination of the Hydraulic Conductivity

One of the most difficult aquifer parameters to determine is hydraulic conductivity (K); many methods exist (some better than others), including use of grain size distributions, aquifer tests and computer modeling. In most low budget studies, however, K is either conveniently assumed from a limited amount of field data or taken from values previously published for similar material. In this study, K was calculated from a combination of slug and/or bail tests, a water level recession curve analysis and previously published data.

Slug and Bail Tests

After a sudden stress (water suddenly removed or added) has been imparted to the water column in a piezometer, Hvorslev (1951) reasoned that the rate of inflow q through the piezometer point at time t , is proportional to the hydraulic conductivity K and to the unrecovered head $(H-h)$, and can be summarized by the following equation:

$$\frac{dh}{H-h} = \frac{F K}{\pi r^2} dt \quad (G-1)$$

where h is the head at time t in the piezometer; dh is the change in head that takes place during the time interval dt ; H is the initial equilibrium level in the piezometer; F is a shape factor that depends on both the shape and dimension of the piezometer; K is the hydraulic

conductivity and r , the inside radius of the piezometer.

Starting with equation G-1, Hvorslev (1951) derived several equations for different piezometer or piezometer-screen geometries.

Those used in this study are listed below:

$$K_h = \frac{d^2}{8} \frac{\ln \left(\frac{2mL}{D} \right)}{LT} \quad \text{for } \frac{mL}{D} > 4 \quad (G-2)$$

$$K_m = \frac{\pi D}{11T} \quad \text{for } d = D \quad (G-3)$$

$$K'_v = \frac{\pi}{11} \frac{\frac{D}{m} + L}{T} \quad \text{for } K'_v = K_v \text{ and } d = D \quad (G-4)$$

where : K_h is the horizontal hydraulic conductivity of the sediments, cm/s.

K_v is the vertical hydraulic conductivity of the sediments, cm/s.

K'_v is the vertical hydraulic conductivity of the sediments in the casing, cm/s.

$K_m = (K_h \cdot K_v)^{1/2}$ (mean hydraulic conductivity), cm/s.

d is the inside diameter of the pipe, cm.

D is the inside diameter of the well screen or sand point, cm.

$$m = (K_h/K_v)^{1/2}.$$

L is the length of the well screen or sand point (for equation G-2) or of the sediments in the casing (equation G-4), cm.

T is the basic time lag, sec. See Hvorslev (1951) for the definition of the basic time lag.

For more details on piezometer tests, the reader is referred to the excellent discussion by Hvorslev (1951).

The results of the slug/bail tests are listed in Table G-1. Some of the main factors which may have affected the piezometer tests, principally by increasing the lag time, include gas trapped in the sediments, gas bubbles accumulating near the piezometer screen and the clogging of the screen by fines as the test progresses (Hvorslev, 1951). However, in view of the very consistent results obtained combined with the types of piezometers used and the modest disturbance imparted to the system during the tests, these problems were probably minor, except in two cases (OS-1, OS-17) for which the data were discarded. Most of the results are in fact compatible with the geology of the area and data previously published for similar sediments. Nonetheless, the major handicap of slug/bail tests stems from the fact that they only generate K-values for the sediments that surround the piezometers, and so ideally many such tests should be performed to determine an average K for the area. This was unfortunately not possible everywhere around Lake Wingra, because of the limited number of

TABLE G-1 : K-values in ft/s calculated from the slug/bail tests, assuming four different anisotropies for the sediments around and below Lake Wingra

PIEZOMETER #	K_h or K_v ¹	K_h or K_v ¹	K_h or K_v ¹	K_h or K_v ¹	$K_m = (K_h - K_v)^{1/2}$ (x 10 ⁵)
	if $K_h = K_v$ (x 10 ⁵)	if $K_h = 10K_v$ (x 10 ⁵)	if $K_h = 100K_v$ (x 10 ⁵)	if $K_h = 1000K_v$ (x 10 ⁵)	
OS-2I					1.03
OS-4D	0.017	0.024	0.031	0.038	
OS-9Sh	0.33	0.45	0.56	0.68	
OS-15D	4.08	5.47	6.85	8.24	
OS-16I	0.90 ¹	0.86 ¹	0.85 ¹	0.85 ¹	
OS-18D					2.00
OS-22D	4.00	5.35	6.71	8.06	
OS-25D	3.79	5.26	6.74	8.22	
IL-9	0.53	0.74	0.95	1.16	
MP-9a	0.56	0.76	0.96	1.17	
MP-9b	0.40	0.55	0.70	0.85	
MP-10a	2.41	3.29	4.17	5.06	

¹ for OS-16I the value listed in the table is for K_v . All other values are for K_h .

available piezometers and the variety of sediments present.

Water Level Recession Curve Analysis

It was observed (see Chapter IV) that in the upper few feet of the west marsh and wetland area, a substantial secondary permeability and a shallow lakeward groundwater flow had developed. Since no piezometers were emplaced in the shallow zone, no slug/bail tests could be performed. However, the automatic water level recorder monitored the water table in the west marsh sediments and so, an analysis of the water level recession curve was done in order to obtain a horizontal hydraulic conductivity value that could be used to quantify the lateral, lakeward groundwater movement in this zone.

Following an episode of recharge, the water level in an aquifer slowly subsides until either the pre-recharge equilibrium is reached or another recharge event takes place. In this case, recharge is induced by precipitation events and a true steady state equilibrium is never achieved; water levels essentially decay until the next storm occurs.

The stage of the water level recession curves primarily depends on two aquifer parameters: T (transmissivity) and S (storage coefficient). It is directly proportional to the former and inversely proportional to the latter. An analysis of the curve yields a value for

FIGURE G-1 : Figure showing (1) the piezometer and mini-piezometer sites (small black dots) where slug and/or bail tests were conducted and (2) the infinite strip around the automatic water level recorder (large black dot labelled "recorder") used in the water level recession curve analysis.

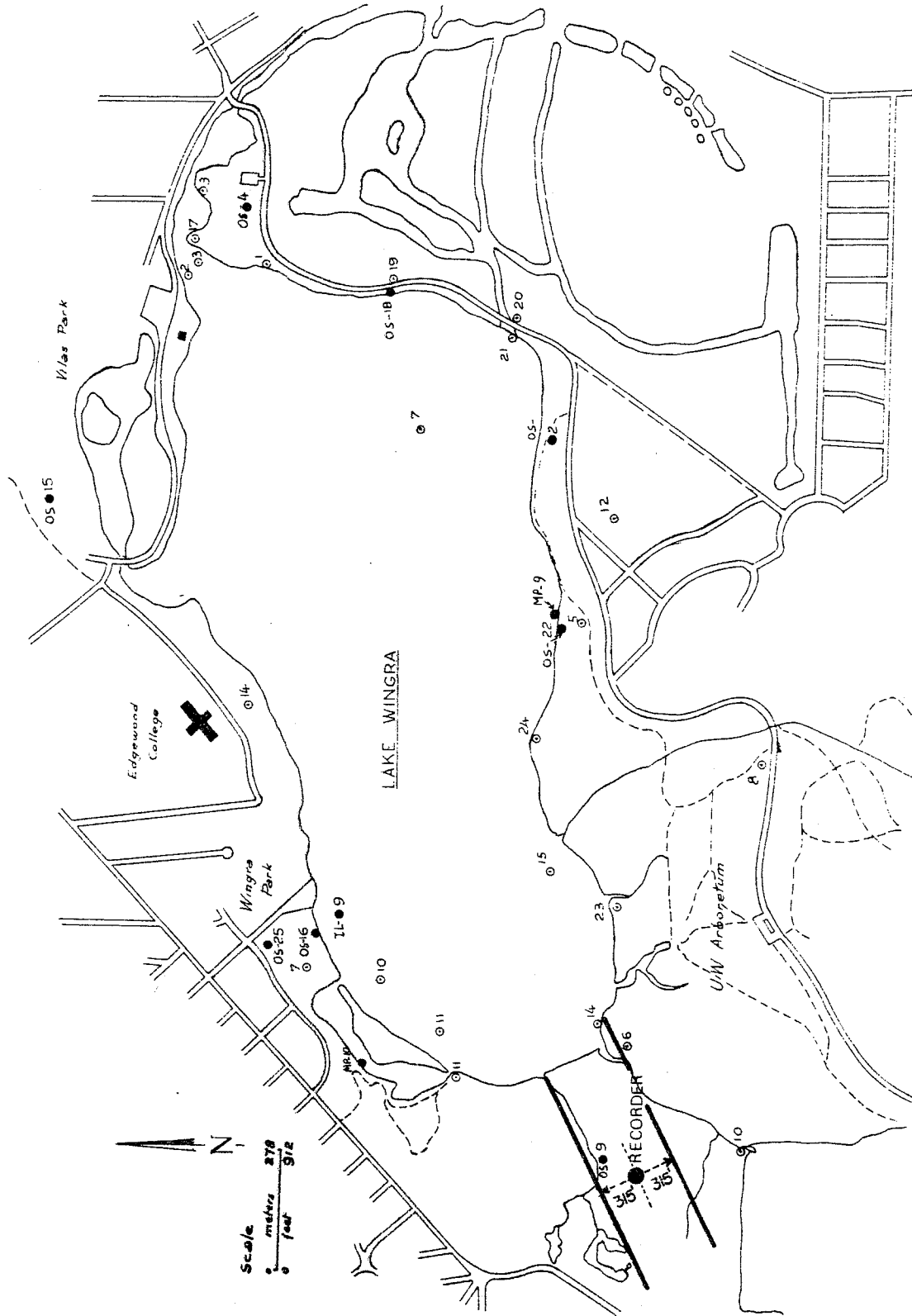


FIG. G-1

T/S.

The analytical method used to extract the T/S-value from the recession curve is derived from heat flow theory, not surprisingly since the analogy between heat flow and groundwater flow was recognized by Theis (1935) and other earlier workers. To be feasible, the method calls for a set of assumptions and simplifications (Weeks, 1964): 1. the aquifer needs to be idealized to a basic geometric shape; 2. the aquifer must be assumed homogeneous and isotropic; 3. the recharge should be uniform over the entire aquifer portion of interest, and it could either be instantaneous or at an equilibrium rate that ceases instantaneously.

The aquifer portion around the automatic water level recorder was approximated as an infinite strip, bounded by two streams (see Figure G-1). Due to the short and intense nature of most summer storms, recharge was assumed to be instantaneous. Under these assumptions, the relation between groundwater level decay after a recharge episode and T/S is given by Brown (1963) in equation G-5; it was adapted to hydrogeology from the equation describing the one-dimensional heat flow in an infinite slab, derived by Ingersoll and Zobel (1948).

$$\frac{h_t}{h_o} = \frac{4}{\pi} \sum_{n=1,3,5}^{\infty} \frac{1}{n} e^{-\left(n^2 \pi^2 \frac{T t}{w^2 S}\right)} \cdot \sin \frac{n \pi X}{w} \quad (G-5)$$

where h_o is the instantaneous change in head due to recharge.

h_t is the residual change in head remaining at time t .

n is an odd, positive integer.

T is the transmissivity.

S is the storage coefficient.

w is the width of the strip.

X is the distance from the observation well to one strip boundary

Seven major and minor storms were chosen from the automatic water level recorder charts, and were analysed following the procedure outlined by Brown (1963) and Weeks (1964). A value of .075 for S was obtained from Huff and Young (1980). The results are listed in Table G-2.

The restrictive assumptions needed to use this method cast some doubts on the validity of the results. For example, the aquifer is neither homogeneous, nor isotropic, nor is the infinite strip model the best geometric shape to represent the upper aquifer portion in the west marsh. Nevertheless, the results display relative consistency, and although the average K_h obtained is significantly larger than the other K_h -values that characterize the sediments around the lake, it is not unreasonable in view of the substantial secondary permeability that has developed in the upper few feet (root zone) of the west marsh.

TABLE G-2 : K_h -values obtained from the water level recession curve analysis. The thickness of the unconsolidated aquifer was assumed to be 150 ft (based on deep well logs, Figure 2 in Chapter I and a seismic reflection profile done during this study).

STORM DATE	K_h in ft/s
6/9	9.9×10^{-5} 6.3×10^{-5}
6/15	1.8×10^{-4} 1.0×10^{-4}
8/31	2.6×10^{-4}
9/7	2.5×10^{-4}
9/30	3.8×10^{-4}
10/3	3.7×10^{-4}
10/4	2.2×10^{-4}
average	2.1×10^{-4}

Literature

There has been essentially two previous studies which generated a value for the vertical hydraulic conductivity around Lake Wingra; Oakes et al. (1975) obtained a K_v of 7.6×10^{-7} ft/s from five shallow cores of lake bottom sediments and Huff and Young (1980) determined in the laboratory a K_v of 1.14×10^{-7} ft/s for the west marsh sediments.

APPROVED

Mary P. Anderson

Dr. Mary P. Anderson
Associate Professor
August 20, 1982