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An environmental isotope and computer flow model investigation of the freshwater aquifer in the Lake Huron to Lake Erie corridor (Ontario, Michigan).

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# AN ENVIRONMENTAL ISOTOPE AND COMPUTER FLOW MODEL INVESTIGATION OF THE FRESHWATER AQUIFER IN THE LAKE HURON TO LAKE ERIE CORRIDOR

by

Bosiljka Crnokrak

A Thesis
Submitted to the
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1991

Canadä

ABSTRACT

An environmental isotope and computer flow model investigation was undertaken to assess the distribution of groundwater flow in the freshwater aquifer located in southwestern Ontario and southeastern Michigan within the Lake Huron to Lake Erie corridor. The freshwater aquifer consists of a discontinuous thin layer of granular material located at the glacial overburden - bedrock interface which is nearly everywhere confined by a thick clay till layer.

Groundwater samples were obtained from the freshwater aquifer for which ^{18}O concentration and electrical conductivity values were measured. These values were combined with ^{18}O, deuterium and tritium concentrations and electrical conductivity values from previous studies completed within the study area to determine the residence times and movement patterns of groundwater within the aquifer. In addition, a discrete state mixing cell model was developed utilizing ^{18}O concentration history to estimate groundwater flow rates in the freshwater aquifer.

The $\delta^{18}O$ distribution generally indicates that groundwater is progressively becoming older from areas of recharge, where relatively enriched $\delta^{18}O$ values (-9 to -11°/$_{oo}$) occur, toward areas of discharge, where depleted values (-16 to -18.5°/$_{oo}$) pres e. The physical hydrogeology, electrical conductivity and tritium values generally concur with the direction of groundwater implied by the $\delta^{18}O$ contours.

A one-dimensional discrete state mixing cell computer model, called the Simplified Discrete State Flow Model or SDSFM, was developed for this study from the equations of continuity. The model utilized ^{18}O concentration history to estimate groundwater flow rates in the freshwater aquifer. The approximate representative unit flow rates in the study area ranged from 0.3 m³/yr/m to 0.8 m³/yr/m.

Hydraulic conductivity (K) and average linear groundwater velocity ($\bar{v}$) was calculated by inputting the model derived flow rates into the appropriate version of the Darcy equation. The average 'K' values for the study area ranged from 6.04 x 10⁻⁶ m/s to 2.26 x 10⁻⁵ m/s. The average '$\bar{v}$' values ranged from 0.62 m/yr to 2.2 m/yr.

DEDICATION

The author would like to dedicate this work to her parents.

Mnogo hvala mama i tata.

ACKNOWLEDGEMENTS

TABLE OF CONTENTS

TABLE OF FIGURES

TABLE OF TABLES

1.0 INTRODUCTION

A thin layer of granular material and fractured bedrock occurring at the glacial drift and bedrock interface is pervasive throughout the area of the Lake Huron to Lake Erie corridor located within southwestern Ontario and southeastern Michigan. This unit referred to as the 'freshwater aquifer' is a regionally confined water supply aquifer that serves thousands of rural homes and farms. The study area on the Ontario side of the corridor consists of the counties of Essex, Kent and Lambton and on the Michigan side includes Macomb, Monroe, Oakland, St. Clair and Wayne Counties.

Injection of liquid industrial waste through deep well disposal practices into underlying porous and permeable bedrock units has been an object of concern with respect to the possibility of contamination of the freshwater aquifer. Another potential contaminant problem is the movement of landfill leachate from surface sources through the glacial drift to the freshwater aquifer. Deep well disposal of liquid industrial waste was commonly practiced in Lambton and Wayne Counties and landfilling is common throughout the study area. To estimate the rate of pollution contaminant movement through the freshwater aquifer it is necessary to quantify groundwater flow through this medium.

1.1 Objectives and Scope

The first objective of this study was to map the pattern of groundwater flow in the freshwater aquifer and identify areas of recharge and discharge by interpreting the following: physical hydrogeology of the study area, environmental isotope (oxygen-18 (^{18}O) and tritium) distributions and the distribution of electrical conductivity. For this study groundwater samples from domestic wells located on the Michigan side of the study area were analyzed for oxygen-18 concentration and electrical conductivity values. Additional oxygen-18, deuterium and tritium concentrations as well as electrical conductivity values were included from previous studies: MacGregor (1985) in Essex County; Scott (1986) in Lambton County; Baxter (1987) in Kent County; Crnokrak (1987) in northeastern Essex County; Erdmann (1987) in Macomb and St. Clair Counties; and Williams (1988) in the Sarnia area.

1

The second objective was to determine the age distribution of groundwater in the freshwater aquifer. This was accomplished by extrapolating the temporal variations of oxygen-18 concentration determined by Edwards (1990) from fossil wood-cellulose studies for Brampton, Ontario to the oxygen-18 concentration values of this study. The origin of the groundwater was also established through the qualitative interpretation of the environmental isotopes.

The final objective of this study was to estimate the rates of recharge, discharge and groundwater flow through the freshwater aquifer given the distribution of oxygen-18 concentrations acquired from the collective sample sites. A finite-state mixing cell model named the Simplified Discrete State Flow Model (SDSFM) was developed for this study to generate flow rates given the temporal variation of oxygen-18 from initial to final (present) values and the estimated volumes of aquifer sections.

1.2 Structure of Thesis

This thesis is divided into three main sections: the study area description, the natural tracer study and the computer model study. There are five chapters that make up the thesis framework. Chapter 1 discusses the objectives and scope of the study. The study area description of Chapter 2 covers the geology and hydrogeology of the region. Chapter 3 encompasses the theoretical considerations of environmental isotopes and electrical conductivity, the methods of their study and the results of the distribution of these natural tracers within the study area. Chapter 4 discusses the theory of the computer flow model developed for this investigation, the manner in which it is applied to the study area, the methods of determining the model parameter inputs and the interpretation of the resultant model flow rates. The last chapter summarizes the conclusions and recommendations of the natural tracer and computer model studies.

2.0 STUDY AREA

2.1 Location

The area of investigation in this study and its relative location is shown in Figure 1. The study area covers approximately 13000 km² of land surrounding the St. Clair River-Lake St. Clair-Detroit River corridor. This region forms part of the border between southwestern Ontario in Canada and southeastern Michigan in the United States. The river corridor is the connecting channel between Lake Huron and Lake Erie. The study area encompasses the counties of Essex, Kent and Lambton on the Ontario side and all of Wayne and Macomb Counties, the southern and eastern section of Oakland County, the northeastern corner of Monroe County and the southern two-thirds of St. Clair County on the Michigan side. The study area lies approximately between latitudes 41°53'N and 43°19'N and longitudes 81°47'W and 83°32'W.

2.2 Climate

The mid-latitude and continental position of the study area results in seasonal weather changes which are tempered by the moderating influence of the Great Lakes (Niedringhaus, 1966). The migration of the polar front results in the study area being influenced during the summer by hot, humid air from the south and cold, dry Arctic air during the winter (Sanderson, 1980).

Mean annual temperatures in the study area range from a high of 10.1°C (U.S. Department of Commerce, 1978) to a low of 8.3°C (Environment Canada, 1982). Precipitation annual means range from 101.7 cm/yr in Monroe County (Michigan Department of Agriculture Weather Service, 1981) to 76.2 cm/yr in Kent County (Brown et al., 1980). The prevailing wind direction is from the southwest for Wayne, Monroe, Oakland and Essex Counties and from the west for Kent, Lambton, St. Clair and Macomb Counties.

2.3 Topography

The surface topography of the study area consists of 2 distinct terrains: a hilly belt that crosses the northwestern section of the study area through Oakland County in a northeast-southwest trend, and a flat glacial lake plain that comprises the rest of
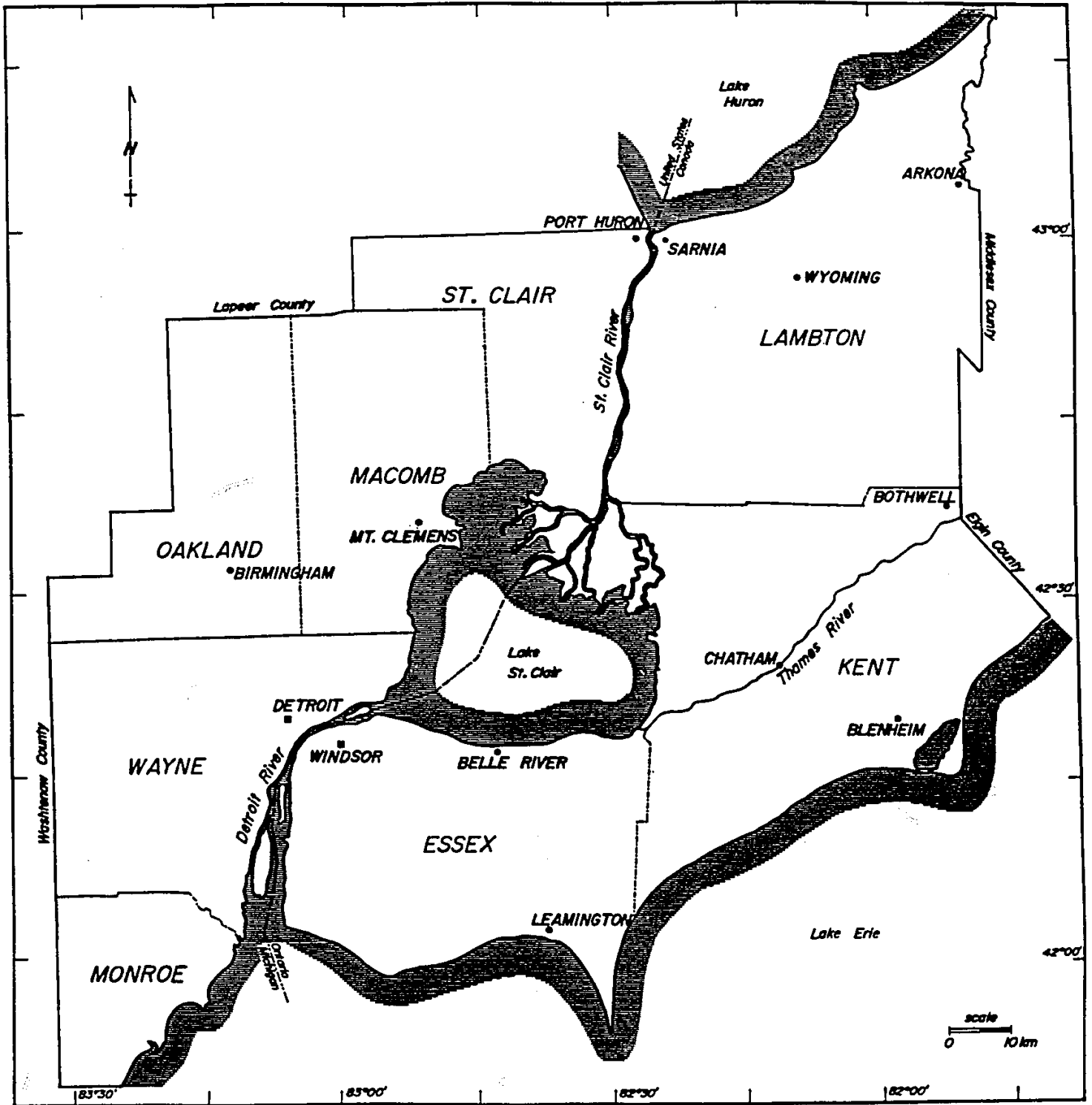
Figure 1   Location of the study area.

the study area (Figure 2). North of the hilly belt is a broad pitted flat plain, also trending northeast-southwest, which forms the topographic divide on the Michigan side and represents the highest elevation in the study area (340 m). The topography of the lake plain on the U.S. side slopes gently southeastward. On the Ontario side, the lake plain is even flatter, sloping westward and northward in Lambton County and westward in Kent County from a topographic high on the eastern edge of Lambton County (250 m) and a topographic high in the eastern corner of Kent County (220 m). In Essex County the plain gently slopes primarily northward and westward from a topographic high of 210 m near Leamington.

The bedrock topography (Figure 3) is similar to the surface topography. A bedrock divide on the Michigan side of the study area exists at the junction between the Thumb Upland and the Erie-Lowland physiographic regions. This divide, which trends northeast-southwest, runs just across the northwestern edge of the study area at approximately the 230 m to 260 m bedrock contour levels. This divide approximately coincides with the topographic divide. The surface topographic highs on the Ontario side correspond with underlying bedrock highs. The slope of the surface topography generally reflects the bedrock surface slope.

2.4 Geology

The study area is underlain by Paleozoic sedimentary rock sequences composed of shale, limestone, dolomite and sandstone which are mantled by unconsolidated Quaternary deposits. The Quaternary deposits primarily consist of glacial drift in thicknesses varying from a few metres to over 100 m. Localized alluvium deposits of Recent age, in turn, are scattered on top of the glacial deposits. Detailed information about the bedrock geology of southwestern Ontario and southeastern Michigan can be found in the following sources: Lilienthal (1978), Windsor and Sanford (1972), Brigham (1971), Dorr and Eschman (1970), Sanford (1968), Beards (1967), Sanford and Brady (1955), and Mozola (1953, 1969, 1970). Detailed summaries of the Quaternary geology within the study area are provided by: Morris (1988), Quigley and Ogunbedejo (1976), Mozola (1953, 1969, 1970), and Dreimanis (1961).
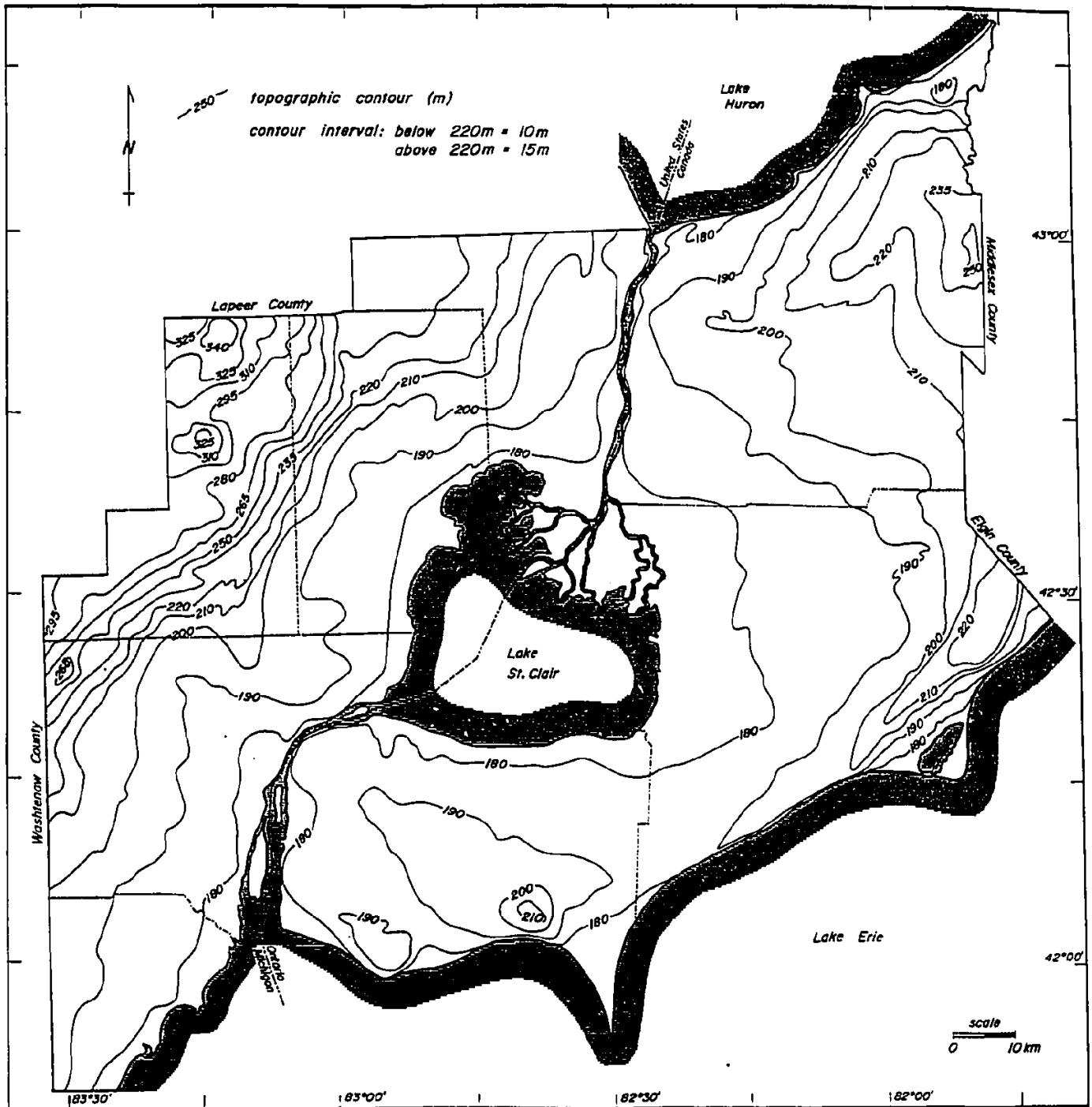
Figure 2   Topography of the study area (after Department of Energy, Mines and Resources, 1973, 1978; U.S. Army Map Service, 1953,
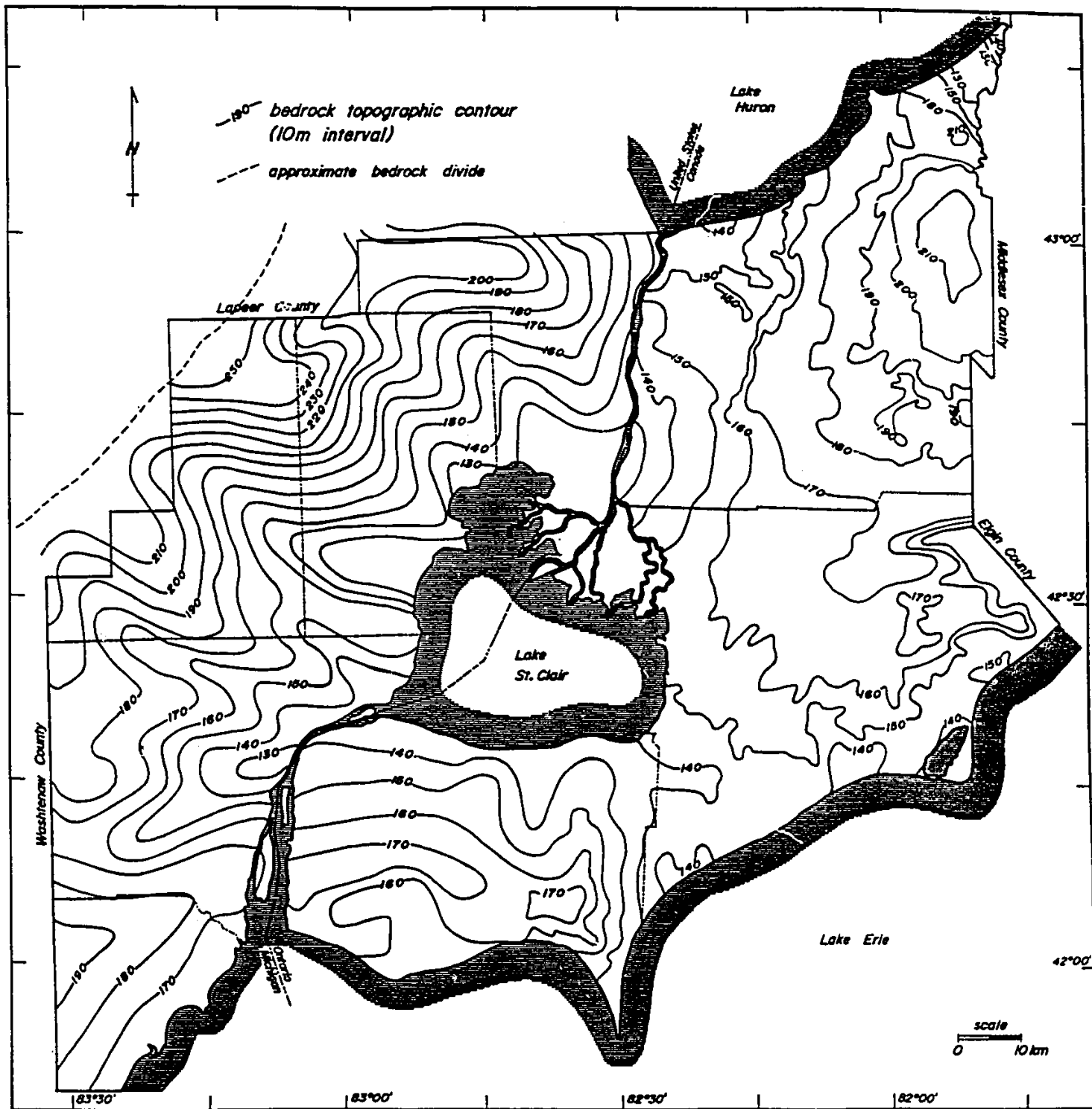
Figure 3 Bedrock topography of the study area (after Cooper, 1976(a & b); Cooper, 1978(a, b & c); Fitzgerald et al., 1979; Sado and Faught, 1981(c & d)).

7

2.4.1 Bedrock Geology

2.4.1.1 Structural Setting

The majority of the study area lies on the southeastern margin of the Michigan Basin (Figure 4). The Algonquin Arch enters the study area from the northeast and the Findlay Arch enters from the southwest. These structural highs form a structural flexure which separates the Michigan Basin to the northwest and the Appalachian Basin to the southeast. The Chatham Sag, formed by the plunge of the two arches, cuts through the study area in a northwest-southeast trend and is centered in Kent County.

The position of the study area on the Michigan Basin gives the outcropping edges of the study area's rock strata on the Michigan side a strong northeast-southwest trend and establishes a slight dip to the northwest at a rate of 6-9 m/km. Local variations in dip are a result of salt bed dissolution in deeper Silurian strata creating collapse structures and/or sediment draping over reef structures (Bringham, 1971). Physiographically the study area bedrock surface occupies the rock lowland of the Erie-Huron plain and the southeastern slope of the Thumb Upland rock surface (Mozola, 1953, 1969, 1970).

2.4.1.2 Stratigraphy

The study area is underlain by two general lithologic sequences: an evaporite-carbonate sequence which includes the Salina Formation, Bass Islands Dolomite, Detroit River Group, Dundee Formation and the Hamilton/Traverse Group, and a clastic sequence which includes the Kettle Point/Antrim Shale, Bedford-Berea-Sunbury/Port Huron Formation and Coldwater Shale (Figure 5).

The sequence of the rock formations ranges from late Silurian to Early Mississippian. Figure 6 is a partial stratigraphic column which displays all the rock groups and formations which outcrop or come into direct contact with the overlying glacial drift. The oldest formation which subcrops in the study area is the late Silurian Salina formation. It subcrops in the extreme southwestern corner of the study area. The Coldwater Formation is the youngest rock and subcrops in the northwestern edge of the
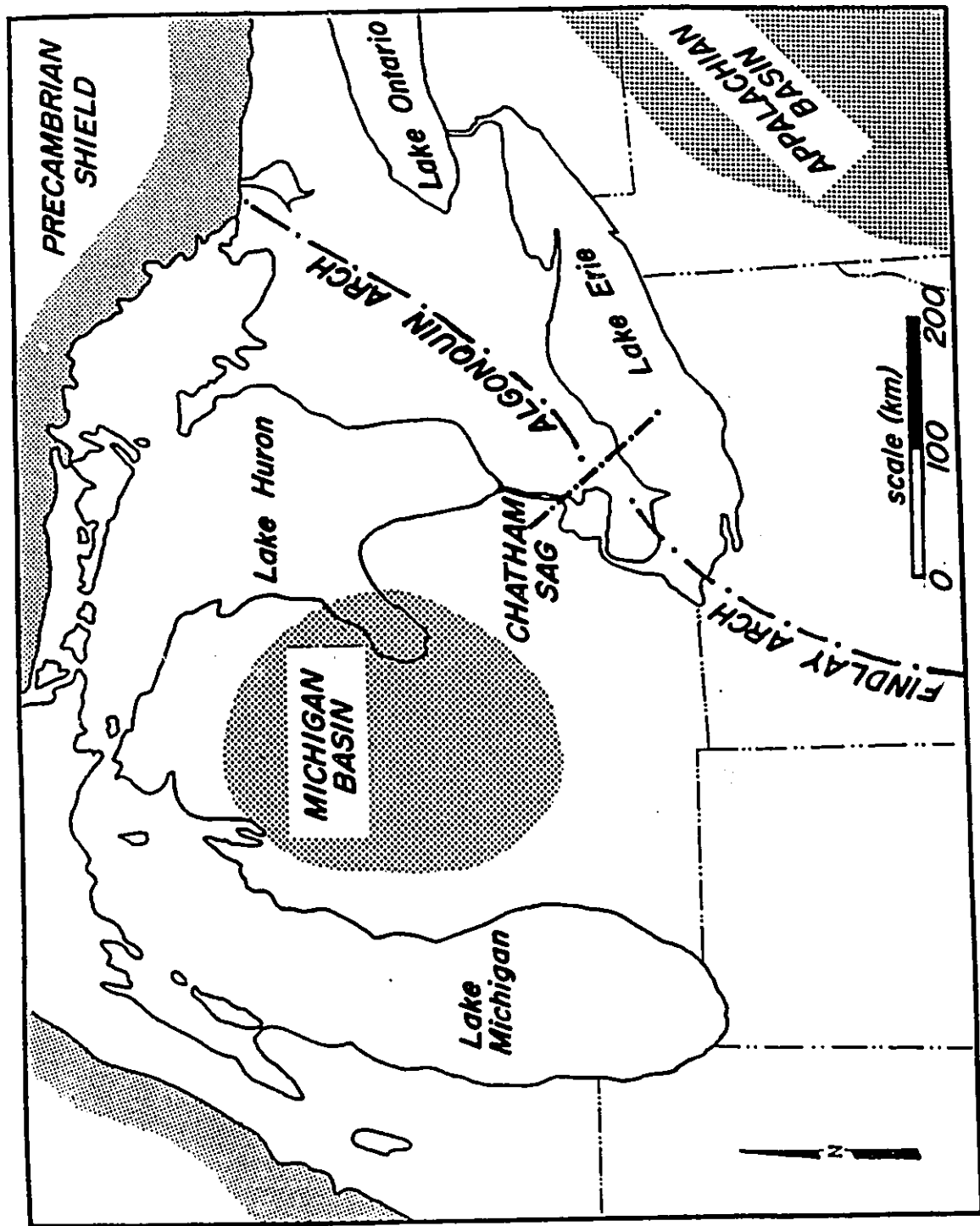
Figure 4   Major structural features of the study area.

Coldwater Formation
-shale

Port Huron/Berea-Bedford-
Sunbury Formation
-shale, sandstone

Kettle Point / Antrim
Formation -black shale

Hamilton / Traverse Group
-shale, shaley limestone

Dundee Formation
-limestone

Detroit River Group
-limestone, dolomite

Sylvania Formation
-sandstone

Bass Islands Formation
-dolomite

Salina Group
-shaley dolomite

N

Lake
Huron

United States
Canada

Lapeer County

Middlesex County

43°00'

Elgin County

42°30'

Lake
St. Clair

Washtenaw County

Lake Erie

42°30'

42°00'

Ontario
Michigan

scale
0   10 km

83°30'   83°00'   82°30'   82°00'

Figure 5   Bedrock geology of the study area (after WMU, 1981a; Sanford, 1969).

10

SOUTHWESTERN ONTARIO				SOUTHEASTERN MICHIGAN			
PERIOD	ERA	GROUP/FM	LITHOLOGY	PERIOD	ERA	GROUP/FM	LITHOLOGY
Devonian and Mississippian		Drift	Clay,Gravel	Mississippian	Early (lower)	Drift	Clays,Gravel
		Port Lambton	Sandstone Shale			Coldwater	Shale
						Sunbury Berea Bedford	Shale Sandstone Shale
Devonian	Middle	Kettle Point	Shale	Devonian	UPPER	Antrim	Shale
		Hamilton	Limestone Shaly Limestone		Middle	Traverse Group	Shale Limestone Shale
		Dundee	Limestone			Dundee	Limestone
		Detroit River Group: Lucas Amherstburg	Dolomite Cherty Limestone			Detroit River Group: Lucas Amherstburg Slyvania	Dolomite Dolomite,Limestone Sandstone
	LOWER	Bois Blanc	Cherty Limestone		LOWER	Bois Blanc	Cherty Carbonates
Silurian	Upper	Bass Island	Dolomite with Shaly Interbeds	Silurian	Upper	Bass Island	Dolomite
		Salina — Unit G	Shaly Dolomite			Salina — Unit G	Shaly Dolomite
		Salina — Unit F	Shaly Dolomite, Anhydrite Salt			Salina — Unit F	Evaporite
		Salina — Unit E	Dolomite with Shaly Interbeds			Salina — Unit E	Dolomite
		Salina — Unit D	Anhydrite Salt			Salina — Unit D	Evaporite
		Salina — Unit C	Shale, Dolomitic Shale			Salina — Unit C	Shaly Dolomite
		Salina — Unit B	Anhydrite Salt			Salina — Unit B	Salt, Shale, Dolomite
		Salina — Unit A-2	Dolomite, Salt, Anhydrite			Salina — Unit A-2	Limestone, Dolomite
		Salina — Unit A-1	Limestone, Dolomite, Anhydrite			Salina — Unit A-1	Limestone, Dolomite, Anhydrite

Figure 6   Partial Stratigraphic column for Southwestern Ontario and Southeastern Michigan (modified after Lilienthal, 1978; Vandenburg et al., 1977; Sandford and Brady, 1955).

study area. Because of the large number of rock formations/groups which directly underlie the drift in the study area, the lithology of each bedrock unit is described in Appendix I.

2.4.2 Quaternary Geology

The unconsolidated sediments which overlie the bedrock were deposited during the late Wisconsinan stage of glaciation (Dreimanis and Karrow, 1972). These deposits may be divided into two distinct areas: an extensive clay till plain of little relief which covers almost the entire study area and an interlobate belt made up of moraines, till plains and outwash deposits (Figure 7). The position and succession of glacial features in the Michigan study area are related to the advance and withdrawal of the Erie-Huron ice lobe (Mozola, 1969).

2.4.2.1 Clay Till Plain

The clay plain is composed of lake beds of clay and sand overlying glacial till which are interrupted by morainic ridges. The texture of the clayey till is fairly uniform throughout the clay plain; typically composed of 40 to 60% clay, 30 to 40% silt, 5 to 10 % sand, and usually less than 5% gravel and fairly uniform throughout its depth (Soderman and Kim, 1970).

The fluctuation of the Wisconsinan glacier ice front has resulted in a heterogeneous assortment of sediments with a wide range of porosities and permeabilities within the till. Associated with the clayey till are complexly distributed sand and gravel deposits, known as lenses, which usually interfinger with and grade into finer sediments.

The glacial till is nearly everywhere thinly veneered by clayey, glaciolacustrine deposits on the Michigan side and on a tract of land east of Lake St. Clair and smaller localized areas on the Ontario side.

There is a large plain of glaciolacustrine fine sands and silts in Kent County which formed in the Thames River delta of glacial Lake Warren and is referred to as the Bothwell Sand Plain (Chapman and Putnam, 1984). It forms a layer of sand over the clayey till floor which ranges in thickness from approximately 1 to over 9 m at Bothwell (OWRC, 1970). The present day Thames River flows through the center of the sand plain and divides it into two. Similar sand plains of fine sands and silts have been

Legend:

- ☐ TIII plains (ground moraine)
- Clay plains
- Sand plains
- Moraines
- Waterlaid moraines
- ☐ Outwash and glacial channels
- Boulder belts
- Peat and muck
- ⟋ Beaches

Lake Huron

United States / Canada

Lake St. Clair

Lake Erie

Lapeer County

Washtenaw County

Middlesex County

Elgin County

Ontario — Michigan

scale
0   10 km

43°00'
42°30'
42°00'

83°30'   83°00'   82°30'   82°00'
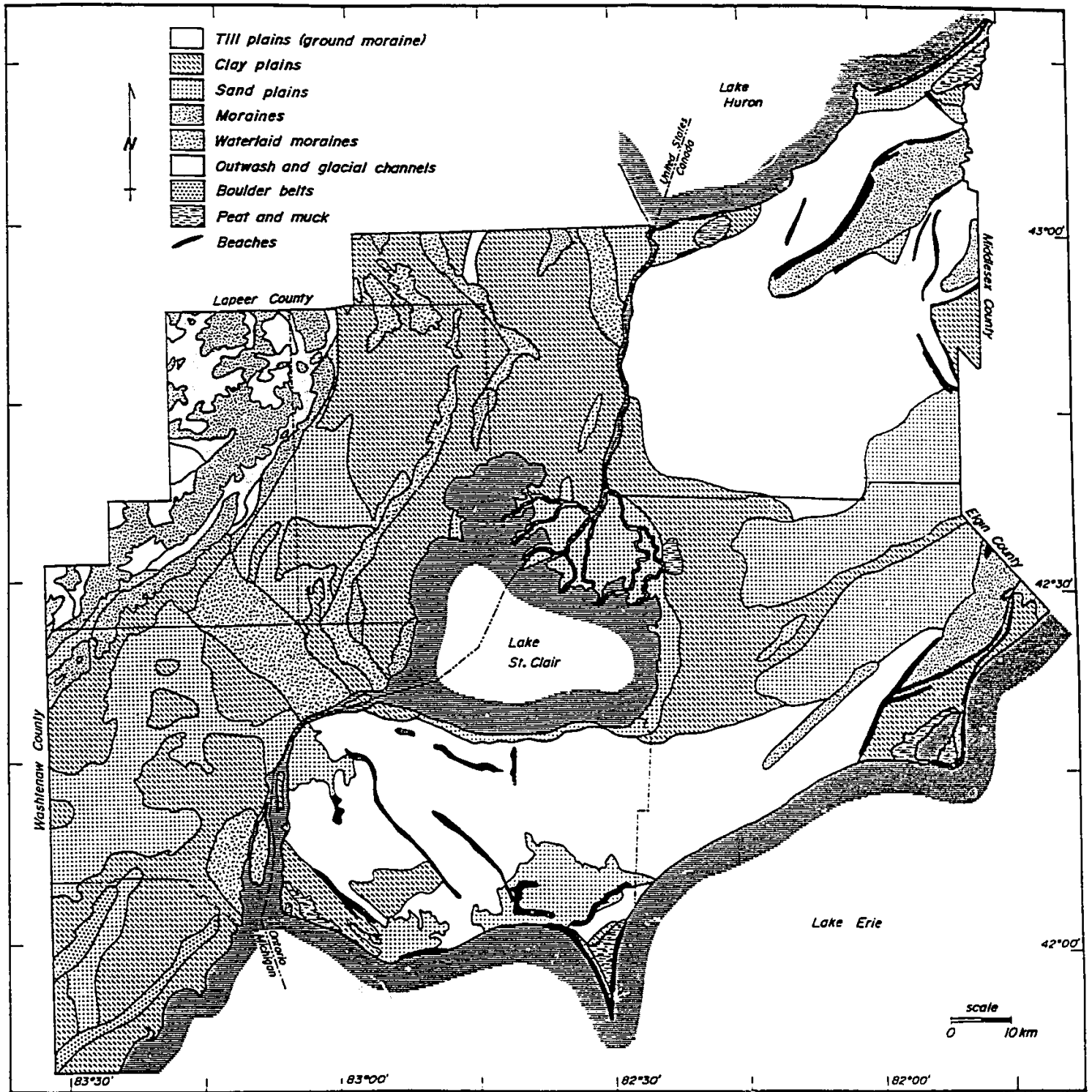
Figure 7   Quaternary geology of the study area (after WMU, 1981a; Chapman and Putnam, 1984).

formed by ancient glacial deltas at the mouth of the St. Clair River, on the border between Oakland and Macomb Counties, and through the center of Wayne County in the form of an elongated sand plain which trends northeast-southwest.

The continuity of the till plain is interrupted by a series of moraines. Within the Ontario area there are five major morainic hills: the Leamington, Blenheim, Charing Cross, Seaforth, and Wyoming Moraines. Within the Michigan till plain area there are six major morainic ridges: the Birmingham, Detroit, Mt. Clemens, Emmett, Grosse Isle, and Port Huron Moraines. The location of these moraines are shown in Figure 8 and a brief description of each moraine is given in Appendix I.

2.4.2.2 Interlobate Belt

This area is a northeast-southwest aligned series of morainic ridges and outwash plains which runs through the northwest corners of Macomb and Wayne Counties and most of Oakland County save the southeast corner (Figure 7). It was formed when the Saginaw and Erie-Huron lobes of the Wisconsinan glacier, which had approached each other and joined, withdrew and separated. The resulting meltwaters from both lobes deposited the outwash sands and gravels found within the study area (Mozola, 1953). The outwash plains separate the two morainic hill belts subsequently deposited by the retreating ice lobes. Only the southern belt of moraines formed by the Erie-Huron lobe are found within the study area. The bedrock and topographic divides previously mentioned are roughly coincident with this interlobate area.

The outwash plains form the largest area of stratified drift within the study area and reach thicknesses of up to 110 m with sand and gravel predominating throughout their depth. Clays are generally subordinate except where morainic hills rise above the plain level. Parts of the outwash plain probably extend beneath the morainic belts since the ice lobe fronts fluctuated during the glacial period (Mozola, 1953). The outwash deposits may, in turn, be underlain by till (Mozola, 1969).

The morainic ridges and till plains found within the interlobate area are composed largely of till deposits, within which are sand and gravel layers which probably represent outwash material deposited earlier in the area's glacial history. During the development of the outwash area the first moraine deposited by the Erie-Huron lobe was the Fort Wayne Moraine (Figure 8). Its junction with the outwash plains marks the area of
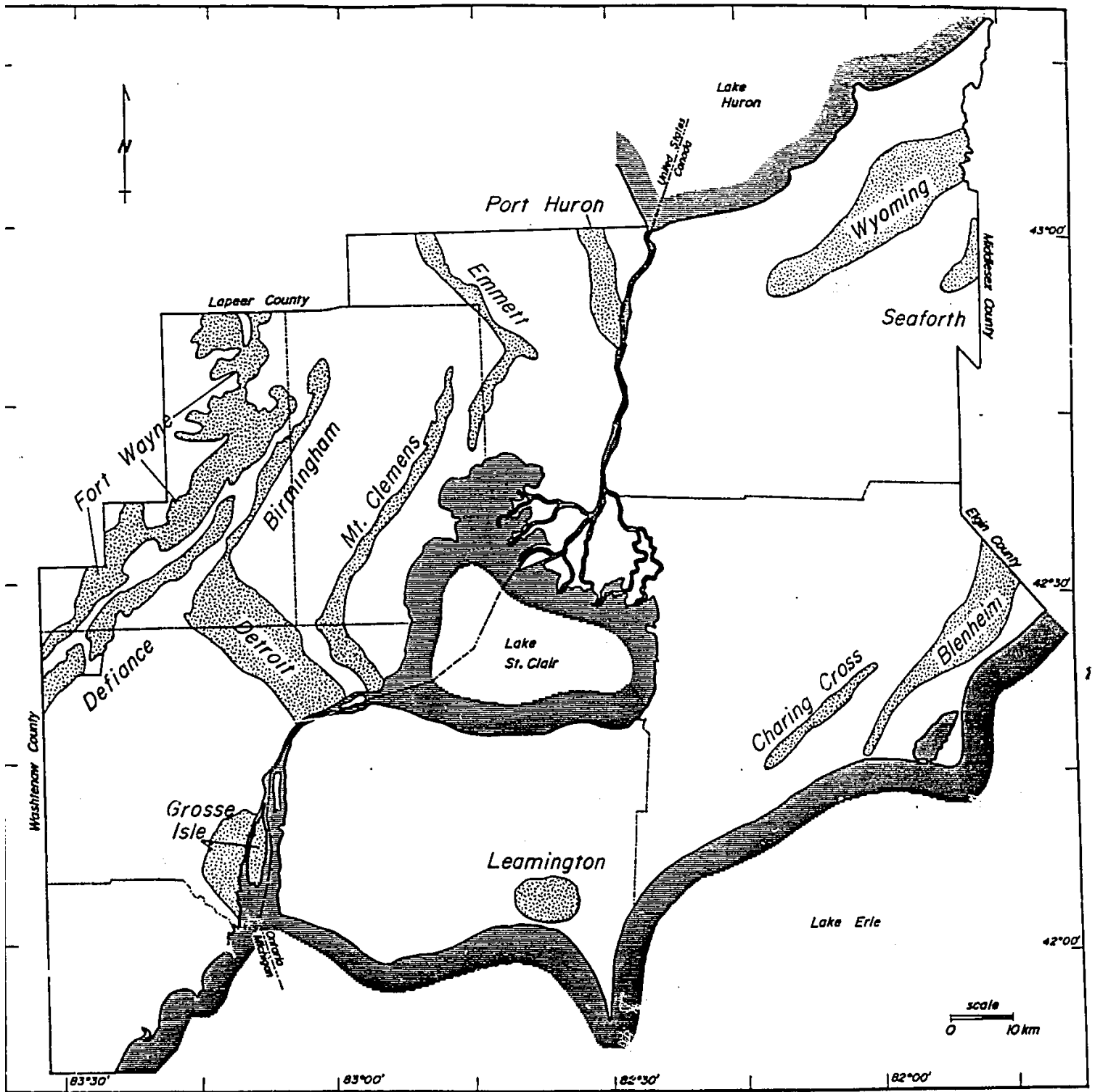
Figure 8   Location of moraines in the study area (after WMU, 1981a).

greatest drift thickness within the study area from 90 to 119 m. The next known position of the Erie-Huron lobe is marked by the Defiance Moraine, a narrow hilly belt having a maximum width of 5 km. The till plain moraines, mentioned earlier, have less prominent relief than the older interlobate moraines.

2.4.2.3 Drift Thickness

Though the till plain is essentially flat, the glacial overburden is not uniform in thickness. The drift thickness strongly reflects the topography of the underlying bedrock surface. Most bedrock topographic highs are associated with relatively thin drift, except where they form divides between adjacent ice lobes, and bedrock valleys are associated with relatively thick drift, therefore the bedrock slope is not apparent on the surface (Rieck and Winters, 1977). In general, the drift thickness map (Figure 9) indicates that till and glacial lake plains are dominantly areas of thin drift relative to moraines which correspond to areas of thicker drift.

Aside from the interlobate region, the thickness of drift in the study area changes gradually. The thickness of the till plain on the Michigan side decreases steadily towards the southeast from a maximum of about 50 m in Wayne and Oakland Counties to a minimum of 5 m in eastern Monroe County. Slightly deeper overburden within the till plain occurs within St. Clair County where the drift thickness decreases eastward from 70 to 40 m and Macomb County where the maximum thickness of 65 m decreases to a minimum of about 35 m along Lake St. Clair. A localized area of thicker drift occurs in the vicinity of the Detroit Moraine in eastern Wayne County where thicknesses reach 65 m.

On the Ontario side an area of thin drift, approximately 15 m thick, cuts diagonally across Kent and Lambton Counties north of the Thames River. From this area the overburden gradually thickens northwestward toward the St. Clair River and Lake Huron (maximum 50 m) and southeastward toward Lake Erie to a maximum of about 50 m. The drift within the Blenheim Moraine thickens to 75 m. Within Essex County drift thickness increases from west to east ranging from 5 to 45 m. Just east of the Leamington Moraine the overburden thickness decreases to 15 m.

drift thickness contour (m)

N

Lake Huron

United States / Canada

Lapeer County

Middlesex County

43°00'

Washtenaw County

Elgin County

Lake St. Clair

42°30'

Ontario — Michigan

Lake Erie

42°00'

scale
0   10 km

83°30'   83°00'   82°30'   82°00'
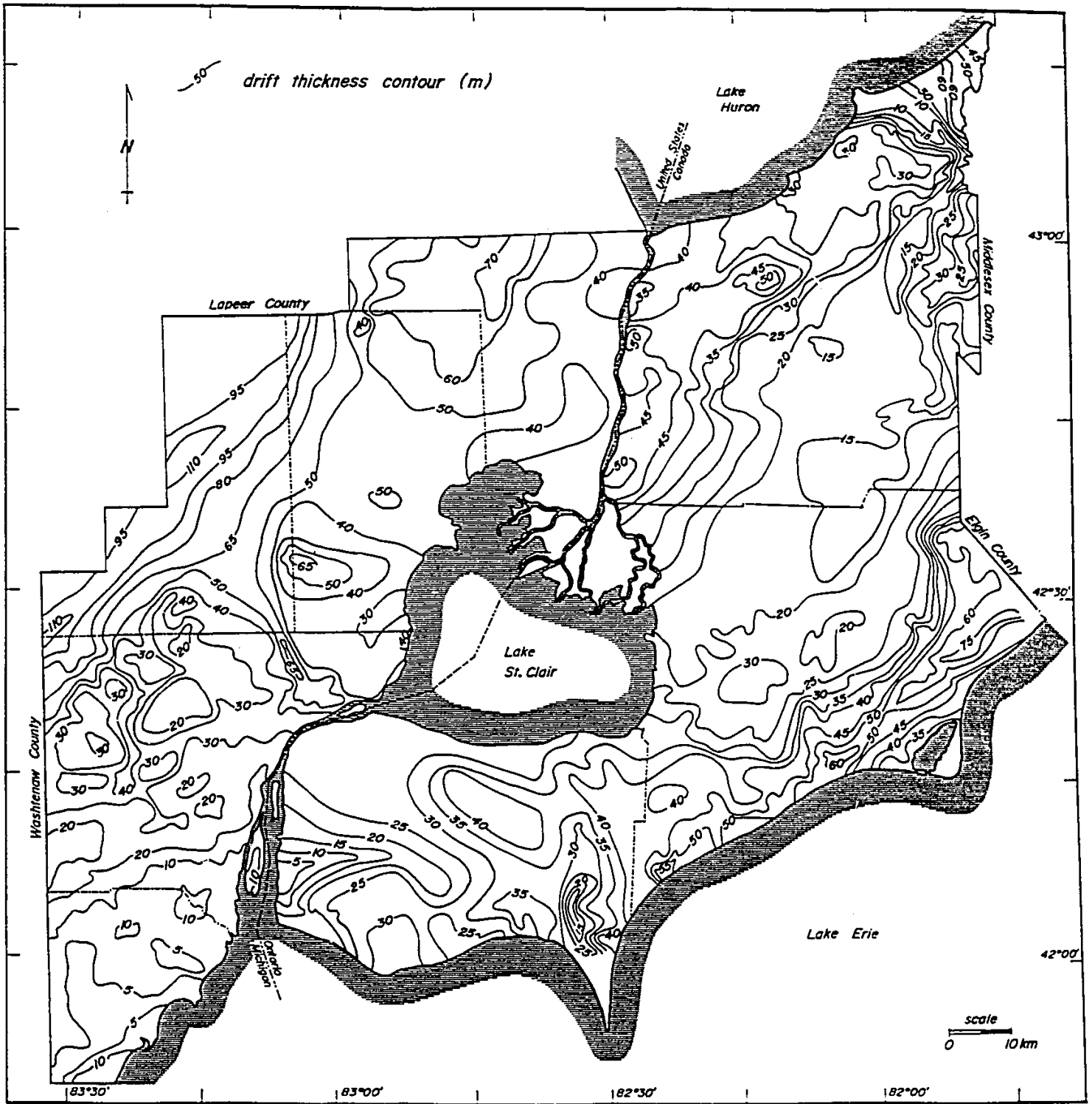
Figure 9   Isopach map of overburden in the study area (after Cooper, 1981(a & b); Cooper and Nicks, 1981(a, b & c); Fitzgerald et al., 1979b; Sado and Faught, 1981(a & b); Vagners et al., 1973(a & b)).

The interlobate region of Oakland County represents the area of thickest overburden. The outwash plains reach thicknesses in excess of 110 m. The average thickness of drift for the entire interlobate region within the study area is about 75 m.

2.5 Hydrogeology

2.5.1 Hydrostratigraphic Units

The study area may be divided into three distinct hydrostratigraphic units: the glacial overburden, the glacial overburden-bedrock interface, and bedrock. A brief discussion of each unit follows. For more detailed discussions of the hydrostratigraphic units the reader may refer to: WMU (1981a, 1981b), Mozola (1953, 1969, 1970), Intera (1989), GTC (1986), and OWRC (1969, 1970, 1971).

2.5.1.1 Glacial Overburden Unit

In areas where the drift is principally composed of glacial lake deposits or clayey till, the material is considered to be an aquitard. This glacial drift is composed of fine-grained clays and silts too impermeable to be productive aquifers. Most of the study area, excluding the northwestern two-thirds of Oakland and the northwestern corners of Wayne and Macomb Counties, is underlain by this material.

At depth, the lake/till plain aquitard may contain thin interbedded aquifers (WMU, 1981a, 1981b). The frequency of occurrence, thickness and extent of these buried aquifers decreases from the junction of the glacial lake plain with the interlobate belt toward the southeast (Mozola 1953, 1969). These confined aquifers also occur with greater frequency within the study area's moraines. Profiles drawn across the glacial lake plain show that most of the individual sand and gravel zones cannot be correlated between test holes (Mozola, 1953). This suggests that these confined aquifers are discrete, discontinuous, of small areal extent and with limited storage capacity and recharge capability (WMU, 1981a, 1981b; OWRC, 1970, 1971).

There are several localized areas of drift composed of glaciofluvial/glaciolacustrine sands and gravels at or near the surface which form unconfined aquifers. At depth this drift generally consists of interbedded aquifers, aquicludes and aquitards (WMU, 1981a, 1981b). The nature of these deposits allows water to infiltrate down directly or along a

tortuous path between interconnected buried aquifers until bedrock is eventually reached. The most prominent aquifer of this type in areal extent and depth is located in the outwash deposits area of the interlobate region in the northwestern corner of the study area. Within these deposits several well records show that clays are almost absent throughout their profile. The Bothwell Sand Plain of eastern Kent and southeastern Lambton Counties is another areally extensive but thinly deposited unconfined aquifer. The Leamington Moraine forms a smaller unconfined aquifer with buried inter-till aquifers at depth.

2.5.1.2 Glacial Overburden - Bedrock Interface Unit

The interface which separates the glacial unit and the bedrock unit is routinely referred to as the "freshwater aquifer". This interface unit is composed of a nearly continuous, thin and uneven layer of sand, gravel, and broken and fractured bedrock (Gillespie and Dumouchelle, 1989; Intera, 1989; URM, 1984). This unit is usually 1.5 to 6 m thick (Gillespie and Dumouchelle, 1989) but, locally, it may thicken considerably within buried bedrock valleys (Intera, 1987, 1989). This aquifer is confined where it is situated between shaley bedrock (Kettle Point/Antrim shales and younger) and overburden composed of clayey till. When considering the areal extent of the various sources of groundwater, the most significant source is the freshwater aquifer.

2.5.1.3 Bedrock Unit

The study area is predominantly underlain by impermeable shaley bedrock (Figure 5). South of the mouth of the Detroit River, carbonate rocks are prevalent, subcropping in Essex, southeastern Wayne and Macomb Counties. The well yields from the carbonate bedrock aquifer in this area is quite good especially where solution has taken place (GLBC, 1975). Wells that penetrate deep into the rock are common in southern Essex County where the Detroit River Group subcrops (OWRC, 1971). Deeper rock wells are also common within Monroe and the extreme southern section of Wayne Counties where the drift is less than 20 m thick and fresh water supplies can be obtained from the carbonate rocks (Gillespie and Dumouchelle, 1989).

2.5.2 Groundwater Flow

The following discussion relates to flow within the freshwater aquifer.

2.5.2.1 Direction of Flow

The hydraulic head distribution of the freshwater aquifer within the study area was determined from static water levels obtained from water well logs. Figure 10 is the resultant potentiometric surface. The resultant horizontal hydraulic gradients range from 0.0002 to 0.01 and these gradients generally decrease in the direction of the connecting channels.

The correlation of both bedrock topography (Figure 3) and surface topography (Figure 2) to the direction of groundwater flow is readily evident when they are compared to the direction of groundwater flow implied by the potentiometric surface map. The bedrock topographic, topographic, and potentiometric surface contours essentially mirror each other; this implies that the topography and the potentiometric surface follow the trend of the bedrock surface. These contoured surfaces indicate that flow within the freshwater aquifer is concentrated toward the centre of the study area (Lake St. Clair).

Several regional flow systems can be identified from the countenance of the hydraulic head distribution. Figure 10 indicates that on the Michigan side, flow is unvaryingly southeast toward Lake St. Clair and the St. Clair and Detroit Rivers. In addition, the hydraulic gradient steadily decreases in the downgradient direction. Another flow system seems to originate near the eastern edge of Lambton County. Flow is primarily to the west and southwest toward the St. Clair River and Lake St. Clair with the southern edge of flow eventually reaching Lake Erie. Flow in the northeastern corner of Lambton County indicates that a small component of the flow is north toward Lake Huron. A third flow system is located within Essex County. Groundwater flow radiates out in all directions from an area in the southeastern portion of the county toward Lake St. Clair, the Detroit River and Lake Erie. The horizontal hydraulic gradient is much steeper south of this area (i=0.0043) than in the other downgradient directions (i_{ave}=0.0011).
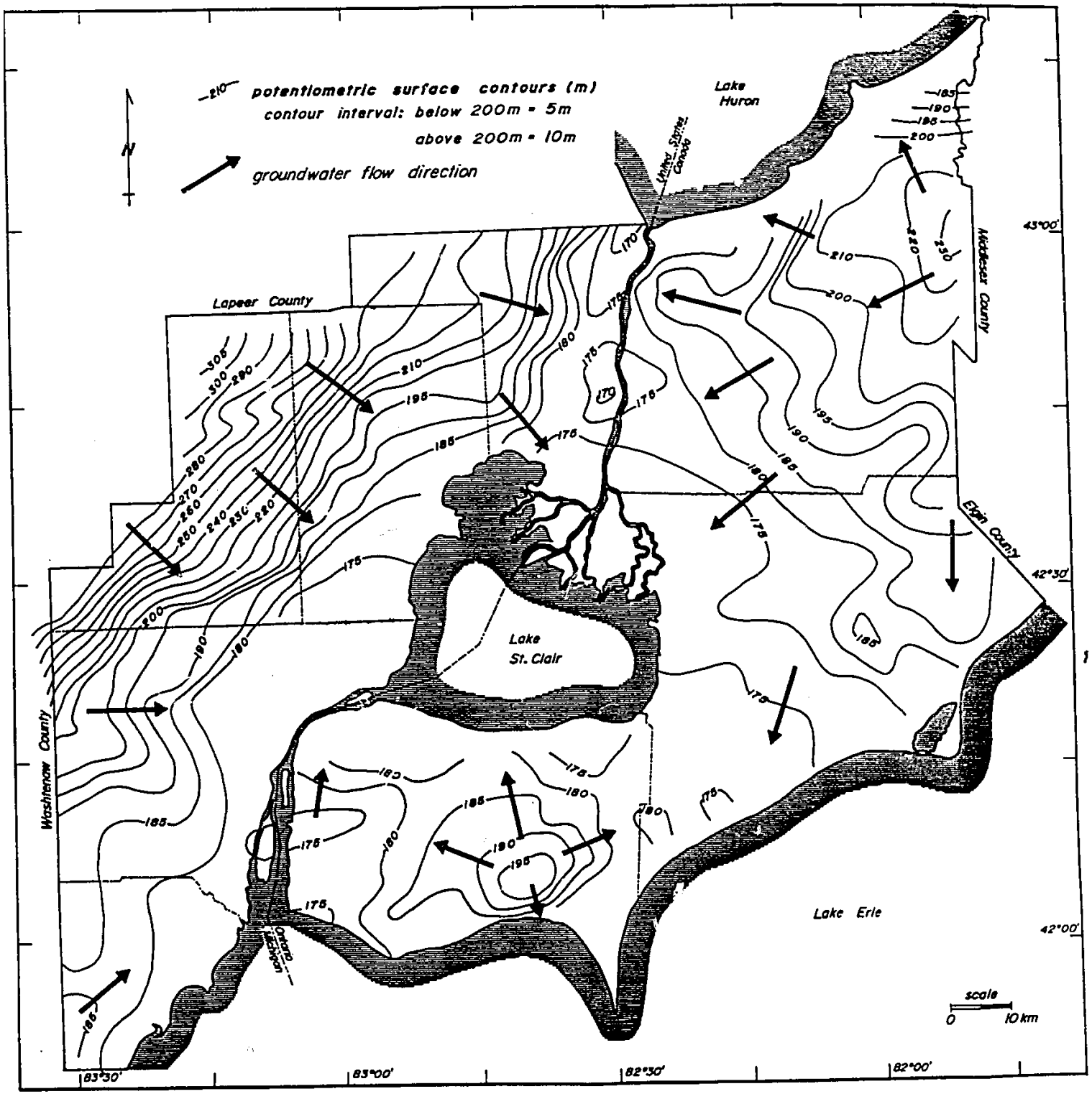
Figure 10   Potentiometric surface of the freshwater aquifer.

The approximate location of the bedrock divide in the northwestern portion of the study area is shown in Figure 3. Comparing this divide with the positions of the groundwater and topographic divides estimated from Figures 10 and 2, it is apparent that the three divides are coincidental and that flow is to the southeast within the Michigan study area. Similarly, the positions of the circular piezometric highs near the northeastern edge of Lambton County and in southern Essex County are almost exactly concurrent with the positions of their respective bedrock and topographic highs.

2.5.2.2 Recharge and Discharge Areas

Interpretation of the surficial geology, bedrock topography and piezometric surface of the freshwater aquifer has resulted in the identification of five areas of recharge within the study area: the interlobate region in the northwestern portion of the study area, the topographic high on the eastern edge of Lambton County, the Thames River-Bothwell Sand Plain area of northeastern Kent County, the Leamington Moraine in southern Essex County and the area of thin overburden (<5 m thick) in Monroe County.

The bedrock, topographic and groundwater divides of the Michigan side are located within the area of outwash deposits of the interlobate region. This area constitutes the major intake area of the Michigan study area (Mozola, 1953). Coarse grained sands and gravels throughout the vertical drift column provides a relatively uninhibited infiltration path for precipitation which ultimately reaches the freshwater aquifer.

Recharge into the aquifer located on the bedrock high near Lambton County's eastern border is possible through the sandy beach deposits which edge the Wyoming and Seaforth Moraines and the buried stratified sands and gravels associated with these recessional moraines.

The Bothwell Sand Plain occurs in an area of relatively thin drift which, according to water well records, contains inter-till aquifers. The Thames River channel may cut through these inter-till aquifers and be hydraulically connected to them. These aquifers may also be connected to the freshwater aquifer since the river flows through thin drift approximately 15 m thick which could allow infiltration of surface water from the river into the inter-till aquifers and eventually into the freshwater aquifer.

The Leamington Moraine, located in southeastern Essex County, forms a topographic high and is situated above a bedrock high. It is comprised of surficial deposits of sand and gravel which form an unconfined aquifer and confined inter-till sand/gravel units.

The southern part of the Monroe County study area contains an area of very thin drift which is less than 5 m thick. A conduit for precipitation may be allowed wherever vertical fractures occur through this thin till. Inter-till sand lenses, although infrequent in this part of the glacial till plain, can also provide an infiltration path for recharge water, especially if the till plain is mantled with surficial sands.

Groundwater flow within the freshwater aquifer occurs by the movement of water from areas of high potential to low across the potentiometric surface. Figure 10 indicates that all of the surface water bodies adjacent to the study area are discharge areas: Lake St. Clair, which is the most common discharge area, Lake Huron, Lake Erie, the St. Clair River and the Detroit River.

2.5.3 Groundwater Chemistry

Water in the bedrock is usually more ionized than that recovered from the glacial drift, consequently dissolved solids tend to be highest in samples from bedrock wells. The degree of ionization increases with depth of penetration into bedrock (WMU, 1981). Glacial drift water tends to dilute the ionized bedrock water thereby increasing its own ionic content.

Mineral content and type may vary greatly within the freshwater bedrock aquifer depending upon which rock unit the ionized water originated from. For example, hydrogen sulfide is a common water quality problem throughout the lake plain and till plain sections, particularly where the freshwater aquifer overlies the Kettle Point/Antrim shale formation (Mozola, 1953). High chloride content is also a common problem in the freshwater aquifer within the extent of the till and lake plains. Vandenburg et al. (1977) has shown that the freshwater aquifer within Lambton County is dominated by chloride water. The salt water in these sediments are attributed to the upward seepage of brine from deeply buried bedrock formations (Mozola, 1953). Potable supplies are possible from most rock formations within their subcrop area provided the wells are of shallow penetration into rock (Mozola, 1969).

2.5.4 Previous Studies

Hydraulic conductivity values which averaged 5×10^{-6} m/s were reported for the freshwater aquifer in the Sarnia area by Intera (1989) and were determined by a variety of field methods (slug, withdrawal, recovery and pump tests). The average hydraulic gradient across the Sarnia area was 0.0016. The average groundwater flux per unit aquifer width (assuming aquifer thickness to be 2 m) was 0.5 m^3/yr, using Darcy's Law and the field measured 'K', and 0.57 m^3/yr using a finite difference flow model. Assuming an average porosity of 0.3 the average fluid velocity through the freshwater aquifer was calculated to be 2.7×10^{-8} m/s.

A two-dimensional finite difference computer simulation of flow through the freshwater aquifer utilizing the distribution of ^{18}O was completed by Hyde (1987) for Lambton County and Pettapiece (1988) for Kent County. The Lambton County computer simulation results indicated an average hydraulic conductivity value of 5.0×10^{-5} m/s and a porosity of 35% for the freshwater aquifer. The Kent County computer simulation results indicated an average hydraulic conductivity value of 1×10^{-4} m/s and a porosity of 34%.

Values of 'K' for the freshwater aquifer differ greatly from those of the overlying glacial till plain which essentially confines the permeable aquifer from above. Within the clay till plain of the study area, Desaulniers (1980) calculated values of field hydraulic conductivity that ranged between 1.0×10^{-10} m/s and 5.2×10^{-10} m/s with a geometric mean of 1.7×10^{-10} m/s. Similar laboratory values of 'K' were determined by Desaulniers et al. (1981) with values ranging between 8.3×10^{-11} m/s and 9.3×10^{-10} m/s. Orpwood's (1984) study of the clay in Essex County resulted in similar values of hydraulic conductivity where field 'K' ranged from 3×10^{-10} to 8×10^{-10} m/s for clayey till greater than 6 m deep.

2.5.5 Groundwater Use

The use of groundwater in the study area is dependent upon several factors: the type of overburden, thickness of the overburden and the availability of surface water sources.

Groundwater is generally plentiful within the stratified, coarse grained interlobate region, while low yields, only adequate for domestic supplies, are generally expected from the till plain. Where the till and glacial lake plains are exposed at the surface, groundwater supplies are generally obtained from the freshwater aquifer (Ontario Ministry of the Environment, 1981).

The Ontario side of the study area in general has a higher percentage of bedrock wells than the Michigan side due to the greater prevalence of the clayey till plain which cannot provide an adequate supply of water. In Lambton County, 90% of the water wells obtain water from the freshwater aquifer (URM, 1984). Within Essex County 80% of the recorded wells terminated in the bedrock (MOE, 1981), with many of these being freshwater aquifer wells (MOE, 1984). Water supplies may also be obtained from buried sand and/or gravel units within these plains. These confined aquifers occur infrequently within the Ontario study area while the volume and frequency of occurrence of these deposits in Michigan increase northwestward toward the interlobate region (Mozola 1953, 1969).

The frequency of wells completed in the bedrock generally diminishes with increasing drift thickness. Most areas within the Michigan study area that have drift thicknesses in excess of 30 m have less than 10% wells completed in the bedrock (WMU, 1981a, 1981b) such as in the Oakland, Macomb and the larger part of the St. Clair County study area. Because of the relatively thin glacial drift within Monroe County and the southern part of Wayne County, bedrock wells in this section make up greater than 10% of the total number of water wells. Monroe County alone has about 90% of its wells completed in bedrock (Mozola, 1970).

The Detroit Metropolitan water system serves southeast Oakland, Wayne, Monroe and Macomb Counties and obtains water from the Detroit River and Lake Huron. The City of Windsor and all the towns in Essex County obtain their water supplies from the Detroit River, Lake St. Clair or Lake Erie. The proximity of these surface water bodies precludes the use of groundwater for most of the larger municipalities within the study area. This is especially true for communities located in the generally non-productive clayey till plain region.

3.0 NATURAL TRACER INVESTIGATION

The tracing of groundwater flow in the study area is an integral part of the determination of the relative ages and origins of the groundwater. Environmental isotopes are naturally occurring tracers whose abundances can be used in hydrogeologic investigations to help trace the movement of groundwater through the subsurface. The traditional method of groundwater tracing, which involves the injection of a tracer into groundwater, yields information on a very small portion of the groundwater system, especially within less permeable material such as clays which are prevalent in the study area. In contrast, use of environmental isotopes allows a much larger scale of study since the natural tracers have been in the groundwater flow system for 10s to 10000s of years (Thatcher, 1967). The measurement of the field parameter, electrical conductivity, can be used to distinguish waters of different chemical character and origin and can aid in determining the flow path of groundwater through the freshwater aquifer.

3.1 Environmental Isotopes and Electrical Conductivity

In this study, the environmental isotopes oxygen-18, deuterium and tritium were used in conjunction with electrical conductivity to interpret the origin, history and movement of groundwater within the freshwater aquifer. These isotopes are ideal tracers for this investigation because they are conservative tracers that are constituents of water molecules. Oxygen-18 and deuterium are stable isotopes which are chemically conservative in groundwater, that is, their concentration within groundwater does not change unless two groundwater bodies with different isotopic concentrations exist in the same area (Thatcher, 1967). Tritium is a radioactive isotope which is not affected by reactions other than decay. These isotopes undergo mixing when waters of differing isotopic concentrations occur together.

3.1.1 Oxygen-18 and Deuterium

Oxygen-18 (^{18}O) is one of the three stable, natural isotopes of oxygen; it contains 8 protons and 10 neutrons. In comparison, the most abundant oxygen atom, oxygen-16, which makes up 99.763% of the total oxygen, contains only 8 neutrons (Fritz and Fontes, 1980). This indicates that ^{18}O is a heavier isotope than ^{16}O.

The hydrogen atom has three isotopes: protium (^{1}H), which is the dominant hydrogen isotope making up 99.984 % of the total hydrogen; deuterium (D or ^{2}H), a stable isotope with a mass number of two; and tritium (T or ^{3}H) (Fritz and Fontes, 1980). D, which is heavier isotopically then ^{1}H, is usually used in conjunction with [18]O for groundwater studies.

Since the concentrations of [18]O and D are low, the use of isotopic ratios to express concentrations is the norm. The ratio, which is the abundance of the heavy isotope divided by the abundance of the light isotope, is determined by mass spectrometry. The isotopic abundance ratios for oxygen and hydrogen is therefore represented in the following manner:

$$R = \frac{^{18}O}{^{16}O} \qquad\qquad R = \frac{D}{H} \quad\text{or}\quad \frac{^{2}H}{^{1}H} \qquad\qquad \text{(Dansgaard, 1964)}$$

Concentrations are expressed in delta (δ) units as parts per mil ($^0/_{00}$) differences between a standard and an unknown. The delta value is obtained using the following equation developed by Craig (1961):

$$\delta = \frac{R(sample) - R(standard)}{R(standard)} \times 1000^0/_{00}$$

where: $R = \frac{^{18}O}{^{16}O} \quad\text{or}\quad \frac{D}{H}$

The standard for water is Standard Mean Ocean Water (SMOW) and is assigned the value of zero for its delta value (i.e. $\delta^{18}O_{SMOW} = 0.0^0/_{00}$ and $\delta D_{SMOW} = 0.0^0/_{00}$) (Craig, 1961).

Light isotopes have higher vapour pressures than their heavy isotope counterparts. This difference in vapour pressure causes the lighter isotopes (^{16}O, H) to tend toward the vapour phase and the heavier isotopes (^{18}O, D) to tend toward the liquid phase through a series of condensation stages. This process is called "isotope fractionation" and is influenced by distance from the source, temperature, vapour exchange and evaporation (Thatcher, 1967). As oceanic water vapour moves through the hydrologic cycle it becomes progressively lighter as the heavier isotopes continually tend toward the liquid phase through condensation (Thatcher, 1967). Glacier or polar ice represents the final stage of fractionation, therefore it is the isotopically lightest water (i.e. it has the most depleted delta values). The concentrations in the stable isotope cycle range from the 0 $^0/_{00}$ of SMOW to approximately -50$^0/_{00}$ for $\delta^{18}O$ and -200$^0/_{00}$ for δD in polar ice (Fritz and Fontes, 1980). Positive values up to about +2$^0/_{00}$ for $\delta^{18}O$ and +20$^0/_{00}$ for δD are encountered under conditions of high evaporation.

Depletion of the heavy isotopes species (i.e. ^{18}O and D) has been related by Dansgaard (1964) to geographic parameters, such as latitude (temperature related), altitude, distance from the coast and the amount of precipitation. The latitude effect over the North American continent is roughly 0.5 $^0/_{00}$ $\delta^{18}O$/degree latitude (Yurtsever, 1975).

The relationship between $\delta^{18}O$ and δD for unevaporated waters, such as those originating from precipitation, is linear with a slope of about 8 (Figure 11). The intercept value depends on the origin of the condensing vapour, where a typical value for oceanic precipitation is about +10°/$_{oo}$. Thus, Craig (1961) found that in general, the linear relationship between $\delta^{18}O$ and δD for such waters can be described by:

$$\delta D = 8.0 \; \delta^{18}O + 10$$
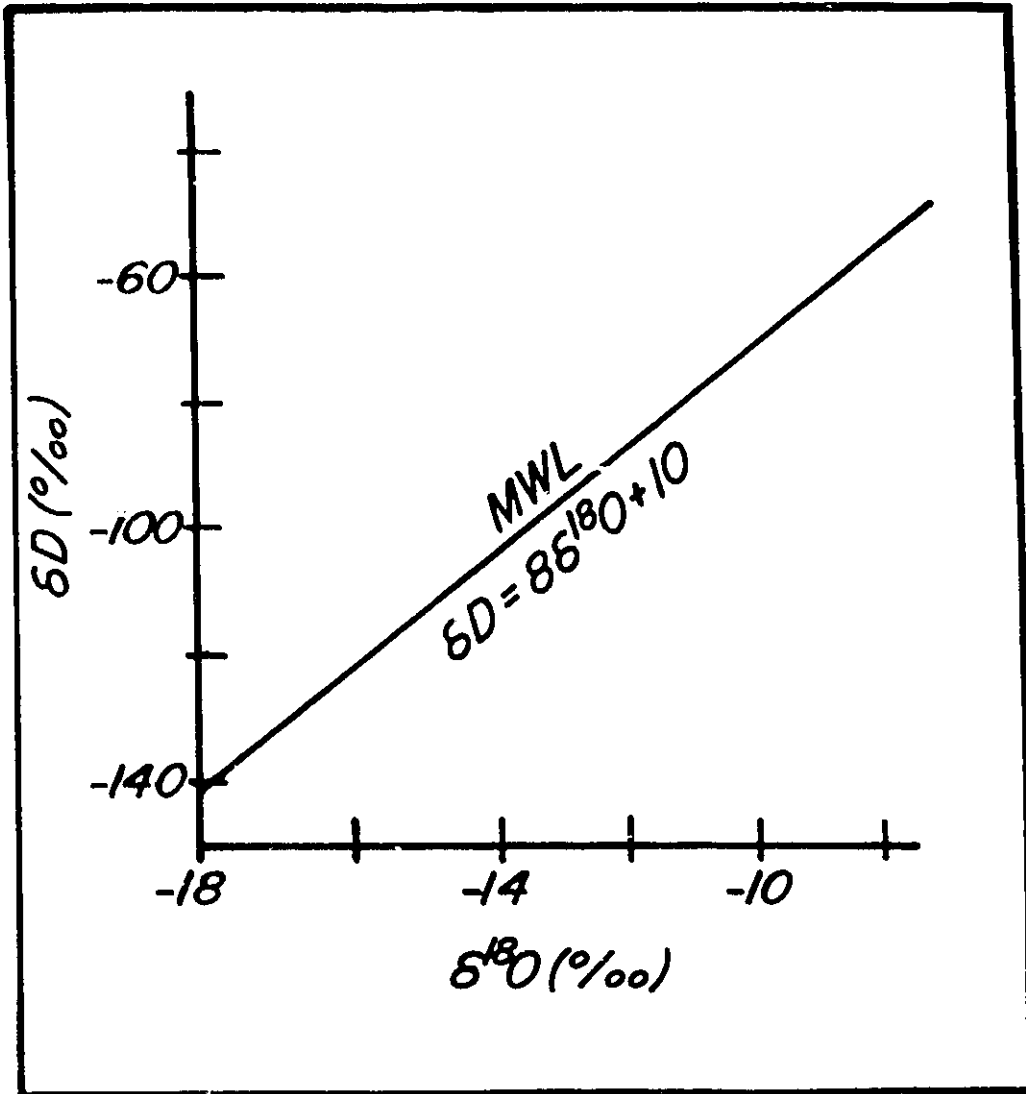
Figure 11   Meteoric water line (after Craig, 1961).

which is called the "meteoric water line". Evaporating bodies of water such as those from closed or inland seas also lie on straight lines. However their intercepts are generally greater than +10°/$_{oo}$ and their slopes range from 5 - 6 which are lower than those for non-evaporated waters (Craig, 1961).

3.1.1.1 Application of ^{18}O and D in Hydrogeologic Studies

A linear relationship exists between mean annual temperature and the $\delta^{18}O$ and δD values of precipitation (Dansgaard, 1964) where the isotopic content of precipitation becomes increasingly more negative (depleted) with decreasing temperature. It has been found that in temperate, humid areas the stable isotopic content of groundwater reflects the weighted average annual isotopic composition of precipitation in the recharge area (Fritz et al., 1976). Together these two relationships indicate that the isotopic composition of groundwater recharge water reflects the climate from which the meteoric waters originated.

The ^{18}O - temperature relationship can be used to identify areas of recharge and discharge and estimate groundwater age. Fritz et al. (1975) compared the ^{18}O content of glacial ice to that of deeply buried fossil shells located beneath Lake Erie. He concluded that very cold glacial conditions prevailed until about 10000 years ago, after which a rapid temperature improvement to current conditions occurred. Based on this evidence and $\delta^{18}O$ values observed in groundwater in tills, Desaulniers et al. (1981) suggested that groundwaters recharged in the past 10000 years in southwestern Ontario have $\delta^{18}O$ values which are similar to present precipitation $\delta^{18}O$ values of about -9 to -11 °/$_{oo}$ (I.A.E.A., 1979) and that groundwater recharged more than 10000 years ago in southwestern Ontario have depleted $\delta^{18}O$ values of about -16 to -20°/$_{oo}$. The distribution of ^{18}O can therefore also be used to determine groundwater flow directions during the last 10000 years. The isotopically depleted older groundwater (>10000 years old) has been progressively displaced by, or mixed with isotopically enriched younger recharge water (<10000 years old) in the downgradient direction.

3.1.1.2 Temporal $\delta^{18}O$ Variation of Meteoric Water

The stable isotopic composition of datable fossil material can be used for paleoclimate reconstructions. Edwards and Fritz (1986, 1988) conducted fossil wood-cellulose studies which related cellulose $\delta^{18}O$ and $\delta^2 H$ values of the wood to the isotopic concentrations of local meteoric water taken up by these plants growing during different time periods at Brampton, Ontario. Table 1 lists the estimated age ([14]C a B.P.) and inferred isotopic composition of past precipitation based on the cellulose isotope values (Edwards, personal communication, 1990).

Table 1 indicates that over a period of 11500 years, the $\delta^{18}O$ values have become enriched with time but not in a steadily increasing manner. From about 4000 a B.P. to the present, $\delta^{18}O$ values, in fact, steadily decreased which may indicate a gradual cooling trend. The cellulose studies of Edwards and Fritz (1986, 1988) resulted in paleotemperature curves which indicate that a postglacial climatic amelioration occurred, climaxing approximately 4500 a B.P. at mean annual temperatures 3-4°C above present temperatures. This increase in temperature is reflected in the enrichment of [18]O in meteoric waters from -15.2 to -8.9°/$_{oo}$ between 11450 to 4000 a B.P. with minor downward [18]O fluctuations during this time interval. The cellulose studies also indicate that from about 4500 to 1500 a B.P. the mean annual temperature in the Brampton area decreased from about 10.3°C to the modern value to 7.1°C. From about 1500 a B.P. to the present, these studies imply that the mean annual temperature has remained relatively constant. Table 1 shows the corresponding changes in the $\delta^{18}O$ value of meteoric water, decreasing to a value of about -11°/$_{oo}$ about 1500 a B.P. and stabilizing thereafter.

3.1.1.3 Regional Studies

Several studies have investigated the age and origin of groundwater in the freshwater aquifer and the overlying drift within the locality of the study area. Desaulniers et al. (1981) measured the $\delta^{18}O$ values of groundwater at three sites situated in clayey till within the present study area. Values ranged from a water-table value of -10°/$_{oo}$ (SMOW) to deep (20-40 m) groundwater values of -14 to -17°/$_{oo}$. The shallow

Estimated Age (^{14}C a B.P.)	$\delta^{18}O$ Meteoric Water ($^o/_{oo}$ SMOW-infer.)
modern	-11.1
1500	-10.7
3850	-9.3
4000	-8.9
6300	-8.9
7510	-10.9
7700	-10.9
7900	-12.9
8700	-12.7
9550	-14.1
10450	-13.6
10950	-14.4
11450	-15.2

Table 1: $\delta^{18}O$ values of meteoric waters in the Brampton area and their estimated time of occurence (Edwards, personal communication, 1990).

groundwater values are indicative of present day precipitation while the deeper values are characteristic of cooler, glacial waters (^{14}C age > 8000 years). An isotopic study of saline near-surface (<100 m deep) groundwater in the east-central Michigan basin by Long et al. (1988) had an isotopic range of -9 to -19$^o/_{oo}$. The geologic and hydrogeologic setting of Long's study area is similar to this study area's. Fritz et al. (1975) measured the ^{18}O contents of shells and mollusks from a basal till unit beneath Lake Erie. The concentration ranged between -18 and -20$^o/_{oo}$ with emplacement occurring between 12800 and 13800 yr. B.P. These studies imply that deep groundwaters in this geographic region are a mixture of modern meteoric water and water that recharged the aquifer system during a cooler time period.

3.1.2 Tritium

Tritium (^{3}H) is a naturally occurring, radioactive isotope of hydrogen with a mass number of 3 and a half-life of 12.35 years. Tritium concentrations are expressed in tritium units (TU) where one TU represents one tritium atom for every 10^{18} protium (^{1}H) atoms (Hoffman and Stewart, 1966).

Tritium is produced naturally in low concentrations in the atmosphere by cosmic radiation, however, the largest amounts of tritium were produced as a by-product of above ground thermonuclear bomb testing which began in the early 1950's. It is estimated by Brown (1961) that pre-bomb background levels in precipitation were in the range of 5 - 20 TU. The record of tritium in precipitation (Figure 12) illustrates how atmospheric tritium concentrations began to increase sharply in the years following the nuclear tests until a peak of several thousand TU's occurred in 1963.

Since the half-life of tritium is 12.35 years, water recharged before 1952 should now have less than about 5 TU, while younger water, recharged after 1952, should have more than about 5 TU. Usually detection limits for tritium are below 5 TU, as is the case for this study, therefore it is generally regarded that water with more than 20 TU was recharged after 1952 (Egboka et al., 1982).

3.1.3 Electrical Conductivity

Electrical conductivity (EC) is a numerical expression of water's ability to conduct an electrical current and is the reciprocal of electrical resistivity (Hem, 1970). In water
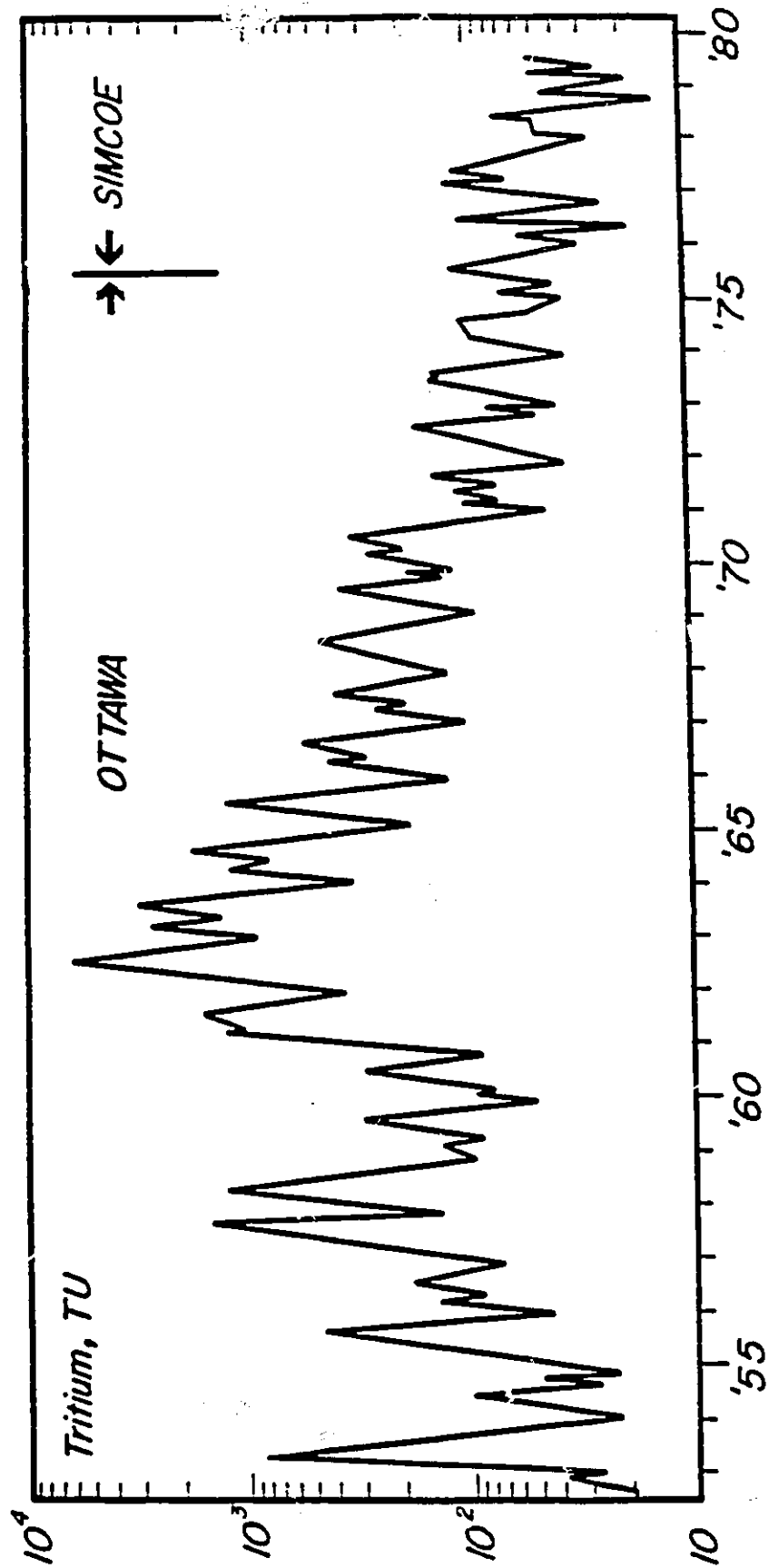
Figure 12   Tritium in precipitation versus time recorded in Ottawa (after I.A.E.A., 1979).

analyses, electrical conductivity is reported in microsiemens/cm ($\mu S/cm$). Specific electrical conductance is defined as the conductance of a cubic centimeter of any substance compared with the conductance of the same volume of water (Driscoll, 1986). It is dependent on temperature and on the type and concentration of the dissolved ions. It is usually defined at 25°C, so that differences in conductance are a function of the concentration and type of dissolved ions only (Matthess, 1982).

There is a strong linear correlation between electrical conductivity and total dissolved solids (TDS) in groundwater, where electrical conductivity increases as TDS concentration increases (Driscoll, 1986). This linear increase occurs as groundwater travels away from a recharge zone, picking up ionized particles along the way. As groundwater moves along its flow paths in the saturated zone, increases of total dissolved solids and most of the major ions normally occur (Freeze and Cherry, 1979). The longer the residence time of the groundwater in an aquifer, the more opportunity there is for ionized particles to dissolve into it.

3.2 Methods

3.2.1 Groundwater Sampling

3.2.1.1 Sample Site Selection

For the study, 67 samples from domestic water wells were obtained within the Michigan study area and 7 samples within Ontario. The total number of sites for the freshwater aquifer investigation is 220 and includes samples from other studies completed within the study area: MacGregor (1985), Essex County; Scott (1986), Lambton County; Baxter (1987), Kent County; Crnokrak (1987), northeastern Essex County; Erdmann (1987), Macomb and St. Clair Counties; and Williams (1988), the Sarnia area. Figure 13 shows the location of the sample sites.

Sample sites were chosen with an attempt to obtain reasonable areal coverage of the study area. Locations for freshwater aquifer wells in Michigan were selected using data from the Michigan Department of Natural Resources' (DNR) Geological Survey Division (GSD) in Lansing, Michigan. Within the GSD, the Glacial and Groundwater Geology
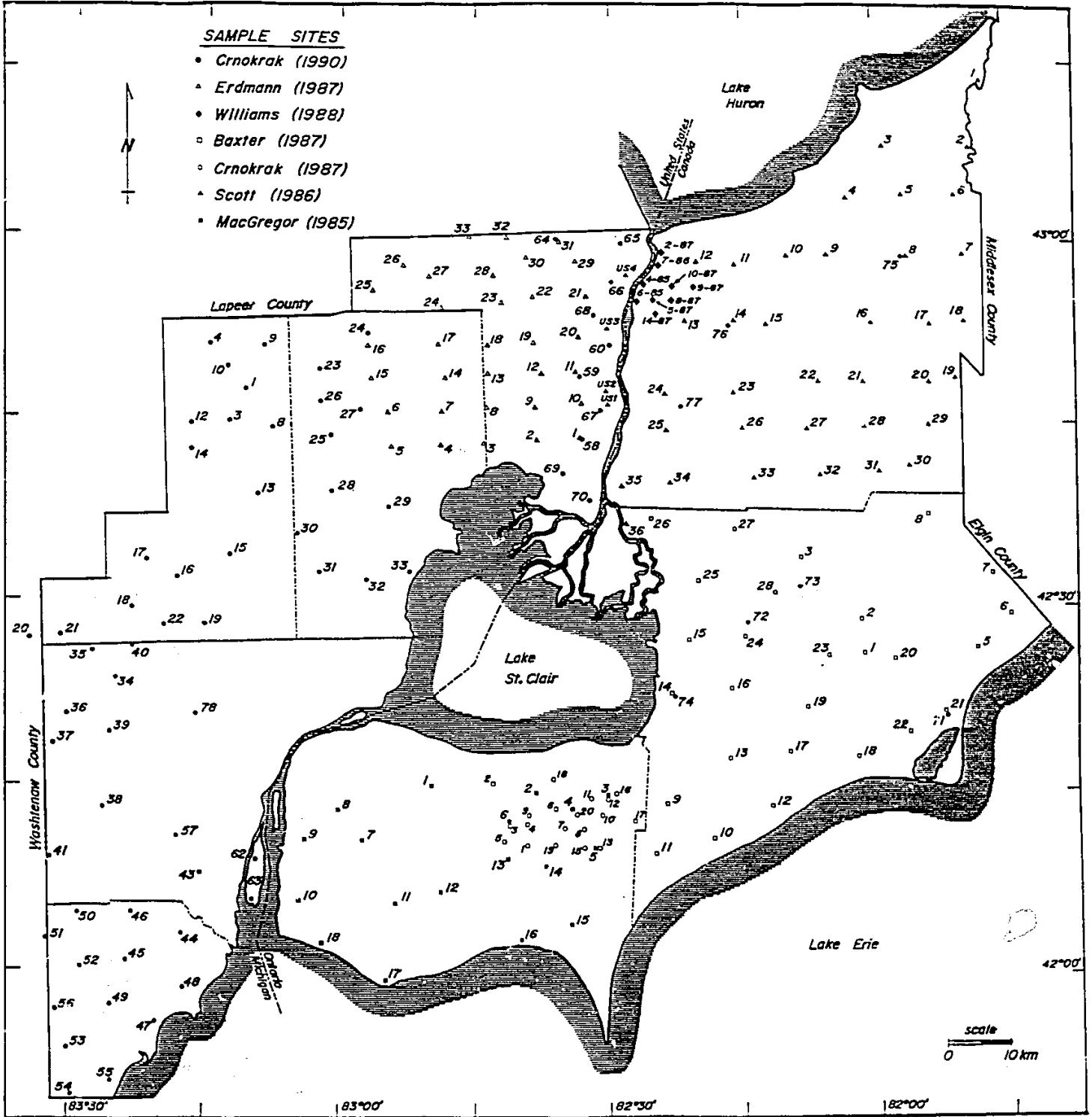
Figure 13   Location of well sample sites.

Unit maintains water well record files from which this study's sample sites were selected. In Ontario, sample sites were chosen from water well records on file with the Ontario Ministry of the Environment (1977, 1984).

Examination of these records revealed a scarcity of freshwater aquifer wells in some areas and as a result some portions of the study area have poor well coverage. The metropolitan Detroit and Windsor areas are completely devoid of productive wells as these localities obtain their water supplies from surface water sources. Freshwater aquifer wells were also scarce in areas where sufficient water supplies are readily available in glacial deposits such as outwash sand and gravel and lacustrine sand. The area of interlobate outwash deposits in the northwestern portion of the study area and the sand plain deposits of southwestern Wayne County and northeastern Kent County were areas of poor well coverage.

Well record examination also revealed that most of the Erdmann (1987) samples had been obtained from shallow drift wells and not from wells which penetrated the freshwater aquifer as was required for this investigation. Only 9 of the 33 samples were freshwater aquifer wells and they are noted in Appendix II.

3.2.1.2 Sampling Procedure

Groundwater samples were collected in 250 mL polyethylene "Nalgene" bottles which were filled completely to avoid any "head space". At every tenth sample site, field blanks and replicate samples were obtained to serve as laboratory checks.

Electrical conductivity measurements were taken at each site using a YSI Model 32 conductivity meter. This instrument has fully automatic temperature compensation to the standard temperature of 25°C. Meter calibration was done before and after each field outing by immersing the instrument's electrode into two standard KCl solutions and observing whether the readings varied from each standard's known electrical conductivity. The readings were recorded and found to fall within the typical ±5% error associated with conductivity meters.

Well samples were taken after pumping for an appropriate length of time (usually 5 to 10 minutes) to ensure a fresh sample. Precautions were taken to ensure that any in-house water-treatment facilities were bypassed.

3.2.2 Isotope Analysis

The 74 samples collected for this investigation were analyzed for oxygen-18 concentration. Oxygen-18, deuterium and tritium analyses had been performed on the samples collected previous to this investigation.

All isotopic analyses were performed at the Environmental Isotope Laboratory of the Department of Earth Sciences at the University of Waterloo. The oxygen-18 and deuterium determinations were done using a Micromass Model 903 mass spectrometer. The analytical precision of the mass spectrometer is $\pm 0.2^{\circ}/_{oo}$ for $\delta^{18}O$ and $\pm 2.0^{\circ}/_{oo}$ for δD. Tritium concentrations in water were measured by direct liquid scintillation counting using an Intertechnique or Beckman Model 7500 Scintillometer. This method of tritium analysis has a detection limit of approximately 6 TU and an analytical precision of ± 6 TU to ± 8 TU. Sample 2-87 from Williams (1988) had its tritium concentration measured by the enriched method which has a detection limit of about 0.8 TU and an analytical precision of ± 1 TU.

3.3 Results and Discussion

3.3.1 Isotopes and Electrical Conductivity

The concentrations of oxygen-18, deuterium and tritium and the electrical conductivity of the freshwater aquifer groundwater samples can be used to determine the age, origin and movement of groundwater in the aquifer. The isotope analysis results and electrical conductivity measurements obtained during this investigation and those acquired from previous studies within the study area (MacGregor, 1985; Scott, 1986; Baxter, 1987; Crnokrak, 1987; Erdmann, 1987; Williams, 1988) are listed in Appendix II.

3.3.1.1 Oxygen-18 and Deuterium

Earlier in this report it had been established that waters of glacial origin (>10000 years old) with $\delta^{18}O$ values of -16 to $-20^{\circ}/_{oo}$ had been emplaced into the freshwater aquifer within the study area during colder climatic conditions. An ameliorating climate since then has resulted in precipitation becoming progressively enriched to present day $\delta^{18}O$ values of -8 to $-10^{\circ}/_{oo}$. The distribution of $\delta^{18}O$ values in the freshwater aquifer (Figure 14) shows this same range of values. The concentration of ^{18}O was found to be

Figure 14   Contour map of oxygen-18 in the freshwater aquifer.

greater near the recharge areas identified in section 2.5.2.2 relative to areas proximal to the discharge water bodies. Between the areas of recharge and discharge, the concentrations of ^{18}O become progressively more depleted towards the discharge area.

The frequency distribution for values of $\delta^{18}O$ in the groundwater samples is illustrated in Figure 15. The distribution is bimodal about a mean of approximately $-10^{\circ}/_{oo}$ and another mean of about $-16^{\circ}/_{oo}$.

The plot of oxygen-18 versus deuterium (Figure 15) shows that the well samples taken from the freshwater aquifer produce the meteoric water line for the study area represented by the equation:

$$\delta D = 7.5\delta^{18}O + 2.89 \qquad (r = 0.99)$$

Since none of the sample points plot below the line and this line is parallel to the meteoric water line that Desaulniers et al. (1981) obtained for southwestern Ontario, it can be assumed that this line represents unevaporated meteoric water.

There are five regional flow patterns which can be recognized from the pattern of the $\delta^{18}O$ contours. These patterns are in agreement with flow that was identified through the physical interpretation of the geology and hydrogeology of the study area discussed in section 2.5.2.1. These five regions have the following locations: 1. the northern 2/3 of the Michigan study area; 2. Lambton and northwestern Kent Counties; 3. southern Kent County; 4. Essex County; 5. the southern third of the Michigan study area. The interpretation of the $\delta^{18}O$ values within each region will subsequently be discussed.

REGION 1 - The northwestern corner of the study area contains $\delta^{18}O$ values of -9 to $-11^{\circ}/_{oo}$ which are representative of recent precipitation $\delta^{18}O$ values for the region (I.A.E.A., 1979). The enriched $\delta^{18}O$ values are concurrent with the location of the interlobate region recharge area. The $\delta^{18}O$ contours become gradually depleted to values of -16 to $-17^{\circ}/_{oo}$ to the southeast. The contours therefore indicate that the flow of
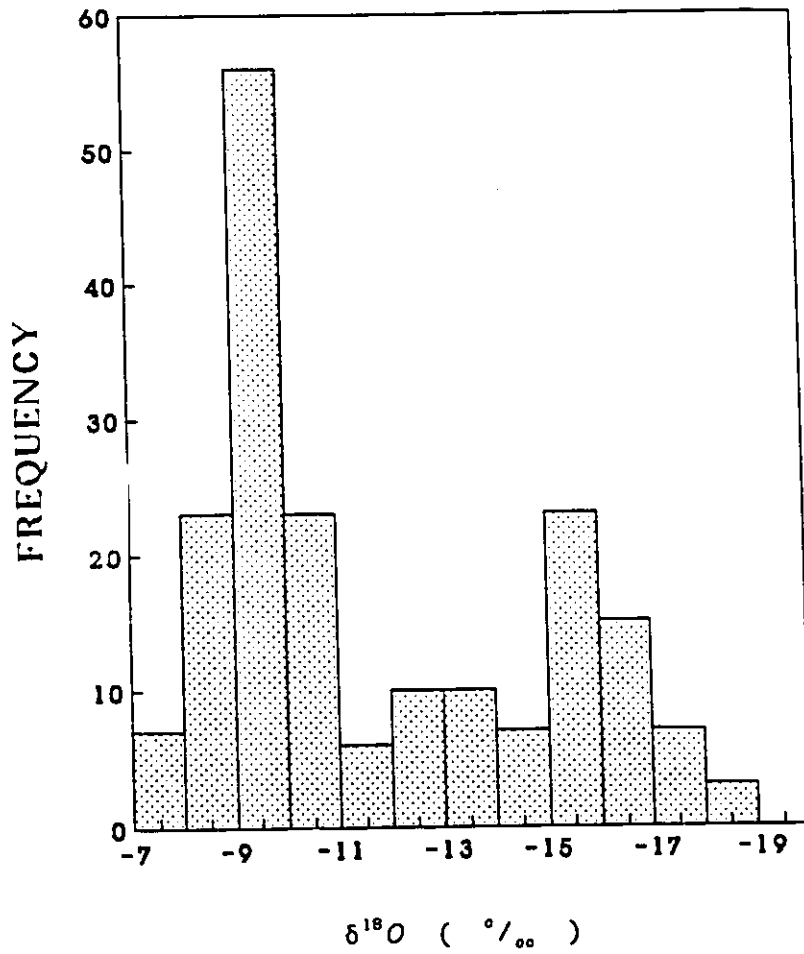
Figure 15   Frequency histogram of $\delta^{18}O$.
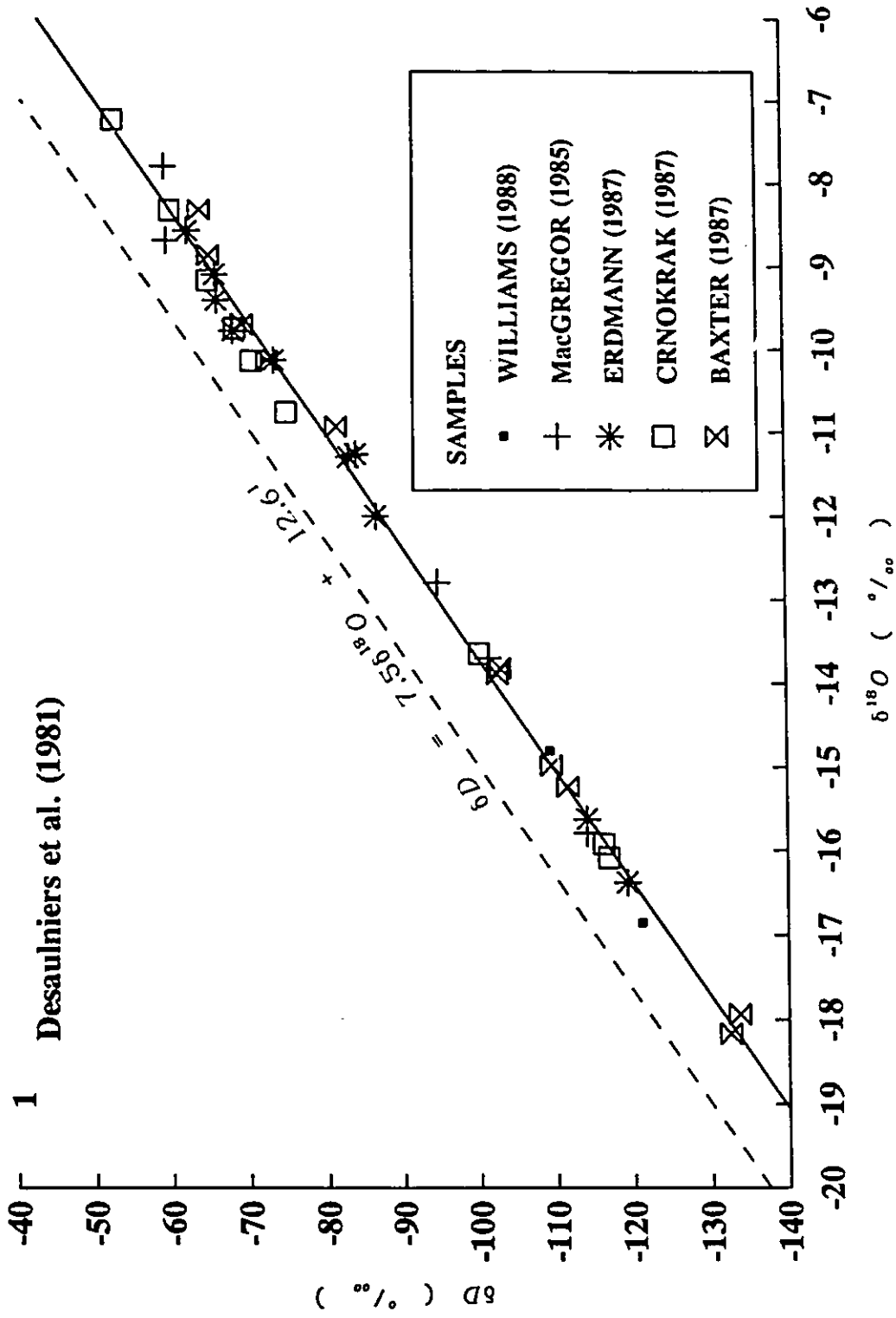
Figure 16 $\delta^{18}O$ vs. δD plot for the freshwater aquifer.

groundwater within this flow region is from the northwest, originating within the interlobate recharge area, to southeast toward the St. Clair River, Lake St. Clair and the Detroit River. The contours also indicate that as the groundwater flows southeast, the water is progressively dominated by older water (depleted $\delta^{18}O$ values) which is being displaced by younger water located upgradient.

There is a prevalence of younger water with values of -9 to $-10^o/_{oo}$ found within the glacial till plain section of northeastern Macomb and central St. Clair Counties. This surge of enriched ^{18}O water would imply that groundwater had either travelled farther and faster from the interlobate recharge area through this region than areas to the south, or that there was infiltration of younger water occurring from above. These enriched $\delta^{18}O$ values have been attributed to the shallow drift well samples obtained by Erdmann (1987) which would naturally have a heavier isotopic signature than deeper freshwater aquifer well samples (Desaulniers et al., 1981).

REGION 2 - Within Lambton County the $\delta^{18}O$ values range from recent precipitation values of -8.5 to $-10^o/_{oo}$ found along the eastern border of the county to very depleted values in excess of $-17^o/_{oo}$ just east of the St. Clair River. Depleted values of this concentration represent waters of glacial origin.

There are two directions of flow implied from the trend of $\delta^{18}O$ contours. One of these is denoted by a westward bulging of the $-10^o/_{oo}$ contour which coincides with the location of the recharge area in eastern Lambton County. The $\delta^{18}O$ contours indicate that flow is westward toward the St. Clair River and northwestward toward Lake Huron. The second flow path is indicated by the broad band of enriched $\delta^{18}O$ values (-8.5 to $-10^o/_{oo}$) which trends northeast-southwest and is located within the Bothwell Plain-Thames River recharge area. The direction of flow from this area is northwest towards the St. Clair River. The flow from these two recharge areas likely converges in the center of Lambton County.

Comparing the contour trends on either side of the St. Clair River it appears that flow is converging towards the St. Clair River. The southeastern flow of groundwater from the Michigan side mirrors the northwestern flowing groundwater system on the Ontario side.

A number of wells on either side of the St. Clair River, particularly along its northern reach, have enriched ^{18}O concentrations relative to neighbouring upgradient wells. Well samples 65, US4, SS-12, 4-85, 6-85 and US1 have $\delta^{18}O$ values which range from -10.32 to -14.81°/$_{oo}$. The $\delta^{18}O$ contours of Figure 14 suggest that most of these enriched samples should have $\delta^{18}O$ values in excess of -16°/$_{oo}$. Available tritium data indicates that these samples are not tritiated, thereby eliminating river water as a possible source of ^{18}O enrichment. Intera (1989) proposed that a possible explanation for the enriched $\delta^{18}O$ values may be the mixing of glacial age meteoric water with isotopically heavier saline waters originating from deeper bedrock formations. Such a scenario had been suggested by Long et al. (1988) to explain the stable-isotope geochemistry of saline near-surface groundwater along the southeastern edge of Michigan.

REGION 3 - South of the Thames River valley another pattern of groundwater flow within the freshwater aquifer is implied by $\delta^{18}O$ contours. Immediately south of the Thames River, $\delta^{18}O$ values of -9 to -10°/$_{oo}$, indicative of the present day precipitation values, increase to values in excess of -18°/$_{oo}$ near Lake Erie. The most depleted value of -18.46°/$_{oo}$ represents the oldest water within the study area.

The pattern of $\delta^{18}O$ contours suggests that groundwater flow is predominantly southeast toward Lake Erie. A southward deflection of the contours, which originates south of Chatham may be influenced by the northeastward movement of groundwater flow from Essex County. The two flow systems converge near southwestern Kent County where mixing of the older glacial waters from each system occurs. The push of directionally opposing flow systems may cause deflection of both system's contours. However, this deflection may simply be due to the low density of sampling points which reduces the confidence of the $\delta^{18}O$ contours. The infusion of enriched recent $\delta^{18}O$ values of -9 to -10°/$_{oo}$ in southeastern Kent County near the Elgin County border may be the result of the presence of a bedrock high which underlies relatively thin drift, 20 to 25 m thick, located just west of the county border. The presence of surficial and interbedded

sands in the vicinity may have allowed the vertical infiltration of enriched ^{18}O recent precipitation into the freshwater aquifer with the bedrock surface slope diverting groundwater flow northeast into Elgin County.

REGION 4 – The $\delta^{18}O$ values in Essex County trend from recent precipitation values of approximately -9 to -10°/$_{oo}$ in the recharge area near the Leamington Moraine, to values of -15.5°/$_{oo}$ in the northwest and -16.5°/$_{oo}$ in the northeast.

The flow directions indicated by the attitude of the $\delta^{18}O$ contours is north towards Lake St. Clair and the northern reach of the Detroit River and east into Kent County. Although there is insufficient data to contour the extreme southern portion of Essex County, groundwater appears to flow south from the recharge area towards Lake Erie, based on several sample points from MacGregor (1985) and the slope of the bedrock surface.

A surge of enriched $\delta^{18}O$ values appears directly north of the recharge area in the central portion of Essex County and extends further north than the contour profiles east and west of this area. The presence of this younger water with $\delta^{18}O$ values of -8 to -9°/$_{oo}$ is puzzling. Till thickness here is in excess of 30 m which suggests that vertical recharge is unlikely, and the lateral flow distance from the Leamington recharge area is large. The discovery of a north-south trending buried esker by Dr. T.F. Morris of the Ontario Geological Survey (personal communication, 1990; Morris, 1988) which coincides with the location of the anomalously young groundwater, may be the answer. Dr. M.G. Sklash of the University of Windsor has formulated an hypothesis whereby the buried esker may be responsible for the anomalously young groundwater (personal communication, 1990). The esker may have a high hydraulic conductivity which has permitted rapid lateral groundwater movement or it may have caused deep, localized, vertical fractures in the clayey till overlying this area as it settled over the esker, thereby providing a rapid vertical infiltration path through the aquitard. Researchers at the University of Windsor are currently verifying the limits of the esker and the existence of the young groundwater. Andrew Ainslie (personal communication, 1991) has identified groundwater, from a well he believes has intersected the top of the esker, which has a $\delta^{18}O$ value of -7.48°/$_{oo}$, which may confirm the 'esker - young groundwater' hypothesis.

REGION 5 - The freshwater aquifer groundwater in this area contains the highest concentration of [18]O within the study area where the southwestern portion of the region contains groundwater with $\delta^{18}O$ values which range from -7.5 to -8.5°/$_{oo}$.

The [18]O concentration depletes to values of -12 to -13°/$_{oo}$ to the northeast. These intermediate $\delta^{18}O$ values suggest mixing of older glacial water with relatively recent water. The direction of [18]O depletion suggests that flow within this system is northeast toward the Detroit River and Lake Erie.

3.3.1.1.1 Relationship to Till Thickness

There is a relationship that seems to exist between the value of $\delta^{18}O$ and the thickness of till over the freshwater aquifer. Figure 17 illustrates the general trend of increasingly depleting $\delta^{18}O$ values with increased till thickness when following the plot of sample sites from separate studies. Samples obtained from the interlobate recharge area in Michigan were not included in this plot as there is little or no clay till at depth in this region. Desaulniers et al. (1981) also found that there was a distinctive shift of $\delta^{18}O$ values with depth suggesting that deeper groundwater originated under cooler climatic conditions. Areas of thicker till within the study area prevent the rapid infiltration of surface meteoric water to the aquifer due to the low matrix permeability of the clayey till. Oxygen-18 concentrations characteristic of recent precipitation were observed within the areas of thin drift in the Thames River valley and Monroe County. The till within these areas is therefore thin enough to allow recent meteoric water to infiltrate to the freshwater aquifer.

Figure 17 also shows that there are sample sites which do not follow the trend of depleting $\delta^{18}O$ values with increased till thickness:

1. Three samples from the Baxter study (JB-5, 6 and 7) plotted at the top of the graph. These samples were obtained from the freshwater aquifer where it is overlain by the Blenheim Moraine. This feature forms an area of great local drift thickening which thereby pushes these samples to the top of the plot.
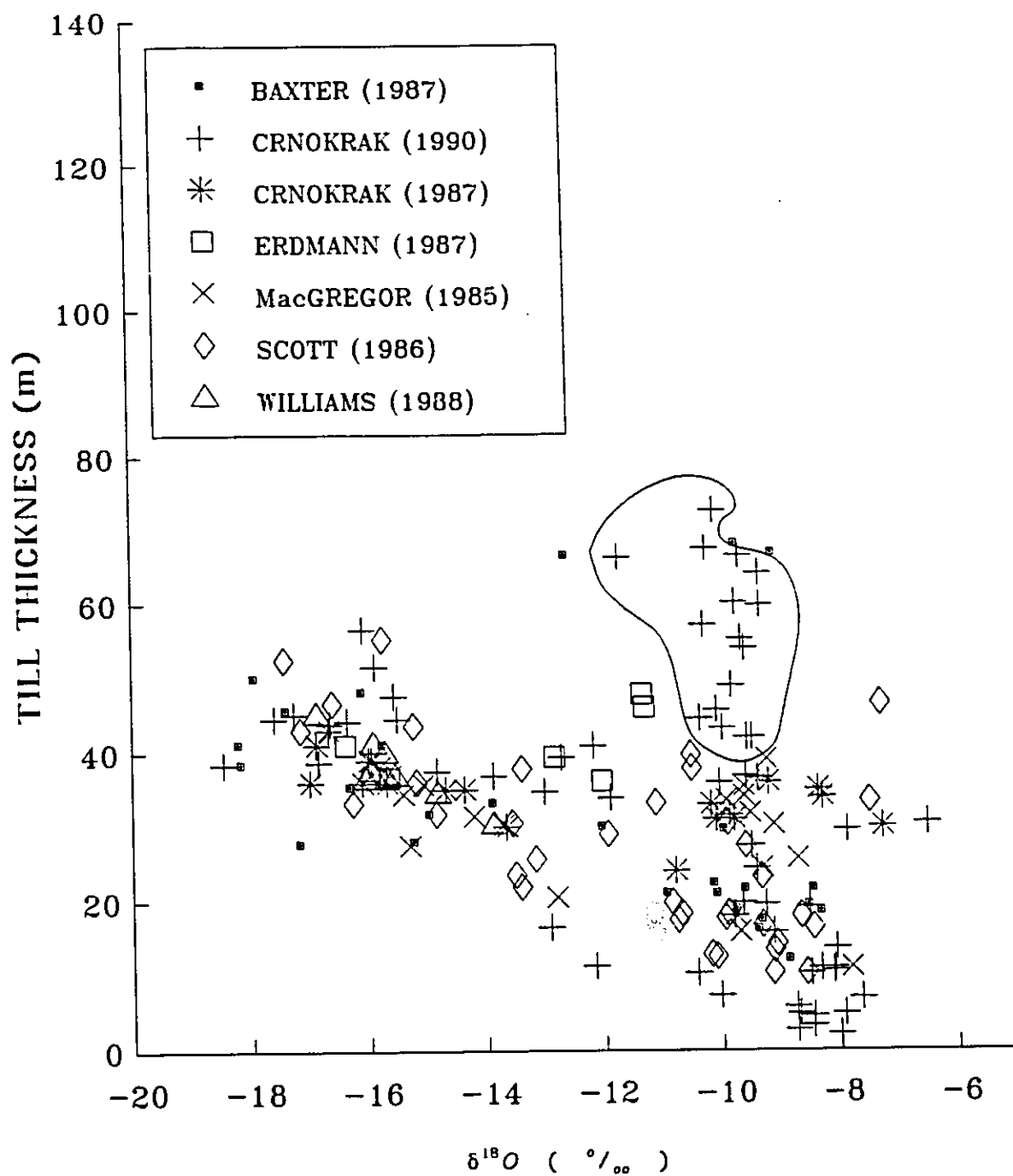
Figure 17   Plot of $\delta^{18}O$ verus till thickness.

2. The group of samples from the Crnokrak (1990) study which are circled on Figure 17 are all located in the glacial till plain on the Michigan side just outside the interlobate recharge area. The $\delta^{18}O$ values in this area are expected to be enriched because of the proximity of these samples to the recharge area.

3. Samples SS-15, SS-24 and BC-19 have isotopic values that are similar to the average St. Clair River water samples ($-7.25^{\circ}/_{oo}$ $\delta^{18}O$ and 89 TU) observed by Mason et al. (1986). These samples were likely obtained from domestic water supplies that were connected to public surface water systems.

3.3.1.2 Tritium

Figure 18 shows the areal distribution of tritium in the freshwater aquifer. The frequency distribution of tritium illustrated in Figure 19 shows that the population distribution is not random and that it is skewed to the left toward lower tritium values (<20 TU) which represent groundwaters that have been recharged prior to 1952.

The distribution of tritium suggests that higher concentrations of tritium are present in the known recharge areas and that the freshwater aquifer within the majority of the tritium sample area contains non-tritiated water. The groundwater samples from the northeastern Michigan study area, collected from the Erdmann (1987) study area, were not obtained from wells penetrating the freshwater aquifer and are therefore not representative of the aquifer in this area. For this reason the Erdmann (1987) tritium concentration data has not been included in Figure 18.

The presence of tritiated water (>20 TU), which implies that water entered the aquifer less than 40 years ago, in central Essex County north of the Leamington recharge area, is coincident with the enriched $\delta^{18}O$ values of the groundwater in this area. This water is too young to have reached the freshwater aquifer by vertical infiltration through the confining till or by lateral movement through the aquifer from the recharge area. These areas of anomalously young water coincide with buried eskers identified by Morris (1988). The eskers may have allowed tritiated water to enter the aquifer in two ways. One way may be laterally, if the esker is composed of highly hydraulically conductive sands and gravels. A second approach may be vertically, where deep fractures may have formed in the clayey till overlying this area as it settled over the eskers, thereby providing a rapid vertical infiltration path through the aquitard. The
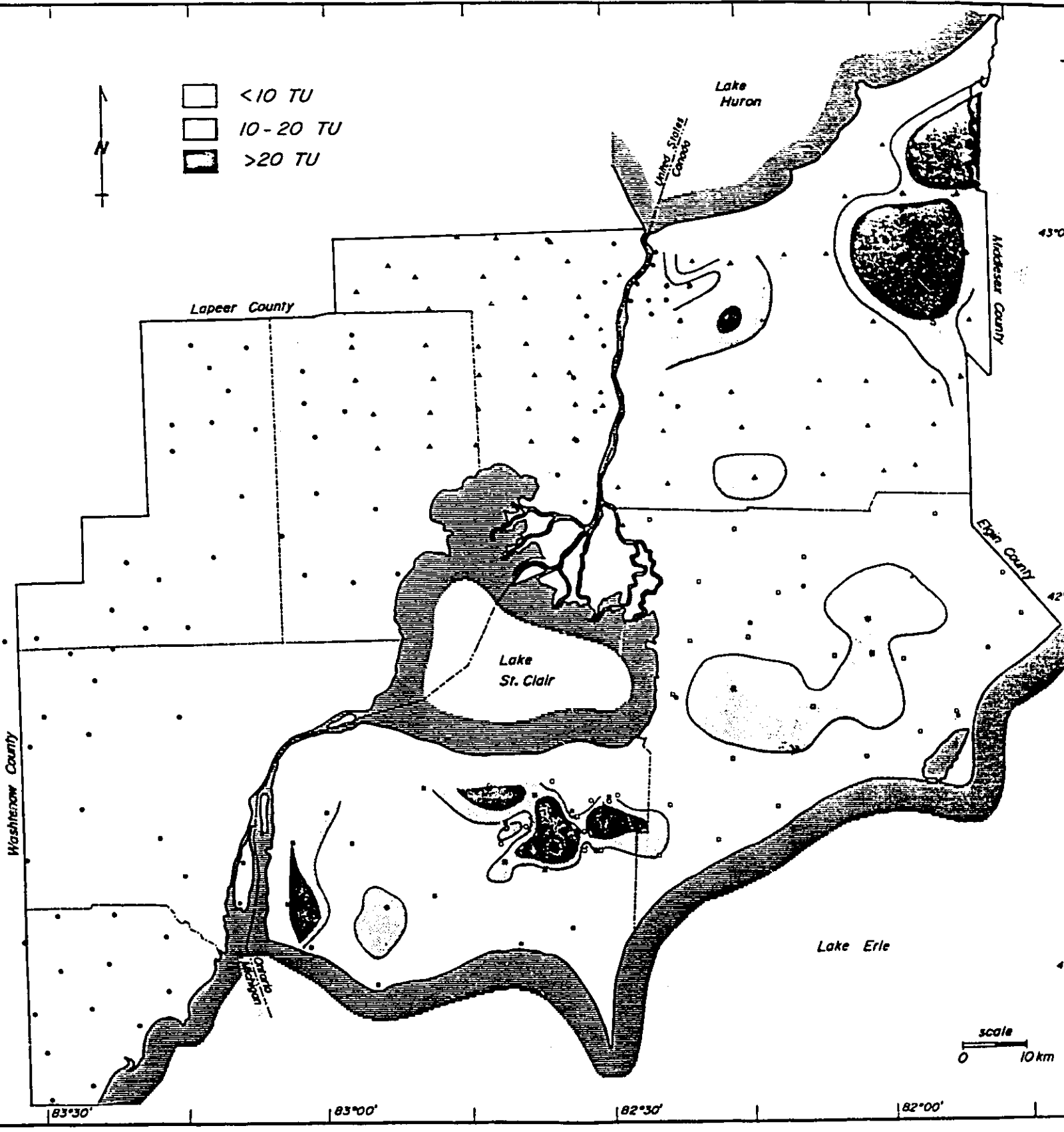
Figure 18   Contour map of tritium in the freshwater aquifer.

49
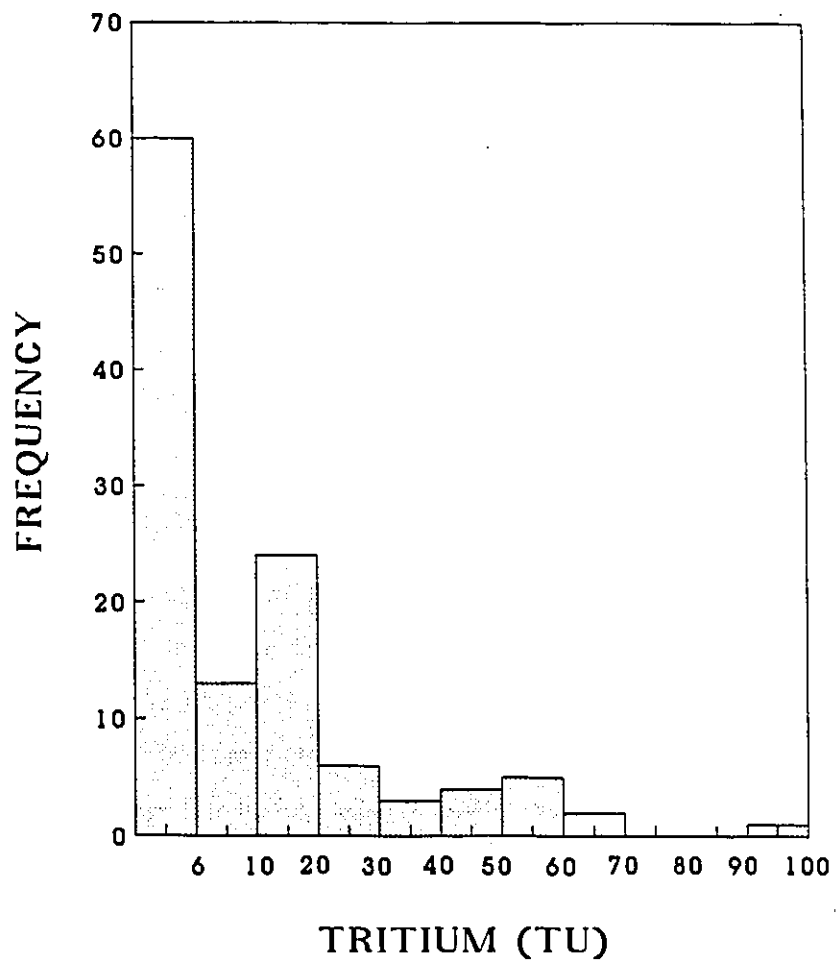
Figure 19   Frequency histogram of tritium.

existence of the esker and the presence of tritiated water in its vicinity, is currently being investigated by researchers at the University of Windsor. Alternatively, the high tritium values may be the result of contamination by surface waters due to faulty wells allowing surface waters to travel to the aquifer along the well casing. The small pocket of elevated tritium in southwestern Essex County can be attributed to the lack of sufficient till cover (~5m) which allows easy infiltration of young waters to the freshwater aquifer.

There appears to be a weak relationship between $\delta^{18}O$ and tritium which is illustrated in Figure 20. The plot shows that most of the samples with very depleted $\delta^{18}O$ values have low tritium concentrations (<20 TU). In addition, samples above 20 TU show a trend toward lower tritium concentrations with depletion of $\delta^{18}O$ values. Similarly these tritiated samples show a tendency to decreased tritium concentrations with increased till thickness (Figure 21). Below 20 TU, a concentration which is interpreted as being non-tritiated with the direct counting method, there is no apparent simple relationship between tritium content and till thickness.

3.3.1.3 Electrical Conductivity

The areal distribution of EC is illustrated in Figure 22. The frequency histogram for EC (Figure 23) shows the sample distribution to be skewed toward lower values.

In general, the EC contours illustrate the same pattern of groundwater flow indicated by the $\delta^{18}O$ data contours and the hydraulic head. The lowest EC values are observed within the five major recharge areas where EC < 1000 $\mu S/cm$. EC steadily increases in the downgradient direction towards the discharge zones. This distribution is reasonable since it suggests that groundwater in the freshwater aquifer is transporting an increasing quantity of solids as it travels from recharge to discharge areas. The major directions of flow implied by the EC contours are: 1. southeast from the interlobate recharge region to the St. Clair River and Lake St. Clair; 2. west and northwest from the recharge area in eastern Lambton County to the St. Clair River and Lake Huron; 3. northwest and south from the Thames River recharge area to Lake St. Clair, the St. Clair River and Lake Erie; 4. north, northeast and west from the southern Essex County
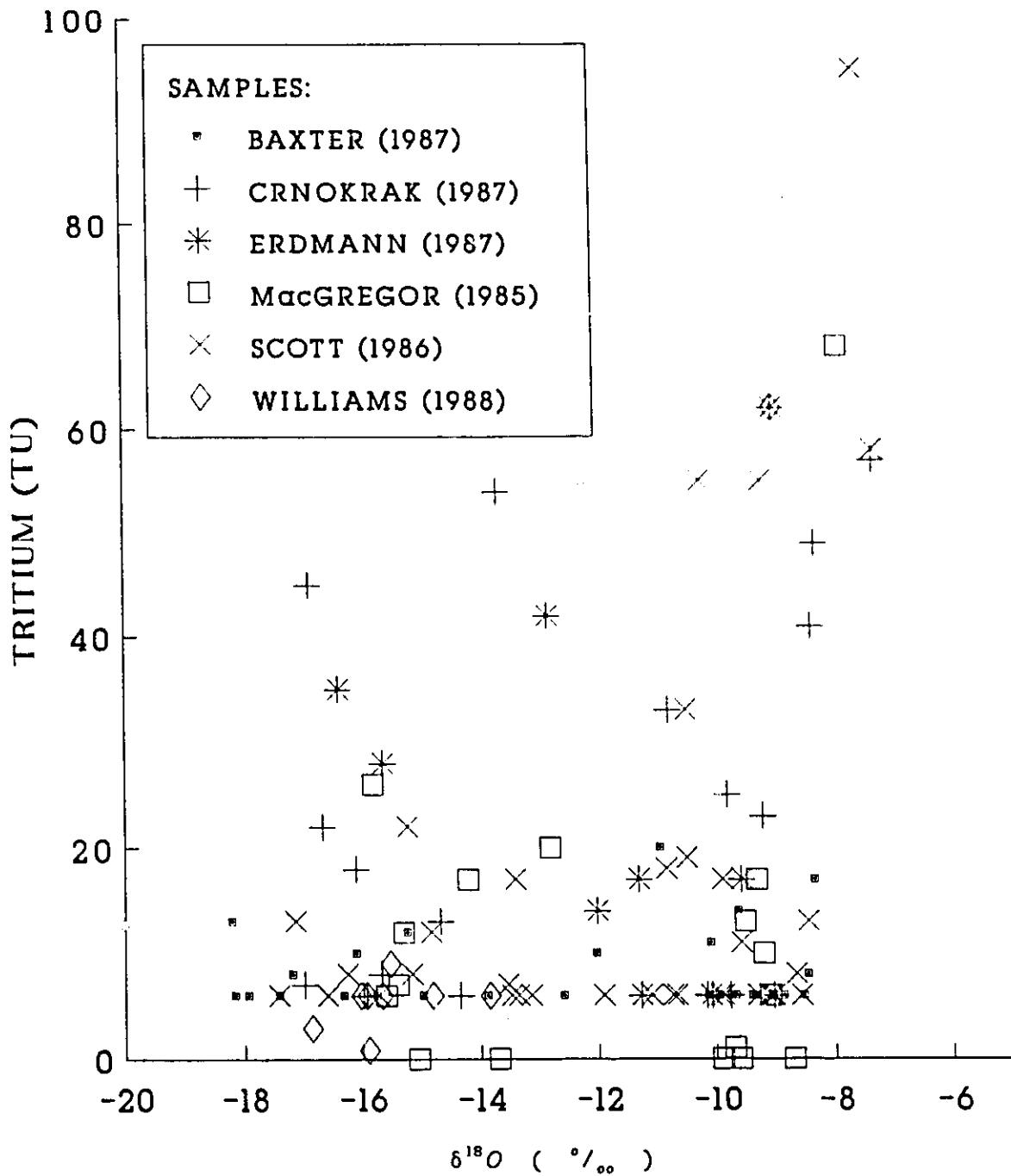
Figure 20   Plot of $\delta^{18}O$ versus tritium.
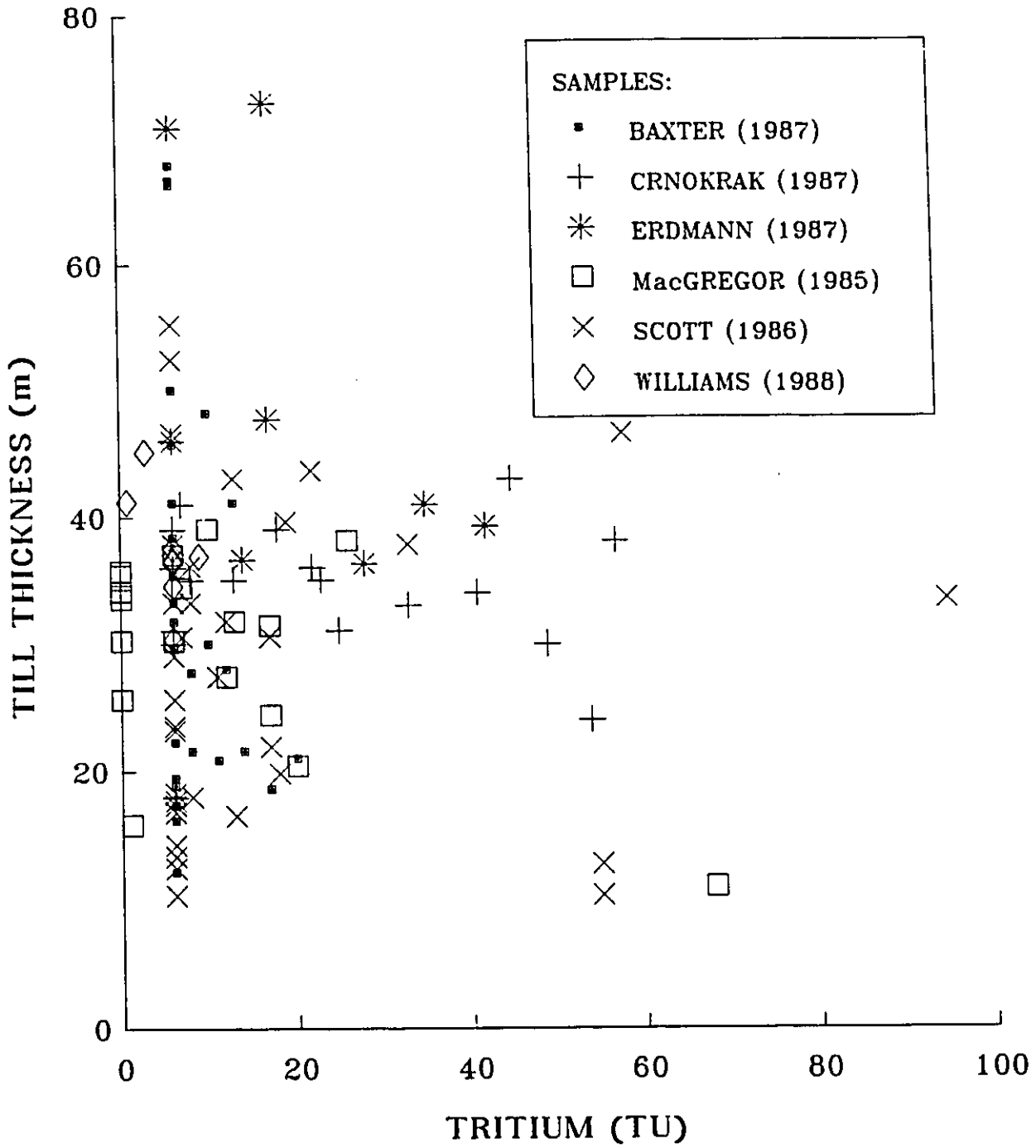
Figure 21   Plot of tritium versus till thickness.
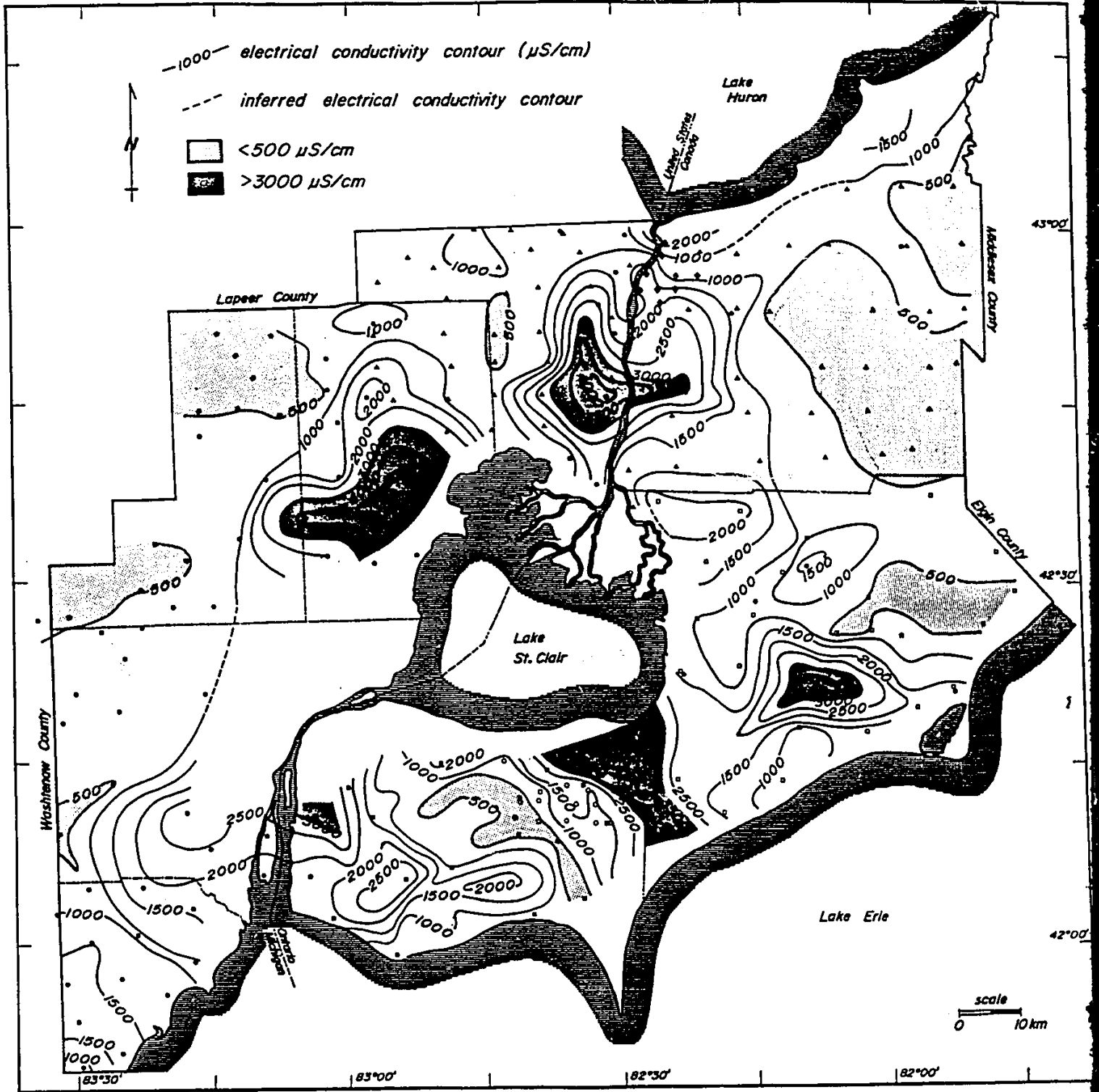
Figure 22   Contour map of electrical conductivity in the freshwater aquifer.
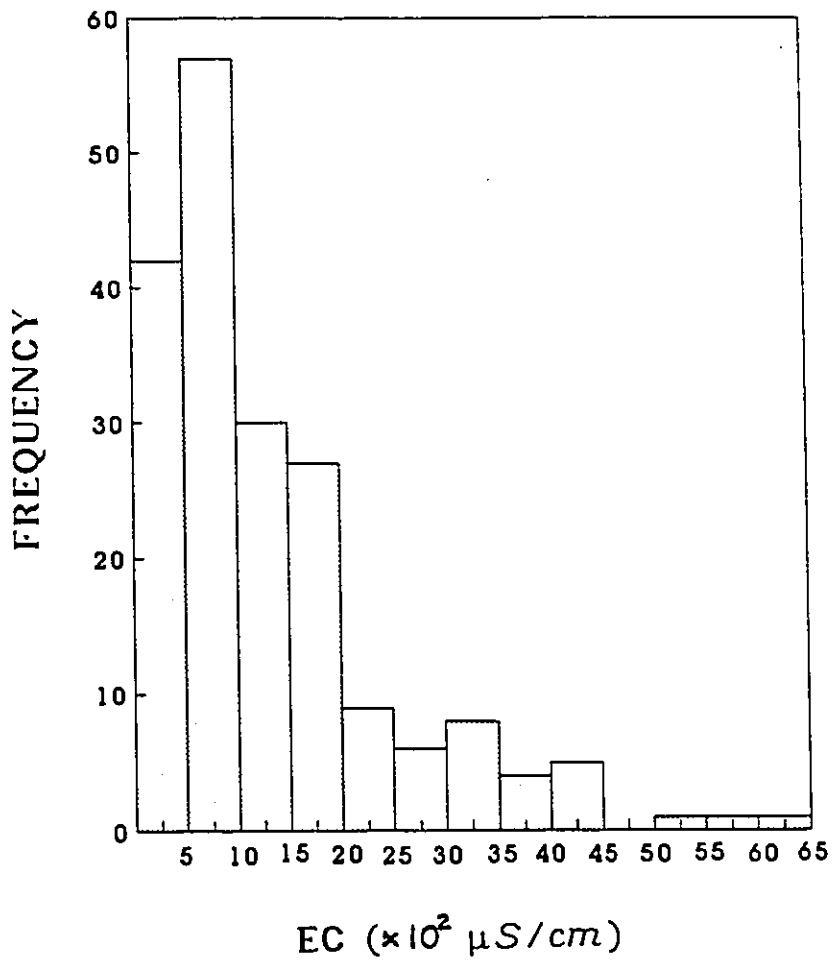
Figure 23   Frequency histogram of electrical conductivity.

recharge area to Lake St. Clair, Kent County and the Detroit River; 5. northeast from the southwestern corner of the study area in Monroe County to the Detroit River and Lake Erie.

Very high EC values of about 24500 $\mu S/cm$ have been reported for two freshwater aquifer wells recently drilled near the town of Belle River in northern Essex County (Osman Ibrahim, personal communication). Less than 10 km south of this location, EC values of 2000 $\mu S/cm$ are common for the freshwater aquifer groundwater. EC of this high a value is indicative of saline water (Freeze and Cherry, 1979) and may reflect exfiltration of deeper formation water into the freshwater aquifer as suggested by Long et al. (1988).

4.0 COMPUTER MODEL INVESTIGATION

4.1 Discrete State Flow Modelling

Isotopic and physical hydrogeology data in an aquifer can provide information on the directions of groundwater flow in an aquifer but these data cannot estimate the rate of groundwater flow in the aquifer. Determining the rate of flow in an aquifer is necessary in predicting the fate of contaminants introduced into an aquifer. Mathematical flow models are convenient methods of determining flow rates in complex aquifer systems.

4.1.1 Flow Models

Typically hydrogeologic flow models rely on head, permeability and porosity data and involve complex finite difference or finite element methods, applied to two-dimensional partial differential equations. These methods model a study area as a large number of small, connected rectangular or triangular "elements". The corners of these elements are referred to as "nodes" and data must be entered for each of these corners. It is unlikely that enough field data can be obtained to act as inputted data for the large number of nodes required for an accurate solution of the mathematics in the finite element and difference methods. The usual practice is to estimate parameters from a general knowledge of the hydrogeology of the modelled area. This estimation of parameters is by far the major source of error in these methods. Often, in an attempt to achieve accurate results, estimated node data is "tweeked" until flow results match known values in the area. Unfortunately, unless there is at least one known value per element, an infinite number of possible flow scenarios can be converged upon. It is then left to the modeller to select a flow scenario which best reflects the modelled area.

Another approach to modelling tracer data in hydrogeologic systems is the use of lumped parameter or discrete state models in which interconnected finite-state mixing cells represent mixing processes and flow conditions. Campana (1975) and Simpson and Duckstein (1976) first introduced this approach in the form of the discrete-state compartment (DSC) model for the interpretation of groundwater flow system patterns. This model utilizes environmental tracers and their distributions in a system represented

by a network of interconnected cells. The model tracks the tracer concentration throughout the system by iterating a recursive mass-balance equation on a discrete time basis.

The DSC model and other such models require prior knowledge of the velocity distributions, usually obtained experimentally or by hydraulic models which describe flow in an aquifer. These DSC type models do not actually yield flow rates. Instead, these models provide a method for judging the accuracy of hydraulic models by using the model's flow velocity distribution as input and solving for a tracer concentration history. The resultant tracer concentration data is then compared against actual field data in order to validate the flow model.

A discrete state mixing cell model has been developed for this study which directly addresses the drawbacks of the above mentioned flow modelling methods. The model does not require massive amounts of data, allowing the exclusive use of field data, yet it yields a "unique" solution of flow rates in a modelled area. This is accomplished by eliminating the solution of flow direction by dividing the modelled area into multiple one-dimensional flow regimes called "subdivisions". These subdivisions are selected based upon knowledge of the hydrogeology of the area and contoured tracer concentration data. When flow direction is eliminated as a variable, the flow can then be solved by tracer concentration history and aquifer volume data alone. This model provides an extremely efficient method of determining flow in an aquifer, which, although simpler mathematically than finite element and difference models, requires less estimated data and ultimately yields a more reliable result.

4.1.2 Discrete State Flow Model (DSFM)

The Discrete State Flow Model (DSFM) determines the rate of flow in an aquifer by discretizing the aquifer into a number of compartments called "cells". The equations of continuity are simultaneously applied to all of the cells in the aquifer and an iterative solution process yields the flow rates between cells otherwise known as the "flow field".

4.1.2.1 Continuity Equation for Water

The first equation used in the flow model is a restated version of the conservation of mass law applied to fluids, commonly known in fluid dynamics as the continuity equation. The continuity equation applied to a cell under the assumption of steady state flow is:

$$\Sigma Q_{in} - \Sigma Q_{out} + \Sigma SBRV - \Sigma SBDV = 0 \qquad (1)$$

where:

Q_{in} the volume of incoming water from other cells in the aquifer per unit time

Q_{out} the volume of water leaving this cell to other cells per unit time

$SBRV$ (System Boundary Recharge Volume) the volume of water entering the cell from outside the aquifer per unit time

$SBDV$ (System Boundary Discharge Volume) the volume of water leaving the cell to outside the aquifer per unit time

These parameters are shown applied to a single cell in a DSFM in Figure 24.

4.1.2.2 Continuity Equation for the Tracer

The second equation used in the model is the continuity equation for the tracer. The tracer concentration varies with time and is therefore presented as a first order ordinary differential equation of the form:

$$\frac{dC}{dt} = \frac{\Sigma Q_{in} C_{in} - \Sigma Q_{out} C + \Sigma SBRV \cdot SBRC - \Sigma SBDV \cdot C}{VOL} \qquad (2)$$

where:

C the concentration of the tracer in the particular cell

$\frac{dC}{dt}$ the change of tracer concentration with time

C_{in} the tracer concentration entering from other cells

$SBRC$ (System Boundary Recharge Concentration) the tracer concentration in the water entering the cell from outside the aquifer
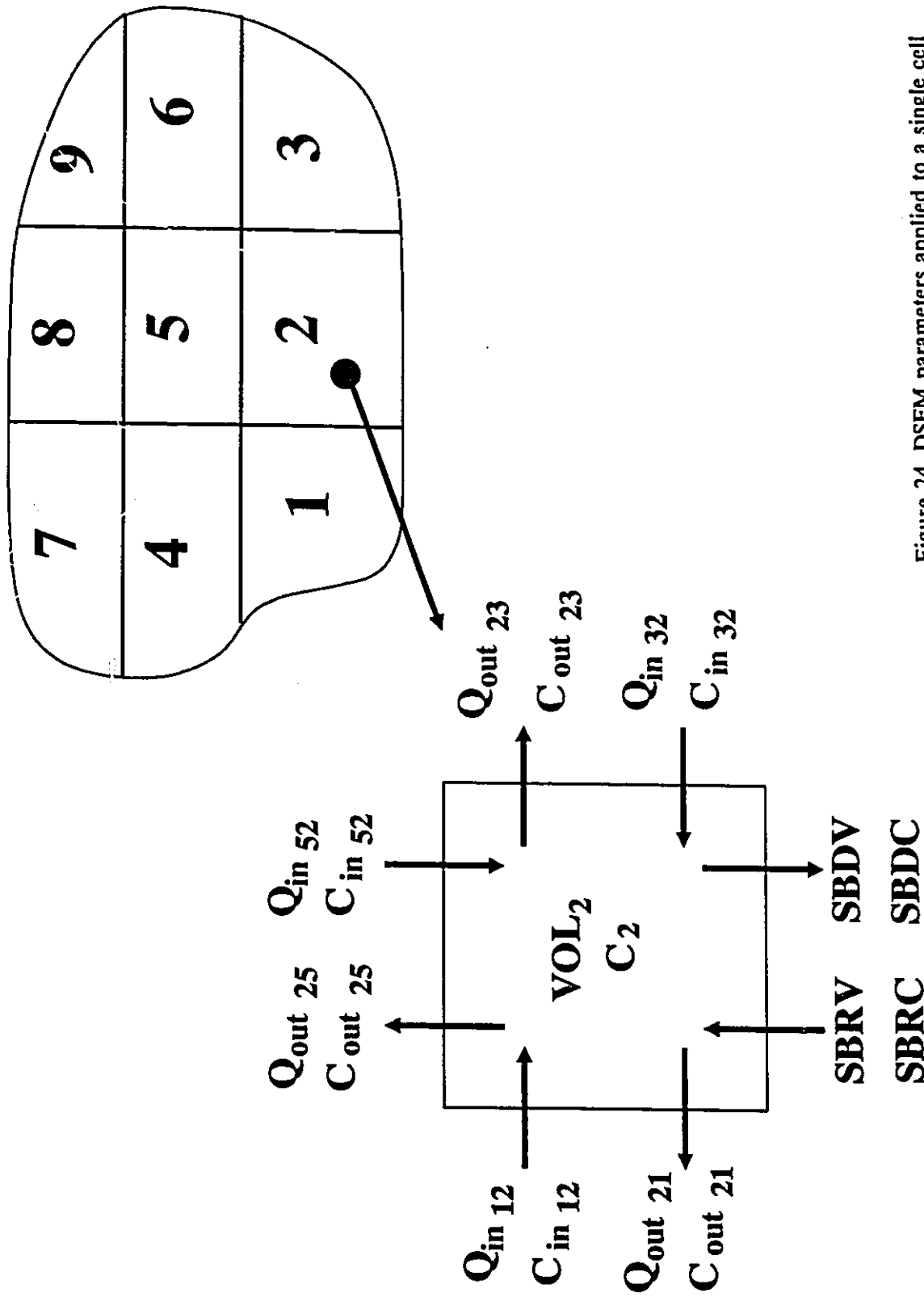
Figure 24 DSFM parameters applied to a single cell

60

SBDC   (System Boundary Discharge Concentration) the tracer concentration in the water leaving the cell to outside the aquifer

VOL   the total porous volume of the cell

In its differential form Equation (2) is of little use in a flow model, but Simple Euler Integration applied to Equation (2) can provide the tracer concentration at any specified time provided an initial concentration is known. Equation (3) is this integration technique expressed mathematically.

$$C(n+1) = C(n) + \frac{dC}{dt}\Delta t \tag{3}$$

where:

n is an increment of time

Δt is the amount of time units per increment

4.1.3 DSFM Applied to an Aquifer

DSFMs can solve only two distinct flow rates per cell which makes it very difficult to solve a flow field where the cells have been selected arbitrarily. For example, Figure 25 is a 16 cell system with a flow field consisting of 80 unknown flow rates. The system can yield two independent equations for each cell, for a total of 32 independent equations, allowing the solution of 32 flow rates. Assuming that incoming concentrations are known and that there is no flow between diagonally adjacent cells, we are left with 48 unsolvable flow rates leaving an infinite number of possible flow scenarios. This simple example would imply that a DSFM is a technique that cannot be readily applied to solving the aquifer flow field. This implication is true unless additional information about the aquifer can be acquired from other data. In this study, isotopic data provide information on the direction of flow in the aquifer and knowledge of the geology of the study area helps delineate the source of recharge to the system and provides the general flow directions within the aquifer. This information helps eliminate the excess number of unknowns and allows us to use a modified form of modelling.
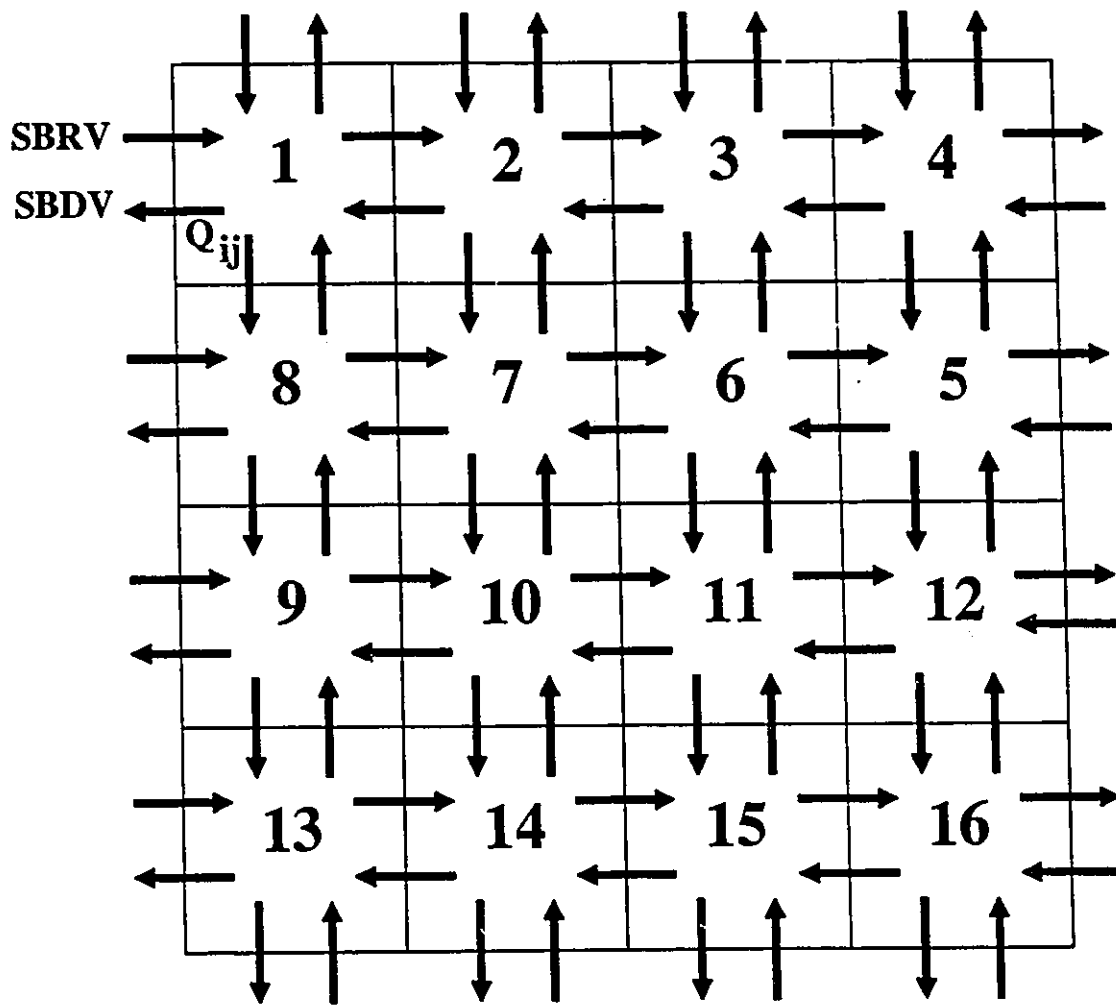
Figure 25   16 cell system showing arbitrary 2-dimensional cell selection.

4.1.4 DSFM Applied to a Study Area

4.1.4.1 Simplification of the DSFM

Figure 26 is a plan view of a hypothetical study area with tracer concentration contours and flow which predominantly run from recharge areas to discharge areas. This study area is divided into smaller subdivisions where isotopic contours are approximately parallel and no discontinuities in contours occur. An assumption can be made that flow between cells in these subdivided areas is one-dimensional and perpendicular to the tracer contours, flowing from the recharge area to the discharge area. Figure 27 illustrates the remaining flow parameters and the equivalent simplified flow model for these subdivisions in plan view. This type of model is the Simplified DSFM (SDSFM).

The equations of continuity are now reduced and presented in the following forms:

continuity of water:

$$Qin_i + Q_{(i-1)i} - (Qout_i + Q_{i(i+1)}) = 0 \tag{4}$$

continuity of tracer concentration:

$$\frac{dC_i}{dt} = \frac{Qin_iCin + Q_{(i-1)i}C_{(i-1)} - (Qout_i + Q_{i(i+1)})C_i}{VOL_i} \tag{5}$$

where:

Qin_i	the volume of recharge water from the environment to cell i per unit time
$Qout_i$	the volume of discharge water from the cell i to the environment per unit time
$Q_{(i-1)i}$	the volume of water entering cell i from the previous cell (i-1) per unit time
$Q_{i(i+1)}$	the volume of water leaving cell i for the next cell (i+1)
C_i	the concentration of tracer in cell i

SUBDIVISION 1

RIVER IN
ALLUVIAL PLAIN

HIGH BEDROCK

SUBDIVISION 2

SUBDIVISION 3

GRAVEL ISLAND
RIDGE

DISCHARGE
AREA

CELL No.

CELL
BOUNDARY

SUBDIVISION
BOUNDARY
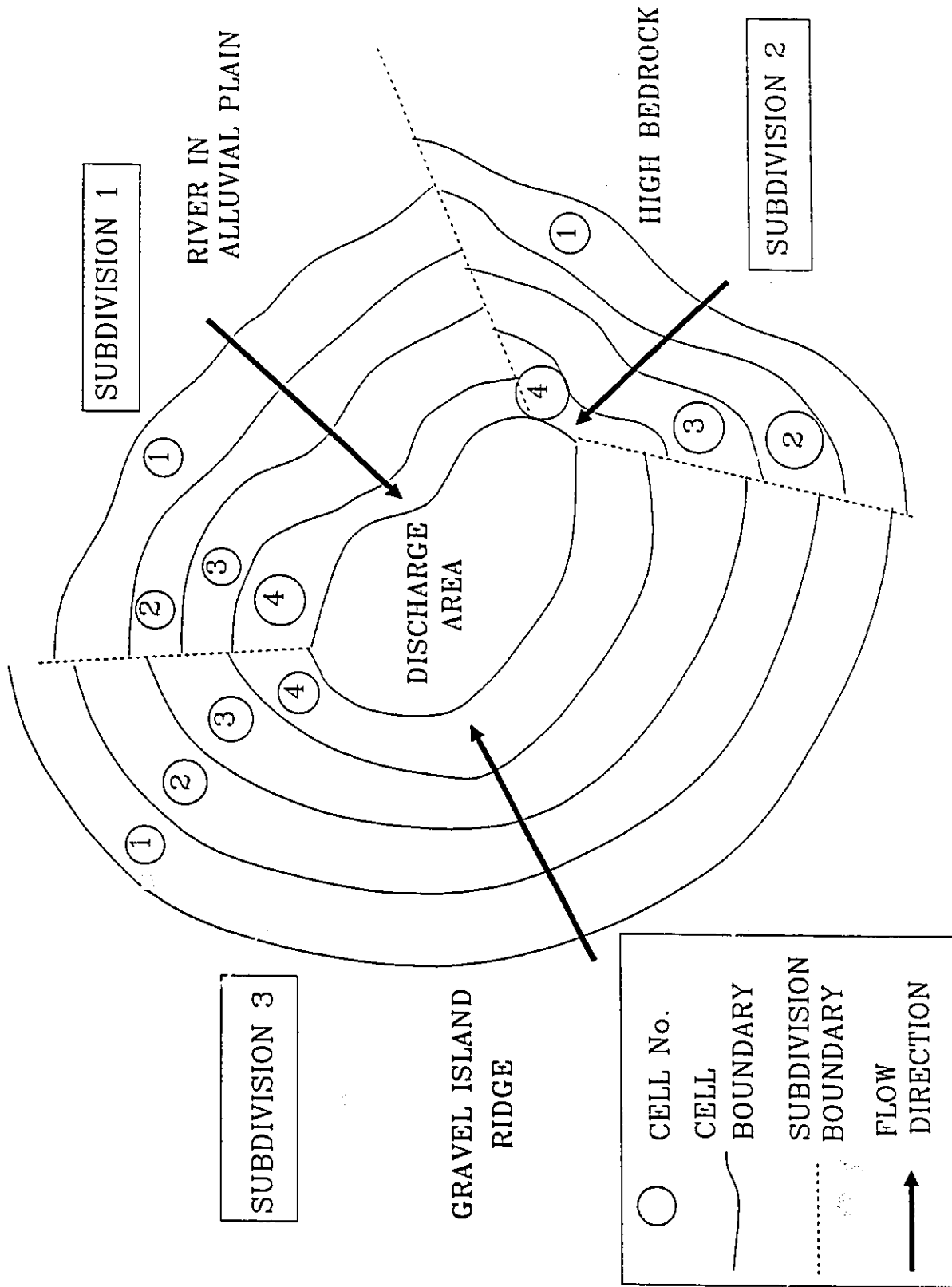
FLOW
DIRECTION

Figure 26   Hypothetical study area.
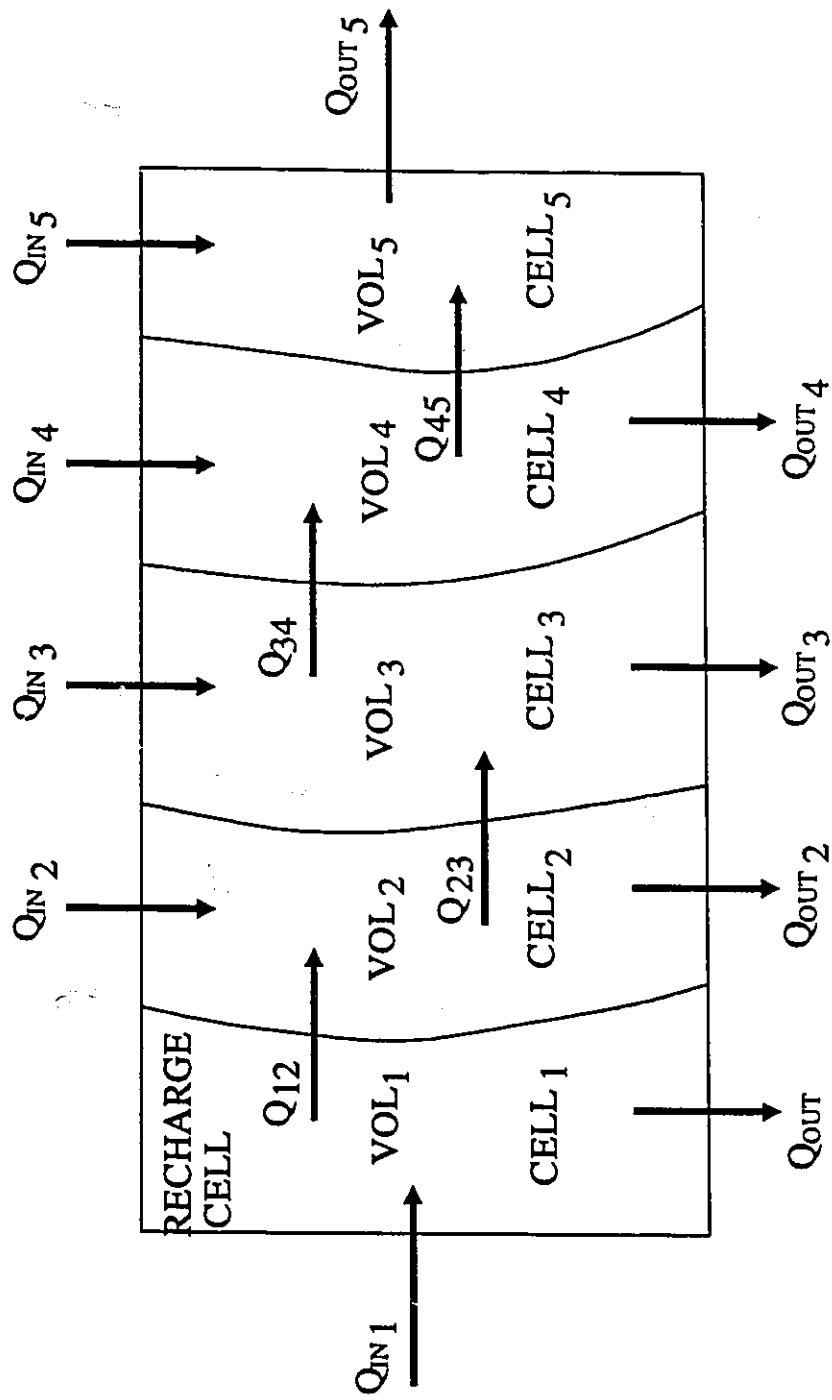
Figure 27   Simplified flow model aquifer subdivision.

Cin_i the concentration of tracer in the recharge water entering cell i

4.1.4.2 Solution of the Simplified DSFM (SDSFM)

The SDSFM, by virtue of the one-dimensional flow assumption, is solved in a cell by cell process starting with the recharge cell and progresses monotonically toward the discharge area. This process will be described in the following sections.

4.1.4.2.1 Solution of Flow in the Recharge Cell

The recharge cell is assumed to be recharged by meteoric water, exclusively. This implies that $Q_{(i-1)(i)}$ is equal to zero. The next step is to regard $Qout_i$ and $Q_{(i)(i+1)}$ as one flow leaving the cell. This simplification has no effect on the governing equations as the brackets in equations (4) and (5) indicate. Iterating various values of Qin_i with equation (4), (5) and (3) will yield Qin_i and the sum of $Qout_i$ and $Q_{(i)(i+1)}$. $Q_{(i)(i+1)}$ is the fraction of Qin_i which flows into the next cell and $Qout_i$ is the fraction which does not. These parameters cannot be separated until the solution of the next cell.

4.1.4.2.2 Solution of Flow in the Intermediate Cells

Initially, an assumption is made that recharge to the current cell in the solution is zero. The only flow into the current cell is assumed to originate from the previous cell. The magnitude of the incoming flow is initially set equal to the total flow rate leaving the previous cell. These assumptions are tested by applying equations (4), (5) and (3) and comparing the final predicted cell tracer concentration against experimental data.

If the assumed flow rate from the previous cell is found to be too large, in other words, if the predicted tracer concentration from the solution of the model is too high, the recharge to the current cell is set equal to zero. Then the incoming flow rate magnitude is reduced and re-tested repeatedly until the predicted final concentration agrees with the current experimental data. The final incoming flow rate is equal to $Q_{i(i+1)}$ for the previous cell. Q_{out} for the previous cell can now be determined by the following:

$$Qout_{i-1} = Qin_{(i-1)} + Q_{(i-2)(i-1)} - Q_{(i-1)(i)} \tag{6a}$$

If the assumed flow rate from the previous cell is found to be too small the incoming flow rate from the previous cell is set at the assumed value and a recharge flow rate is introduced. Different values of recharge flow rate are tested by comparing the models predicted tracer concentration to the current experimental data until they match. The flow rate coming from the previous cell is equal to $Q_{i(i+1)}$ for the previous cell, since all flow from the previous cell is entering the current cell $Qout_i = 0$ for the previous cell.

4.1.4.3 Conditions on the Use of the Simplified DSFM

The simplified DSFM provides a method to estimate rates of discharge, recharge and groundwater flow in an aquifer, but the following assumptions must be considered reasonable when the model is applied:

1) Flow is one-dimensional in nature and progresses from cell to cell.

2) Flow is steady state.

3) Initial and final tracer concentrations are known for each cell.

4) A tracer concentration history is known for recharge to the aquifer from outside the aquifer.

5) The first cell in a subdivision receives incoming flow from a single source with known concentration history, exclusively.

4.1.5 Input for the Simplified DSFM Computer Program

A program has been written such that once the following data are input, the solution is automatically performed:

1) individual cell volume

2) tracer concentration history

3) initial and final cell concentrations

4) the total time elapsed

The output from the program is a table of the solved parameters (the variables illustrated in Figure 27). Figure 28 is a flow chart explaining the program and a listing is provided in Appendix III.

initialize:

cell volumes
Cfinal,Cinitial
tracer history
time (T)

solve:

cell 1:
call ITERATE

for i=2 to Ncells
call ITERATE

if $\dot{V}ij(i-1)>\dot{V}in(i)$
$\dot{V}out(i-1)=\dot{V}ij(i-1)-\dot{V}in(i)$

output:

for i=1 to N
$\dot{V}in(i)$
$\dot{V}out(i)$
$\dot{V}ij(i)$

SUB ITERATE
iterate:
$\dot{V}in(i)$, $\dot{V}out(i)$
call Euler
until C(T)=Cfinal(i)

SUB EULER
for i=1 to T
$C(i)=C(i-1)+\dfrac{dC}{dt}$
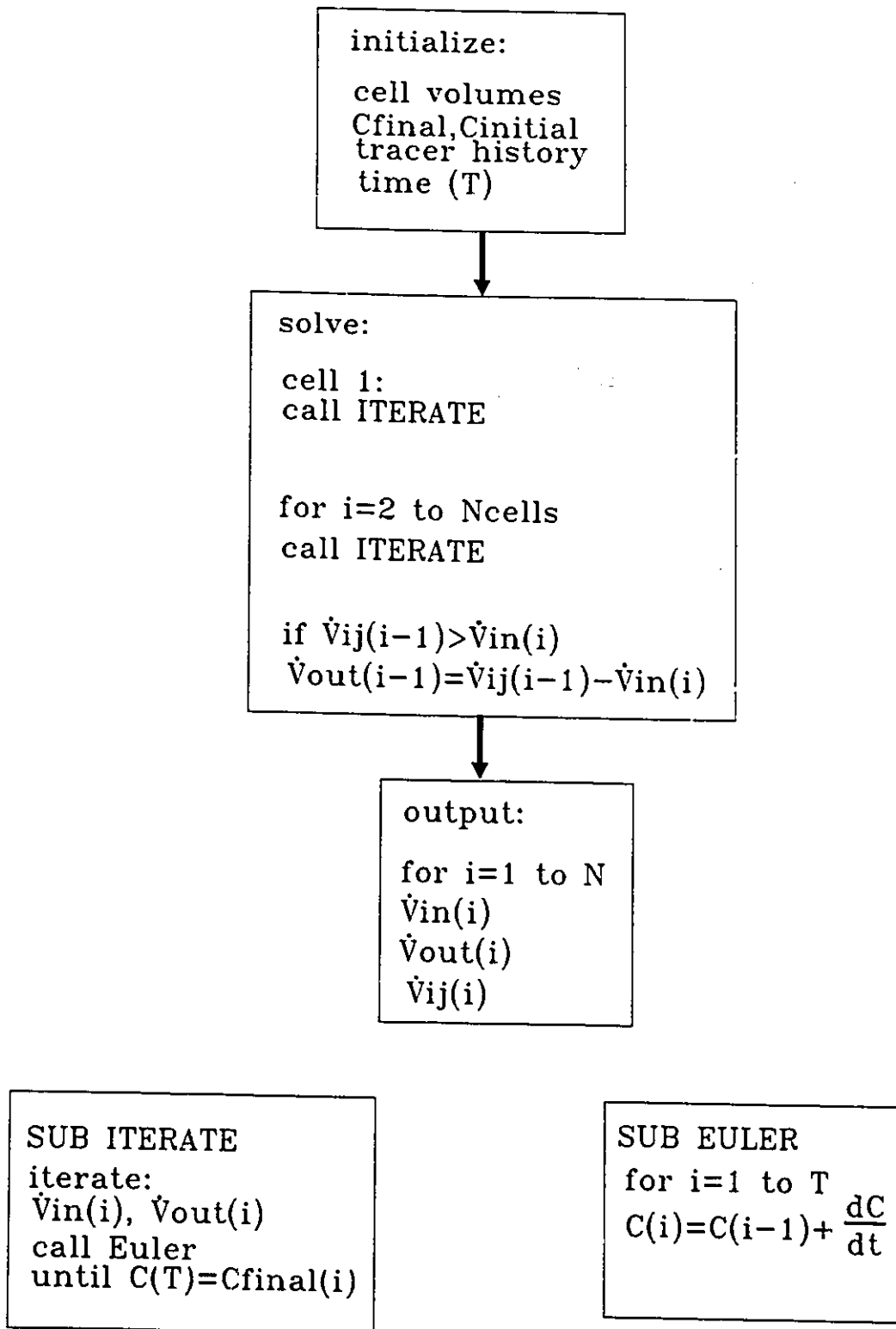
Figure 28   Flow chart of the SDSFM.

4.2 Modelling Methods

4.2.1 Selection of Subdivisions and Cells

The division of the study area into smaller sections, which will be referred to as "subdivisions" in this report, was based on the direction of regional groundwater flow determined from the physical hydrogeology of the study area and the isotopic content of groundwater within the freshwater aquifer. The subsequent separation of subdivisions into smaller cells was based on the positions of the contours of $\delta^{18}O$ fronts from Figure 14. Within each subdivision, the first cell, cell 1, represents the recharge area of the entire subdivision. The size and extent of the recharge cell was also based upon the surficial geology (Figure 7), bedrock topography (Figure 3), surficial topography (Figure 2) and drift thickness (Figure 9) in addition to the position of the $\delta^{18}O$ front contours. The last cell within each subdivision is located at or near the discharge area for each subdivision.

The study area has been separated into four subdivisions: subdivision 1 includes the counties of Oakland, Macomb and St. Clair; subdivision 2 includes Lambton County and the northern third of Kent County; subdivision 3 is comprised of the southern 2/3 of Kent County; and subdivision 4 is located within Essex County. The southwestern portion on the study area, shown as Region 5 on Figure 14, was not modelled as a subdivision because its southern and western limits are located outside of the study area and its northern limits could not be defined due to the lack of data points in the Detroit Metropolitan area. Figure 29 shows the location of the study area subdivisions and the position of cells within each one. The selection of these subdivisions will subsequently be discussed.

SUBDIVISION 1 - This subdivision was chosen to represent the flow of groundwater through the freshwater aquifer from the large interlobate region recharge area on the Michigan side of the study area. This subdivision has a surface area of 4137 km² and it contains eight cells.
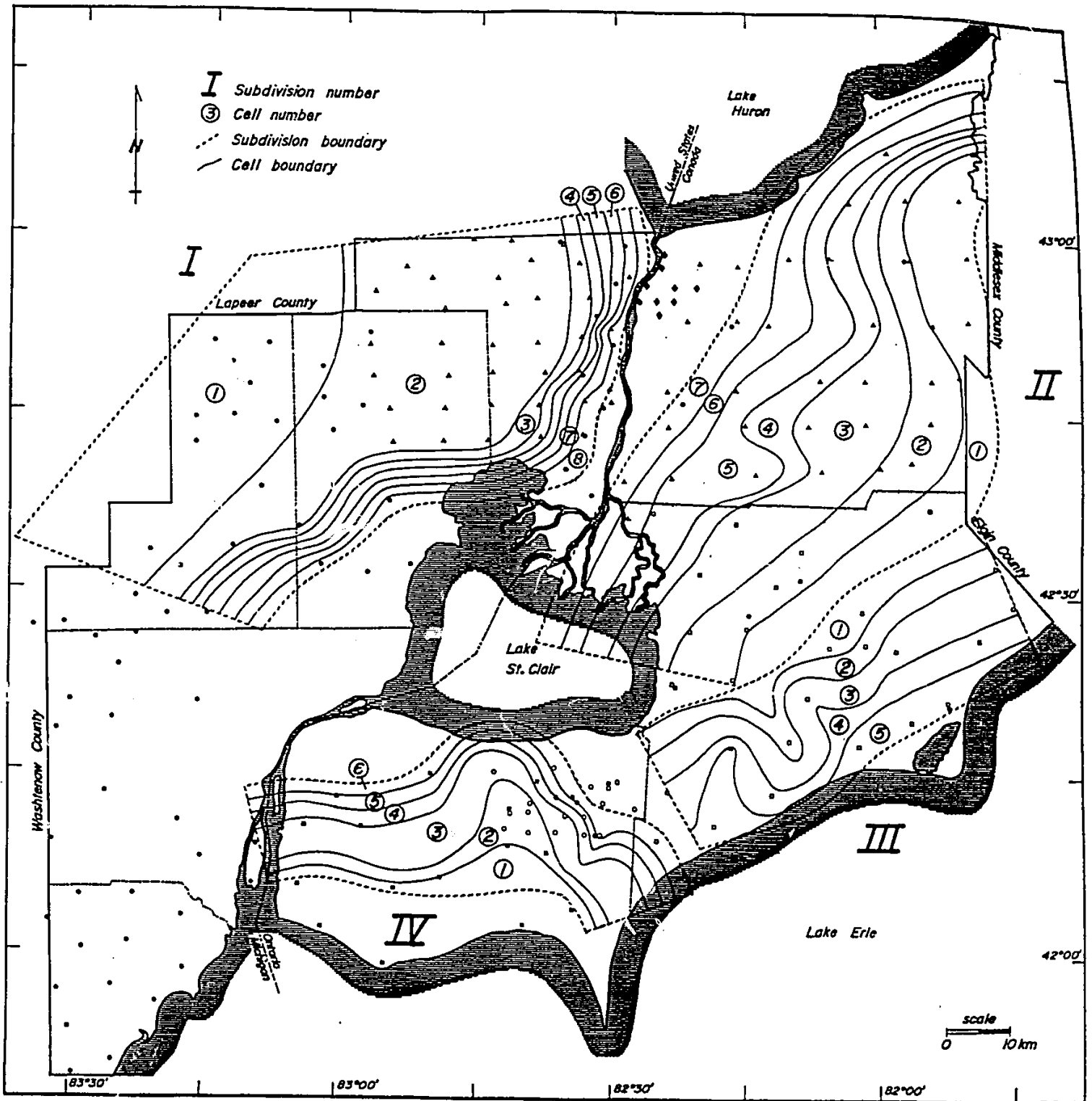
Figure 29   Location of SDSFM subdivisions and their respective cells.

Examination of the potentiometric surface, bedrock topography, topography and $\delta^{18}O$ contours imply that flow within the freshwater aquifer is southeastward from the bedrock and topographic divides which follow a northeast-southwest trend along the northwest edge of Oakland County. This line therefore will also serve as a boundary for subdivision 1 as it represents the division between flow to the northwest and southeast. The northern boundary of subdivision 1 is simply represented by the northern limit of the study area. The southern boundary of the subdivision is set just above the Oakland-Wayne County border due to the paucity of samples in northeastern Wayne County as well as to the lack of samples with depleted $\delta^{18}O$ values in the western third of the county. The discharge boundary was set outside the $-16^o/_{oo}$ $\delta^{18}O$ front which runs along the western edge of the St. Clair River then swings west across the northern bay of Lake St. Clair and the southern portion of Macomb County.

SUBDIVISION 2 - The largest subdivision of the study area has a surface area of 4214 km² and amalgamates the flow pattern of two separate recharge areas. It includes the western flow of groundwater from the recharge area in eastern Lambton County and the northwestern flow of water inferred from $\delta^{18}O$ contours from the Thames River recharge area in Kent County.

The recharge boundary is represented by the Thames River and the approximate bedrock/groundwater divide along the eastern edge of Lambton County. The southern edge is restricted by the lack of $\delta^{18}O$ values within the geologic material beneath Lake St. Clair. The discharge boundary was set within the zone of greatest ^{18}O depletion; through the center of the apparently converging $-17^o/_{oo}$ $\delta^{18}O$ fronts. Past this front the discharge boundary was extended into the northeast corner of Lambton County guided by the position of the enriched $\delta^{18}O$ fronts found upgradient. Subdivision 2 has been partitioned into seven cells.

SUBDIVISION 3 - This subdivision was chosen to represent the southeastern flow of groundwater, implied by $\delta^{18}O$ contours, that is being recharged by the Thames River. The recharge boundary is delineated by the entire course of the Thames River which runs through Kent County. The western boundary of the subdivision was kept within Kent County due to the northeastward movement of a converging groundwater flow

system that originates southwest of subdivision 3. The discharge boundary is represented by the Lake Erie shoreline. This subdivision covers an area of 1524 km² and contains five cells.

SUBDIVISION 4 - This is the smallest subdivision, having a surface area of 1388 km², and it encompasses the groundwater flow system that is being recharged by the Leamington Moraine recharge area located in southern Essex County. The recharge boundary was centered over the Leamington Moraine bedrock high and was extended westward, approximately across the divide of a bedrock ridge located in southwestern Essex County, in order to accommodate the position of enriched $\delta^{18}O$ fronts. The western side boundary was terminated along the Detroit River due to the uncertain distribution of $\delta^{18}O$ values in southeastern Wayne County. The discharge boundary forms the approximate extension of the -16.5 $^o/_{oo}$ $\delta^{18}O$ contour front if it were to extend out into Lake Erie and follow the countenance of the less depleted $\delta^{18}O$ contours situated upgradient to it. This subdivision has been separated into six cells.

4.2.2 Determination of Subdivision Time Span

The concentration of ^{18}O has been found to be directly related to temperature (Dansgaard, 1964). Paleoclimatic conditions at the end of the Wisconsinan glacial period were much colder than at present (McAndrews, 1981). Therefore we must assume that the ^{18}O concentration of water within the freshwater aquifer emplaced during the glacial period must have been more depleted than that of meteoric water presently recharging the study area.

The most negative $\delta^{18}O$ values found within the study area are about -18.5$^o/_{oo}$. Desaulniers et al (1981) found $\delta^{18}O$ values which ranged from -16 to -18$^o/_{oo}$ for deep groundwaters extracted from points at or near the glacial bedrock interface at Sarnia and Wyoming. The basal till unit emplaced during the Port Bruce glacial advance (about 15000 a B.P.) contains shells and mollusks whose $\delta^{18}O$ content Fritz et al. (1975) measured to be between -18 and -20$^o/_{oo}$. Since the most negative values within the study area (sample numbers: 71=-18.46$^o/_{oo}$, JB-21=-18.22$^o/_{oo}$, JB-22=-18.17$^o/_{oo}$) correlate fairly well with the $\delta^{18}O$ values of these studies, it is fair to assume that the

freshwater aquifer contained water of at least a value of $-18.5°/_{oo}$ at the end of the Wisconsinan glacial period. Therefore the initial concentration of $\delta^{18}O$ to be entered into the SDSFM will be $-18.5°/_{oo}$.

There has been an ameliorating climatic trend from the time of the last glacial retreat to the present which has resulted in an enrichment of ^{18}O in the precipitation of the study area. Therefore the temporal variation in $\delta^{18}O$ values of meteoric water is required as input for the flow simulation.

4.2.2.1 Correlation of Temporal $\delta^{18}O$ Variation of Precipitation to Study Area

The record of temporal variation of $\delta^{18}O$ for the Brampton area obtained from the cellulose studies of Edwards and Fritz (1986, 1988), listed in Table 1 (section 3.1.1.2), will be utilized to extract the corresponding temporal $\delta^{18}O$ variation for each subdivision in the study area.

The Brampton study area of Edwards and Fritz (1986, 1988) is further north than this study area. Modern precipitation of the Lake St. Clair study area will therefore be slightly enriched in ^{18}O relative to the Brampton area, reflecting the higher mean annual temperature of the Lake St. Clair area. It is likely that a similar offset in $\delta^{18}O$ values existed in the past (Edwards, personal communication, 1990).

Edwards and Fritz (1986) determined from data in Bryson and Hare (1974) and the International Atomic Energy Agency (1981) that the gradient of $\delta^{18}O$ change with mean annual temperature for meteoric waters in the Great Lakes region is about $0.65°/_{oo}/°C$. This gradient will be assumed to have remained constant over time (at least from the end of the last glacial period) and will be used to determine the temporal variation of $\delta^{18}O$ for each subdivision in the study area based on modern mean annual temperatures.

Environment Canada (1982) records show that Brampton has a mean annual temperature of $7.1°C$ and Edwards and Fritz (1986) measured the $\delta^{18}O$ value of the area's precipitation to be $-11.0°/_{oo}$. The mean annual temperature of the recharge area in subdivision 1 was found to be $8.7°C$, determined from the average of temperature records of two stations of the United States Weather Bureau (U.S. Department of Commerce, 1978). The difference in temperature between this subdivision and Brampton is $+1.6°C$. From the isotope/temperature gradient of $0.65°/_{oo}/°C$, the

difference in isotopic value is therefore $+1.0°/_{oo}$ which suggests that precipitation within the recharge area of subdivision 1 is presently $1.0°/_{oo}$ heavier than that at Brampton. Assuming that a similar $\delta^{18}O$ offset has existed in the past, the value of $+1.0°/_{oo}$ will be added to each isotope value in Table 1. Table 2 summarizes the following for each subdivision: mean annual temperature (T) in each recharge area, the difference in mean annual temperature (ΔT) between each recharge area and the Brampton area, and the difference in $\delta^{18}O$ of meteoric water ($\Delta \delta^{18}O$) between each recharge area and Brampton.

The addition of the $\Delta \delta^{18}O$ for each subdivision area to the $\delta^{18}O$ values of meteoric water of the Brampton area in Table 1 gives the temporal variation of $\delta^{18}O$ in meteoric water recharging into each subdivision from a period beginning 11450 a B.P. However glacial studies of the Great Lakes Basin (Farrand and Eschman, 1974; Eschman, 1985; Eschman and Karrow, 1985; Calkin and Feenstra, 1985) indicate that the study area had been ice-free prior to 11450 a B.P. It must be determined when each subdivision's recharge area was wholly uncovered (i.e. no glacial ice or proglacial lakes). In other words, how many years ago did meteoric water begin to infiltrate each recharge area?

4.2.2.2 Timing of Subdivision Recharge

When the continental glacier began to recede for the last time, most of the study area was covered with the resultant meltwaters which formed proglacial lakes thereby covering the subdivision recharge areas. This scenario prevented the infiltration of water from the recharge areas into the freshwater aquifer due to hydrostatic pressure conditions (i.e. pressure head is the same above recharge and discharge areas) which would prevent the flow of water due to lack of head difference. Since there was no flow when the recharge areas were covered by glacial lakes, the time at which these waters receded from the recharge areas is the point in time when meteoric waters began to infiltrate these areas. This would also mark the time when the flow model simulations should begin.

For this study, the comparison of recharge area elevations to glacial lake elevations does not take into account the effects of glacial rebound on the timing of recharge or

Subdivision No.	Recharge Area T ($^{\circ}C$)	ΔT ($^{\circ}C$)	$\Delta \delta^{18}O$ ($^{\circ}/_{\infty}$)
1	8.7	1.6	+1.0
2	8.3	1.2	+0.8
3	9.1	2.0	+1.3
4	9.4	2.3	+1.5

Table 2: Summary of steps to obtain $\Delta\delta^{18}O$ for each subdivision area.

the variation in flow direction from present day. It will be assumed that flow directions and flow system boundaries have always been as they are now and that hydraulic gradients have been constant since flow started.

The length of recharge time for each subdivision was based upon the Lake Erie basin glacial lake sequence of Figure 30. Recharge area elevations for each subdivision were compared to glacial lake elevations to determine the time at which lake levels had dropped below the elevation of the recharge areas.

SUBDIVISION 1 - The recharge area of this subdivision, which ranges in elevation from about 240 to 340 m.a.s.l., was the first part of the study area to become exposed during glacial ice retreat. The area's topography is higher than the level of Middle Lake Maumee (III) and was ice covered during Highest Lake Maumee (I), therefore recharge of meteoric waters likely began at the transition of Maumee I to Maumee III lake levels. Figure 30 indicates that this transition occurred approximately 14250 a B.P. and will mark the length of time that meteoric water has infiltrated subdivision 1's recharge area.

SUBDIVISION 2 - The glacial beaches of Early Lake Arkona which surround the Wyoming Moraine, denote the exposure of the northern part of the recharge area of this subdivision. These beaches are located between the 215 m and 225 m topographic contours (Chapman and Putnam, 1984) which Figure 30 confirms were formed between Early Lake Arkona levels and Arkona II levels about 13750 a B.P. Figure 30 also shows that Lake Whittlesey inundated most of this recharge area about 13000 a B.P. However, it cannot be ignored that during the Arkona and Ypsilanti lake phases the recharge area was all or at least partially exposed. Since the Whittlesey lake phase was very brief (<200 years) it will be disregarded with respect to the time span of the flow modelling of this subdivision and it will be assumed that recharge of meteoric water has been continuous for 13750 years.

SUBDIVISION 3 - The elevation of the recharge area in this subdivision is substantially lower than those in subdivisions 1 and 2. This indicates that glacial lakes prevented meteoric water from infiltrating this recharge area for a longer period of time.
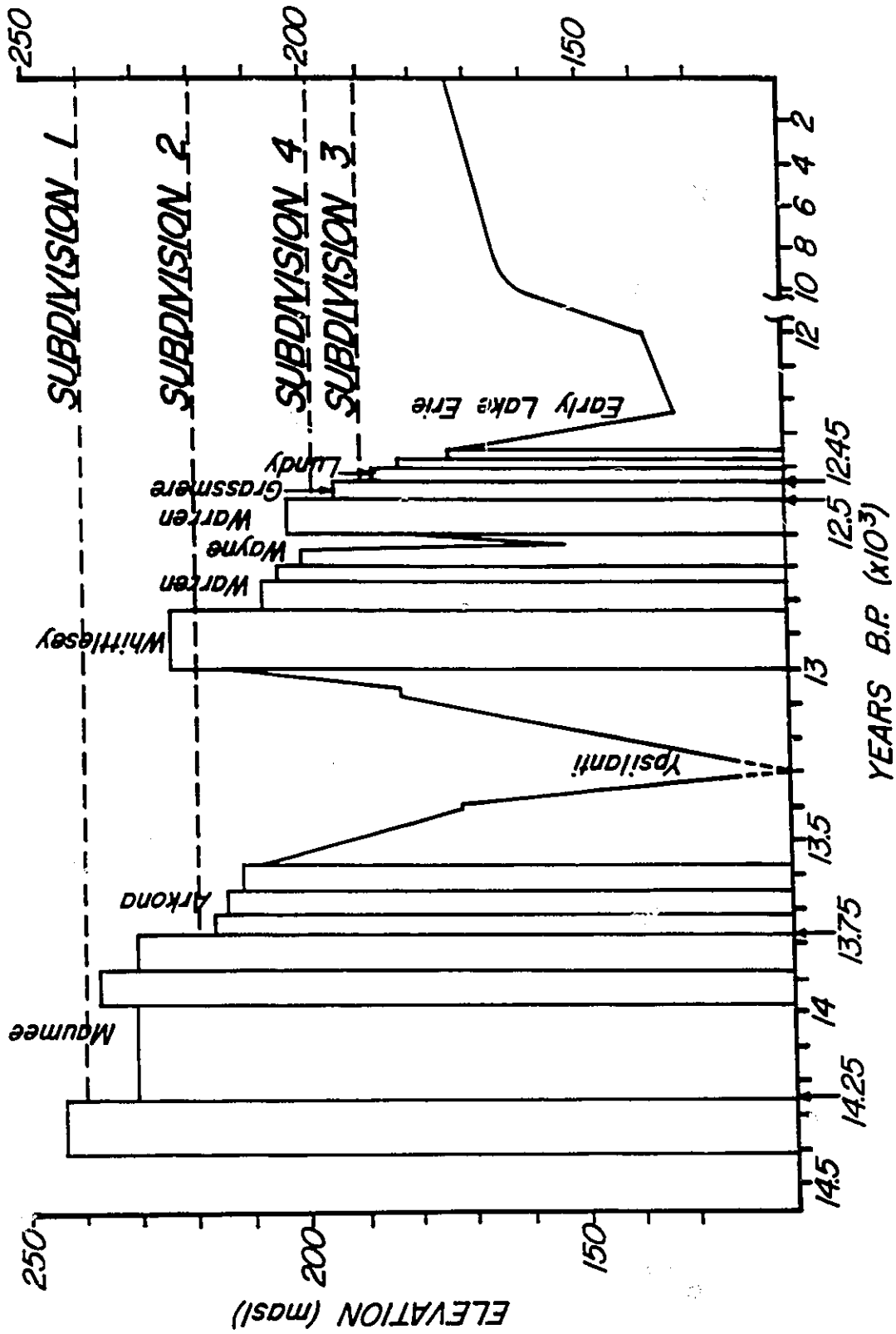
Figure 30   Sequence of glacial lakes in the Lake Erie basin (after Calkin and Feenstra, 1985) showing estimate of subdivision recharge timing.

Recharge area 3 became substantially exposed with the recession of Lake Grassmere from about the 190 m elevation level. This correlates to a recharge period beginning about 12450 a B.P. (from Figure 30).

SUBDIVISION 4 - Warren beaches present in the eastern part of the recharge area (Vagners, 1972) indicate that the elevated area around the Leamington Moraine (~195 m elevation level) was substantially exposed during the Lake Warren phase. That elevation corresponds to the end of the Lake Warren phase receding to Lake Grassmere levels which Figure 30 shows to have occurred about 12500 a B.P. This is the length of time that meteoric water is assumed to have infiltrated the recharge area.

Table 3 is a record of the temporal variation of $\delta^{18}O$ in precipitation for each subdivision inferred from Edwards' data in Table 1. The isotopic data in Table 1 was manipulated for each subdivision by adding the appropriate $\Delta\delta^{18}O$ value from Table 2 to each $\delta^{18}O$ value in Table 1 and linearly projecting backwards the $\delta^{18}O$ content of the meteoric waters until each subdivision has an initial $\delta^{18}O(C_{in})$ value of -18.5°/$_{oo}$ in the length of recharge time determined for each subdivision.

4.2.3 Determination of Cell Volumes

The effective volume of each subdivision cell was determined by multiplying each cell's surface area by the average thickness of the freshwater aquifer in that cell and then multiplying that volume by an appropriate value of effective porosity.

The estimation of cell aquifer thickness was determined by averaging the thickness of the freshwater aquifer (sand/gravel/fractured bedrock) in a number of wells which sufficiently cover each cell area. This information was obtained from water well records.

Values of volumetric porosity, n, were estimated for each subdivision. Based on ranges of porosity values from Freeze and Cherry (1979), it was assumed that a sand/gravel mixture has porosity values that range from 25 to 45%. Intera (1987) estimated the porosity of weathered and fractured shale bedrock to range between 5 and 20%. Considering these two porosity ranges, a porosity value of 30±5% was considered to be representative of the freshwater aquifer material where underlain by shale bedrock. The value of n=0.31 was chosen to represent the value of porosity for the freshwater

YEARS BEFORE PRESENT (ybp)	$\delta^{18}O$ of BRAMPTON METEORIC WATER ($^o/_{oo}$)	$\delta^{18}O$ of SUBDIVISION 1 METEROIC WATER ($^o/_{oo}$)	$\delta^{18}O$ of SUBDIVISION 2 METEROIC WATER ($^o/_{oo}$)	$\delta^{18}O$ of SUBDIVISION 3 METEROIC WATER ($^o/_{oo}$)	$\delta^{18}O$ of SUBDIVISION 4 METEROIC WATER ($^o/_{oo}$)
MODERN	-11.1	-10.1	-10.3	-9.8	-9.6
1500	-10.7	-9.7	-9.9	-9.4	-9.2
3850	-9.3	-8.3	-8.5	-8.0	-7.8
4000	-8.9	-7.9	-8.1	-7.6	-7.4
6300	-8.9	-7.9	-8.1	-7.6	-7.4
7510	-10.9	-9.9	-10.1	-9.6	-9.4
7700	-10.9	-9.9	-10.1	-9.6	-9.4
7900	-12.9	-11.9	-12.1	-11.6	-11.4
8700	-12.7	-11.7	-11.9	-11.4	-11.2
9550	-14.1	-13.1	-13.3	-12.8	-12.6
10450	-13.6	-12.6	-12.8	-12.3	-12.1
10950	-14.4	-13.4	-13.6	-13.1	-12.9
11450	-15.2	-14.2	-14.4	-13.9	-13.7
11950		-15.0	-15.3	-16.2	-16.1
12450		-15.8	-16.2	-18.5	-18.5
12950		-16.6	-17.1		(12500 ybp)
13450		-17.4	-18.0		
13950		-18.2	-18.5		
14250		-18.5	(13750 ybp)		

Table 3: Length of recharge time and estimated record of temporal variation of $\delta^{18}O$ in precipitaion for each subdivision.

aquifer within subdivision 1. The computer simulation results of Hyde (1987) indicate that the freshwater aquifer within Lambton County has a porosity of 35%. This value will be assumed for the aquifer in subdivision 2. Similar modelling by Pettapiece (1988) within Kent County resulted in a porosity of 34%. Subdivisions 3 and 4 will be represented with a porosity of 0.34.

Table 4 is a summary of each cell's surface area, average freshwater aquifer thickness and the resultant effective cell volume given the respective porosity value of each subdivision.

4.2.4 Determination of Cell Concentrations

The concentration of ^{18}O in a subdivision cell (i) was determined by taking the mean of all the $\delta^{18}O$ values taken from each sample site found within that cell. Table 5 summarizes the $\delta^{18}O$ values for each cell.

4.2.5 Parameter Input Summary

Data which are entered into the Simplified DSFM for each subdivision are:

1. Cell volumes from Table 4.
2. Final cell concentration from Table 5.
3. Initial concentration of $-18.5°/_{oo}$.
4. Meteoric $\delta^{13}O$ history from Table 3.

4.3 Simplified DSFM Results and Interpretation

4.3.1 Freshwater Aquifer Flow Rates

Each cell's present $\delta^{18}O$ value and volume, obtained from Tables 5 and 4 respectively, and the temporal variation of $\delta^{18}O$ in precipitation for each subdivision from Table 3 were entered into the Simplified DSFM. The resultant flow rates are presented in Table 6.

CELL #	SUBDIVISION 1 (n=0.31)			SUBDIVISION 2 (n=0.35)		
	Area (x10⁸m²)	Aquifer Thickness (m)	Volume (x10⁸m³)	Area (x10⁸m²)	Aquifer Thickness (m)	Volume (x10⁸m³)
1	13.81	1.90	8.14	8.11	1.58	4.48
2	16.52	1.77	9.07	7.38	1.33	3.42
3	2.14	1.37	0.91	7.46	1.17	3.06
4	1.65	1.52	0.78	4.07	1.22	1.74
5	1.62	2.10	1.05	6.17	1.23	2.66
6	1.75	2.03	1.10	4.57	1.07	1.71
7	1.75	1.83	0.99	4.40	1.524	2.35
8	2.14	1.9	1.26			

CELL #	SUBDIVISION 3 (n=0.34)			SUBDIVISION 4 (n=0.34)		
	Area (x10⁸m²)	Aquifer Thickness (m)	Volume (x10⁸m³)	Area (x10⁸m²)	Aquifer Thickness (m)	Volume (x10⁸m³)
1	2.20	1.22	0.91	1.91	1.80	1.17
2	2.59	1.50	1.32	2.01	1.50	1.02
3	4.14	1.25	1.76	4.37	1.60	2.40
4	3.11	1.56	1.65	1.91	1.12	0.73
5	3.20	1.30	1.42	1.67	1.40	0.79
6				2.02	1.33	0.91

Table 4:   Cell size summary given: subdivision porosity 'n', average freshwater aquifer thickness and volume.

CELL #	CELL $\delta^{18}O$ (‰)			
	SUBDIVISION 1	SUBDIVISION 2	SUBDIVISION 3	SUBDIVISION 4
1	-9.9	-10.16	-10.15	-9.5
2	-10.2	-9.03	-11.75	-10
3	-11	-9.8	-14.25	-13
4	-12	-11.11	-16.25	-14
5	-13	-13.54	-17.75	-15
6	-14	-15		-16
7	-15	-17.22		
8	-16			

Table 5: Cell $\delta^{18}O$ values

SUBDIVISION 1

CELL #(i)	$Qi(i+1)_i$ (m³/yr.)	Qin_i (m³/yr.)	$Qout_i$ (m³/yr)	BOUNDARY LENGTH (m)	$Qi(i+1)$ (m3/yr/m)
CELL 1	91749.5	91749.5	0	68040	1.348
CELL 2	52729.8	45521.9	84541.8	102240	0.516
CELL 3	47054.3	0	5675.5	105840	0.445
CELL 4	47054.3	0	0	107280	0.439
CELL 5	51023.2	3969.0	0	107840	0.474
CELL 6	53637.6	2614.4	0	108360	0.495
CELL 7	53883.9	246.2	0	108360	0.497
CELL 8	54394.5	510.7	0	110160	0.494

SUBDIVISION 2

CELL #(i)	$Qi(i+1)_i$ (m³/yr.)	Qin_i (m³/yr.)	$Qout_i$ (m³/yr)	BOUNDARY LENGTH (m)	$Qi(i+1)$ (m3/yr/m)
CELL 1	105115.6	105115.6	0	113760	0.924
CELL 2	111051.6	145473.5	139537.5	115560	0.961
CELL 3	95103.3	0	15948.1	120240	0.791
CELL 4	95103.3	0	0	130320	0.730
CELL 5	55844.3	7698.0	46957.2	120240	0.464
CELL 6	54062.3	0	1781.9	121320	0.446
CELL 7	54062.3	0	0	123840	0.437

SUBDIVISION 3

CELL #(i)	$Qi(i+1)_i$ (m³/yr.)	Qin_i (m³/yr.)	$Qout_i$ (m³/yr)	BOUNDARY LENGTH (m)	$Qi(i+1)$ (m3/yr/m)
CELL 1	18222.9	18222.9	0	73440	0.248
CELL 2	26732.6	8509.7	0	82080	0.326
CELL 3	30647.8	3915.1	0	82080	0.373
CELL 4	17835.2	607.4	13420.0	83520	0.214
CELL 5	17835.2	0	0	65520	0.272

SUBDIVISION 4

CELL #(i)	$Qi(i+1)_i$ (m³/yr.)	Qin_i (m³/yr.)	$Qout_i$ (m³/yr)	BOUNDARY LENGTH (m)	$Qi(i+1)$ (m3/yr/m)
CELL 1	28449.6	28449.6	0	64080	0.444
CELL 2	37677.4	9227.8	0	79200	0.476
CELL 3	43529.2	5851.8	0	85680	0.508
CELL 4	43928.7	399.5	0	86400	0.508
CELL 5	44486.2	557.5	0	89640	0.496
CELL 6	44868.7	382.5	0	91440	0.491

Table 6: Summary of SDSFM flow rate results

4.3.1.1 Subdivision 1

The computed cell to cell flow rates ($Q_{i(i-1)}$) range from a high of 91749.5 m³/yr

entering the recharge cell (cell 1) to a low of 47054.3 m³/yr. The average continuous

flow rate through this subdivision can be estimated at approximately 56400 m³/yr.

Excluding the flow rate from cell 1, the average continuous flow rate through cells 2 to

8 drops to the more representative rate of about 51400 m³/yr.

The model shows a large Q_{out} term associated with cell 2 indicating that a large

portion of flow is leaving this cell. This is not unexpected since ¹⁸O concentration

gradients are slight over a large area and volume in cells 1 and 2 and then abruptly

increase in cells 3 to 8 which are correspondingly much smaller cells. This directly

implies that high flow rates exist through cells 1 and 2 which drop sharply upon entering

cell 3. To accommodate this sudden drop, flow from cells 1 and 2 must be leaving the

freshwater aquifer system to either the overlying drift or the bedrock beneath or perhaps

leaving the subdivision area altogether. The self-correcting properties of this model

manages the change from high flow to low flow cells through the Q_{out} term. It must be

kept in mind that most of the non-representative samples of Erdmann (1987) are located

within cell 2. The questionably enriched $\delta^{18}O$ values of these samples causes the

concentration gradient across cell 2 to be smaller than it naturally should be. This

magnifies the concentration gradient between cell 2 and 3 creating the impression that a

larger flow rate difference occurs between these two cells.

The large Q_{out} of cell 2 is tempered with an accompanying Q_{in} of substantial

magnitude. This flow likely represents additional recharge obtained through infiltration

of meteoric water through the probable extension of the interlobate outwash sands of

cell 1 into the glacial till plain of cell 2. The net flow from cell 2 out of the subdivision

therefore decreases from 84541.8 m³/yr to 39019.9 m³/yr.

The minor flow rates of cell 3 (Q_{out}) and cells 5 through 8 (Q_{in}) represent surface

flux into or out of the freshwater aquifer. Their magnitudes, which ranges from 510.7

to 5675.5 m³/yr, are a fraction of the continuous flow rate of 51400 m³/yr. In essence,

these smaller values of Q_{in} and Q_{out} correct for minor errors in the volume or ¹⁸O

concentration estimates and possibly in the original one-dimensional flow assumption.

The Q_{out} term in cell 3 probably represents flow loss to the underlying bedrock. The Q_{in} terms of cells 5 through 8 represent recharge to the freshwater aquifer through the till plus any gain from the bedrock.

The unit flow rates in the last column of Table 6, excluding recharge cell 1, indicate that cell to cell flow is distributed fairly evenly since these values vary little in magnitude. The average unit flow rate for cells 2 through 8 is 0.48 m³/yr/m. Cell 1 has the largest flow rate per unit width (1.348 m³/yr/m) as would be expected for the recharge cell.

4.3.1.2 Subdivision 2

The computed results from Subdivision 2 indicate that cell 2 contributes significant recharge to this subdivision from surface sources since a large Q_{in} term, comparable in magnitude to the recharge Q_{in} of cell 1, accompanies this cell. Though the assumptions of the Simplified DSFM dictate that the first cell must be a recharge cell, it does not imply that the first cell is necessarily the only recharge cell. Instead the model only requires that the first cell receives incoming water from precipitation. Therefore large recharge flow rates in subsequent cells, particularly in the second cell, are expected.

The resultant flow rates through the freshwater aquifer in this subdivision range from 54062.3 to 111051.6 m³/yr. The average of cell to cell flow rates is about 81500 m³/yr. Excluding recharge cells 1 and 2, the representative rate of flow through this subdivision drops to about 70800 m³/yr.

Cell 2 results contain a large Q_{out} term, indicating that flow must definitely be exiting Cell 2 out of the subdivision. This is a reasonable result due to the proximity of the cell to the east Lambton County bedrock high, which is circular rather than ridge-like. Flow from a circular high would radiate outward in all directions, technically violating the one-dimensional flow assumption for this subdivision. The model corrects for this problem with the large Q_{out} term.

A second large Q_{out} term accompanies cell 5, indicating flow loss to outside sources. The increase in [18]O concentration gradient between cells 4 and 5 relative to the preceding cells directly implies that the groundwater flow rate decreases upon entering cell 5. The cause of this decrease in flow rate is not clear. Piezometric contours (Figure 10) suggest that flow in the southern reach of cells is southwest contrary to the

northwest direction implied by $\delta^{18}O$ value contours. Because the cells are oriented one-dimensionally in the direction of ^{18}O concentration depletion, this contradiction in flow direction may have manifested itself in cell 5 as a significant Q_{out}. The effects of this Q_{out} term are moderated by an accompanying flux of water from the overlying drift plus any gain from the bedrock with a Q_{in} of about 7700 m³/yr.

4.3.1.3 Subdivision 3

The average flow rate through this subdivision is about 22300 m³/yr averaged from cell to cell rates which range from 17835.2 to 30647.8 m³/yr. This subdivision has a relatively small range of unit flow rates from 0.214 to 0.373 m³/yr/m. The model suggests that the recharge inputs (Q_{in}) from cells 1 through 3 generally halves itself after groundwater flows through each cell. This is probably due to the position of these cells within the Bothwell Sand Plain, the area of thin drift and intertill aquifers which is mantled by surficial sands (Figure 7). The areal extent of this recharge area is most prominent in cell 1 and reduces in prominence towards cell 3. This results in the corresponding decrease in Q_{in} from cells 1 to 3. The much reduced value of Q_{in} associated with cell 4 shows the insignificant surface flux through the till which is prominent in cells 4 and 5.

The relatively large Q_{out} term of cell 4 appears to have no hydrogeologic basis and is probably therefore a symptom of the poor sample representation in this subdivision. This subdivision has the lowest representation of samples per cell in the study area with only 2 or 3 data points occupying each cell, excepting cell 5. The significant lack of data may have introduced an error into the contouring of the $\delta^{18}O$ values manifesting itself in cell 4 as a large Q_{out}.

Subdivision 3 exhibits the lowest flow rates of all the subdivisions. This was anticipated since the most depleted $\delta^{18}O$ values of the study area were concentrated in this area. In subdivision 3, older glacial water, isotopically characterized by $\delta^{18}O$ values <-16°/$_{oo}$, occupies a position in the freshwater aquifer that is closer to its source of recharge than water of equivalent $\delta^{18}O$ value in the other subdivisions. This thus indicates a lower rate of groundwater flow in subdivision 3.

4.3.1.4 Subdivision 4

The cell to cell volume rate in this subdivision steadily increases from a value of 28449.6 m³/yr in recharge cell 1 to 44868.2 m³/yr in cell 6. The Q_{in} values of cells 2 through 6 represent the surface flux from the drift plus any gain from the bedrock. These surface flux values steadily decrease in the direction of ^{18}O depletion from a high of 9227.8 m³/yr in cell 2 to a low of 382.5 m³/yr in cell 6. The absence of Q_{out} indicates that there is no flow loss to the underlying bedrock or to the drift overtop and that no water is leaving the subdivision until the discharge cell is reached. The flow per unit width is very evenly distributed from cell to cell varying between 0.444 and 0.508 m³/yr/m. The representative subdivision unit flow rate averages 0.487 m³/yr/m.

It is worth noting that the results in subdivision 4, which has the most samples per cell area, are clearly superior to those of the other subdivisions in terms of adhering to the original assumptions, evidenced by no Q_{out} terms.

4.3.1.5 Discussion of Subdivision Flow Rates

The three subdivisions whose flow discharges to Lake St. Clair and the St. Clair River have very good agreement among their discharge cell flow rates. Subdivision 1/cell 8, subdivision 2/cell 7 and subdivision 4/cell 6 have unit flow rates of 0.49, 0.44 and 0.49 m³/yr/m respectively. The similarity of these discharge flow rates is a good indicator of the validity of these values, especially between subdivisions 1 and 2 where $\delta^{18}O$ contours mirror each other. The equal and opposite countenance of these contours is an indication that groundwater flow rate toward the St. Clair River should be equal on either side of the river.

The model derived flow rates compare favourably with the rate of groundwater flow determined within the freshwater aquifer in the Sarnia area of Lambton County by Intera (1989). The Intera study reported a unit flow rate of 0.5 m³/yr/m through the freshwater aquifer determined using field measured hydraulic conductivity and Darcy's Law and a similar rate of 0.57 m³/yr/m derived from a finite difference flow model. These values are similar to those of the cells in closest proximity to the Intera study area, subdivision 1/cell 8 and subdivision 2/cell 7, which have unit flow rates of 0.49 and 0.44 m³/yr/m respectively.

4.3.2 Hydraulic Conductivity

The model derived flow rates listed in Table 6 were used to determine values of hydraulic conductivity (K) in order to compare the resultant K values with other sources. K was determined from the Darcy equation of the following form:

$$K = \frac{Q}{iA}$$ (7)

where:

$Q =$ cell flow rate determined from the Simplified DSFM and referred to as $Q_{i(i+1)}$ with units (m³/yr)

$i =$ average hydraulic gradient across each cell $(\Delta h/\Delta l)_{av}$

$A =$ cross-sectional area of each cell along its length (m²)

The cross-sectional area of each cell, listed as Cell C.A. in Table 7, was calculated by multiplying the cell length by the average thickness of the freshwater aquifer in that cell provided from Table 4. The average hydraulic gradient across each cell was estimated from the potentiometric contours in Figure 10. These values ranged from 0.00316 in subdivision 1/cell 1 to 0.00048 in subdivision 3/cell 5.

The calculated hydraulic conductivities, which are presented in Table 7, ranged from 3.10×10^{-6} to 4.72×10^{-5} m/s which translates to a geometric mean of 1.40×10^{-5} m/s. From Freeze and Cherry this range of K falls within the range of values for a silty to clean sand.

These estimated values of K compare well with values reported from other studies. Intera (1987, 1989) reported an average K of 5×10^{-6} m/s for the freshwater aquifer in the Sarnia area which was determined from a variety of field methods. This compares favourably with the K value of subdivision 2/cell 7, the most proximal cell in Lambton County to Sarnia, calculated to be 1.06×10^{-5} m/s. The average K for subdivision 2 has a value of 2.26×10^{-5} m/s and compares very well with the computer model generated K value of 5.0×10^{-5} m/s estimated by Hyde (1987) for Lambton County using a finite difference two-dimensional solute transport model. Pettapiece (1988) used the same

SUBDIVISON 1

CELL #(i)	CELL C.A. (m²)	$(\Delta h/\Delta l)_{(ave)}$	$Q_{i(i-1)}$ (m³/yr)	K (m/s)	$\overline{V}$(m/yr)
1	124853.4	0.00316	91749.5	7.369x10⁻⁶	2.37
2	160516.8	0.00257	52729.8	4.05x10⁻⁶	1.06
3	153150.5	0.00200	47054.3	4.77x10⁻⁶	0.99
4	194391.4	0.00189	47054.3	4.06x10⁻⁶	0.78
5	222276.6	0.00235	51023.2	3.10x10⁻⁶	0.74
6	209134.8	0.00177	53637.6	4.59x10⁻⁶	0.83
7	202091.4	0.00157	53883.9	5.38x10⁻⁶	0.86
8	209304.0	0.00055	54394.5	1.50x10⁻⁵	0.84

SUBDIVISION 2

CELL #(i)	CELL C.A. (m²)	$(\Delta h/\Delta l)_{(ave)}$	$Q_{i(i-1)}$ (m³/yr)	K (m/s)	$\overline{V}$(m/yr)
1	165520.8	0.0011	105115.6	1.83x10⁻⁵	1.81
2	144450.0	0.00052	111051.6	4.72x10⁻⁵	2.20
3	144288.0	0.00055	95103.3	3.81x10⁻⁵	1.88
4	159642.0	0.00092	95103.3	2.05x10⁻⁵	1.70
5	138276.0	0.0009	55844.3	1.42x10⁻⁵	1.15
6	157109.4	0.00115	54062.3	9.48x10⁻⁶	0.98
7	188732.2	0.00086	54062.3	1.06x10⁻⁵	0.82

SUBDIVISION

CELL #(i)	CELL C.A. (m²)	$(\Delta h/\Delta l)_{(ave)}$	$Q_{i(i-1)}$ (m³/yr)	K (m/s)	$\overline{V}$(m/yr)
1	99878.4	0.00078	18222.9	7.38x10⁻⁶	0.54
2	112860.0	0.00052	26732.6	1.43x10⁻⁵	0.70
3	115322.4	0.0005	30647.8	1.69x10⁻⁵	0.78
4	119433.6	0.0005	17835.2	9.50x10⁻⁶	0.44
5	85176.0	0.00048	17835.2	1.39x10⁻⁵	0.62

SUBDIVISION 4

CELL #(i)	CELL C.A. (m²)	$(\Delta h/\Delta l)_{(ave)}$	$Q_{i(i-1)}$ (m³/yr)	K (m/s)	$\overline{V}$(m/yr)
1	105732.0	0.00077	28449.6	1.11x10⁻⁵	0.79
2	122760.0	0.00055	37677.4	1.77x10⁻⁵	0.90
3	116524.8	0.00059	43529.2	2.01x10⁻⁵	1.10
4	108864.0	0.00091	43928.7	1.40x10⁻⁵	1.18
5	122358.6	0.0007	44486.2	1.66x10⁻⁵	1.07
6	121615.2	0.00079	44486.2	1.46x10⁻⁵	1.07

Table 7   Summary of hydraulic conductivity (K) and average linear groundwater velocities ($\overline{v}$) derived from model flow rates.

model (MOC model developed by Konikow and Bredehoeft (1978)) and estimated a K value of 1×10^{-4} m/s for the freshwater aquifer within Kent County. The average K for subdivision 3, which is situated completely within Kent County, was calculated to be 1.24×10^{-5} m/s. The lower K value of this study is likely a better estimate of the K within Kent County as the shale underlying this area is not conducive to high K.

4.3.3 Groundwater Velocity

The average linear groundwater velocity $(\bar{v})$ through each cell was calculated using the Darcy flux equation of the form:

$$\bar{v} = \frac{Q}{nA} \tag{8}$$

where Q and A are the same terms used in Equation (7) and n is the volumetric porosity which varies for each subdivision (n=0.31 for subdivision 1, n=0.35 for subdivision 2, and n=0.34 for subdivisions 3 and 4). The estimated fluid velocities are presented in the last column of Table 7.

The calculated velocities show that groundwater flow through the freshwater aquifer is quite slow, ranging from 2.37 m/yr in the recharge cell of subdivision 1 to 0.44 m/yr in subdivision 3/cell 4. The groundwater velocities of 0.82 m/yr through subdivision 2/cell 7 and 0.84 m/yr through subdivision 1/cell 8 compare very well with the velocity reported by Intera (1989) in the Sarnia area of 0.84 m/yr which was determined from field measured K and the Darcy flux equation.

5.0 CONCLUSIONS AND RECOMMENDATIONS

5.1 Conclusions

5.1.1 Principal Flow Direction

Interpretation of $\delta^{18}O$ concentration contours, EC contours and the physical hydrogeology were used to determine the principal flow directions through the freshwater aquifer. Five well defined flow regions have been identified:

1) The first flow system occupying Oakland, Macomb, St. Clair and northern Wayne Counties flows principally southeast from the interlobate recharge area located in Oakland County toward the St. Clair River, Lake St. Clair and the Detroit River.

2) The second flow system located in Lambton County and northwestern Kent County is comprised of flow emanating from two recharge areas. Flow is west and northwest toward the St. Clair River and Lake Huron from the northern recharge area located on the eastern edge of Lambton County. Flow from the Thames River Valley recharge area is northwest flowing towards the St. Clair River and Lake Huron.

3) The third flow system originates from the Thames River Valley with flow to the southeast toward Lake Erie.

4) The fourth flow system located in Essex County has flow to the north towards Lake St. Clair and to the east towards Kent County. Flow originates from the Leamington Moraine recharge area in southern Essex County.

5) The fifth flow system originating from the region of thin till in the southwestern corner of the study area has flow to the northeast toward Lake Erie and the Detroit River.

5.1.2 Origin and History of Groundwater

The distribution of $\delta^{18}O$ contours and tritium concentrations suggest that groundwater proximal to recharge areas are of recent origin; $\delta^{18}O$ values of about -9 to -11°/$_{oo}$ occur in every recharge area and tritium concentrations >10 TU occur in most

recharge areas. Depletion of [18]O concentration in the downgradient direction indicates that groundwater is progressively older toward the discharge regions. This is corroborated by tritium values of <10 TU in most regions away from the recharge areas.

5.1.3 Flow Rates

Principal flow direction implied from contoured data was used to determine regions where flow could be approximated as one-dimensional. These regions called subdivisions were analyzed to determine flow rates by entering $\delta^{18}O$ concentration history and aquifer volume approximations for each subdivision into the discrete state modelling program (SDSFM) developed for this study. From the model derived flow rates, estimates of hydraulic conductivity (K) and average linear groundwater velocity $(\bar{v})$ were determined using various forms of the Darcy equation. The values for flow rate, K and $\bar{v}$ compared favourably against available published data and are summarized in the following list. The fifth flow system was not modelled due to the lack of sample points in eastern Wayne County.

1) The first flow system (subdivision 1) has a representative unit flow rate of approximately 0.5 m³/yr/m, an average K of 6.04 x 10⁻⁶ m/s, and an average $\bar{v}$ of 0.87 m/yr.

2) The second flow system (subdivision 2) has a representative unit flow rate in the range of 0.45 to 0.80 m³/yr/m, an average K of 2.26 x 10⁻⁵ m/s, and $\bar{v}$ ranging from 0.82 to 2.2 m/yr.

3) The third flow system (subdivision 3) has a representative unit flow rate of approximately 0.3 m³/yr/m, an average K of 1.24 x 10⁻⁵ m/s, and an average $\bar{v}$ of 0.62 m/yr.

4) The fourth flow system (subdivision 4) has a representative unit flow rate of about 0.5 m³/yr/m, an average K of 1.57 x 10⁻⁵ m/s, and an average $\bar{v}$ of 1.02 m/yr.

5.2 Recommendations

1) Observation wells penetrating the freshwater aquifer should be drilled in sections of the study area where there is a noticeable lack of wells i.e. the Detroit-Windsor Metropolitan area. These wells would improve the isotopic distribution accuracy for the areas lacking sufficient isotope information.

2) Obtain $\delta^{18}O$ values further into areas of discharge, particularly Lake St. Clair, to determine the limit of ^{18}O depletion, i.e. how much more depleted than the $\delta^{18}O$ value of -18.5 $^o/_{oo}$ does the freshwater aquifer groundwater become.

3) Freshwater aquifer well samples should be taken in the sample area of Erdmann (1987) to replace shallow well sample data.

4) Samples should be collected for tritium analysis within the Michigan study area to investigate the possibility of recent water infiltration into the fresh water aquifer.

5) Additional well samples and data should be collected to increase statistical confidence in the volume and $\delta^{18}O$ concentration estimates presented in this study.

6) Should more oxygen-18 concentration data become available, subdivisions should be redivided into smaller cells and re-analyzed.

7) Utilization of the Simplified DSFM is possible to determine the rate of movement of a contaminant (e.g. exfiltrating liquid industrial waste into the freshwater aquifer) from cell to cell.

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APPENDIX I

**BEDROCK STRATIGRAPHY
&
TILL PLAIN MORAINES**

BEDROCK STRATIGRAPHY

The stratigraphic sequence will be described in ascending order from the oldest to the youngest formation. Refer to Figure 5 for the subcrop extent of each formation.

Salina Group - The upper portion, designated as the unit "G", consists of alternating shale-dolomite sequences. The shale beds are usually greenish-grey with minor amounts of red and may contain some dolomitic beds or stringers of gypsum and anhydrite (Mozola, 1970). The dolomites are grey to greenish-grey and dense to finely crystalline, but occasionally coarse. The subcrop areal extent of the Salina is very small and is found in the extreme southwestern corner of the study area in Monroe County. There are no natural exposures present.

Bass Islands Formation - This rock formation consists of buff to brown with some light grey, dense to finely crystalline dolomites and some shaly dolomites (Mozola, 1970). There are natural outcrops reported to occur in Lasalle Township in Monroe County and an active quarry operates in Monroetown Township.

Detroit River Group - This group overlies the Bass Islands group. It is composed of three formations in the subsurface: the Lucas Formation, the Amherstburg Formation and the Sylvania Sandstone. The Lucas and Amherstburg formations have a carbonate lithology difficult to subdivide in the subsurface. Hence for this report the upper unit of the group will include the two carbonate formations and will be referred to as the Detroit River Dolomites.

Sylvania Sandstone - This is the basal formation of the Detroit River Group. It is a white to light grey, cross-bedded, fine to medium grained, high purity, quartz sandstone. In Monroe, Wayne and Essex Counties the basal beds rest unconformably on the Bass Islands Group. There are several natural exposures of the Sylvania in the study area. They are found along the Raisin River in Monroe County and along the Huron River between Wayne and Monroe counties.

Detroit River Dolomites - This strata section is dominantly buff to brown, finely crystalline dolomites with some limestone, and is occasionally cherty and/or argillaceous (Mozola, 1969, 1970). The basal section, the Amherstburg, is characteristically dark coloured with black, bituminous zones and is more often limestone than dolomite (Telford and Russell, 1981). The overlying Lucas Formation is primarily dolomitic and is characterized by a lighter colour and the presence of anhydrite in lenses or thin beds of limited extent (Telford and Russell, 1981).

Dundee Formation - This unit is composed of grey, buff to light-brown, cherty, finely to coarsely crystalline crinoidal limestone and dolomites. Quartz sand grains are common at the base and small amber-coloured spore cases are present throughout the formation (Telford and Russell, 1981).

Hamilton Group / Traverse Group - This group consists of limestones, shales and dolomites, with the general succession from top to bottom being predominantly limestone-shale-limestone-shale. The shales are usually blue-grey to grey, occasionally brownish, and may contain some thin calcareous or dolomitic beds. Occasional limestone interbeds occur within the shales. The limestone is light-grey, grey or grey-brown in colour, fine to medium grained, argillaceous, high in calcium and richly fossiliferous at times. The dolomite beds are normally grey or buff and thinly or massively bedded. The carbonate rocks commonly contain shale throughout their sections and are occasionally very cherty. The shale-carbonate rock ratio changes are gradual.

Kettle Point Formation / Antrim Shale - The lithology of this formation is dark brown to black, fissile, finely laminated, bituminous shale with occasional interbeds of green shale. The most characteristic feature of this formation is the presence of hard, black to brown, crystalline, and spherical concretions averaging one meter in diameter. These features, called "kettles", are composed of anthraconite, a petroliferous form of calcium carbonate and commonly have pyrite and/or marcasite nodules associated with them. A second characteristic of the Antrim shale is the profusion of small amber-coloured spore cases.

Where the Kettle Point/Antrim is directly overlain by glacial drift, the formation may leak "shale gas" which may seep into the overlying permeable sediments. Drillers throughout the study area have recorded the presence of this gas while drilling in glacial drift deposits (Mozola, 1970).

Port Lambton Group / Bedford - Berea - Sunbury Formations - The Port Lambton beds of southwestern Ontario are equivalent to the Bedford, Berea and Sunbury formations in Michigan (Sanford and Brady, 1955).

Bedford Shale - The Bedford is dominantly a grey to dark grey shale with scattered beds that are blue-grey in colour and slightly micaceous. Occasional micaceous sandstone and/or shaly dolomite or limestone may be present (Mozola, 1953).

Berea Sandstone - This formation has a sandstone and shale lithology. The sandstone layers are fine-grained, micaceous grey to light drab or brown in beds of varying thickness. The sandstone layers are well cemented, but friable, with water-bearing zones. Nearly everywhere the sandstone beds are separated by beds of light grey to blue-grey shales having sporadic zones of calcareous or dolomitic materials.

Sunbury Shale - This formation succeeds the Berea sandstone. Lithologically it is a hard, dark brown or dark grey to black shale with traces of dolomite.

The Berea and Bedford have a contact that is difficult to determine. The extent of the Sunbury Shale is very uncertain due to lack of bedrock well records. For these reasons this report will consider the Berea, Bedford and Sunbury formations as one unit.

Coldwater Formation - The formation is dominantly a grey to bluish-grey shale or sandy shale which is micaceous in some areas. It becomes increasingly arenaceous upwards and is interspersed with thin beds or lenses of limestone or dolomite. Clay-ironstone nodules, concretional masses with shale centers surrounded by concentric limonitic shells, are a distinctive feature of this formation.

TILL PLAIN MORAINES

The following is a brief description of the moraines that are found within the level till plains of the study area:

Leamington Moraine - is a small dome which stands about 30 m above the surrounding plain. It is composed of a heavy, compact till, the sides are mantled by sand and gravel and it is surrounded by gravel beaches of glacial origin (Chapman and Putnam, 1984). Broad aprons of stratified sand and gravel extend beyond the moraine's boundary which were formed by glaciofluvial processes.

Blenheim Moraine - rises 15 to 30 m above the general level and is a wide modest ridge which is encircled by gravel terraces and shorecliffs.

Charing Cross - is a long, low, thin ridge about 25 km long located just southwest of the Blenheim Moraine and is aligned northeast-southwest.

Wyoming Moraine - is a broad ridge composed of clay till which trends northeast-southwest. It begins 6 km west of the town of Wyoming in north central Lambton County and continues northeast, past the study area border.

Seaforth Moraine - the extreme southern reach of this moraine is found within Lambton County just southeast of the Wyoming Moraine and forms a ridge of clay till from Watford north toward Arkona.

Birmingham Moraine - is a narrow ridge which trends northeast-southwest and which parallels the border between the hilly belt located in the northwestern corner of the study area and the flat glacial plain.

Detroit Moraine - was deposited simultaneously with and at right angles to the Birmingham Moraine. It is an inconspicuous, broad, smoothed, low ridge which appears to extend beneath the Detroit River and continues as a buried recessional moraine in northwestern Essex County (Morris, 1988).

Mt. Clemens Moraine - is a long, thin waterlaid moraine, which runs northeast-southwest through Macomb County and then swings southeast at the Wayne - Macomb County border paralleling the Detroit Moraine.

Emmett Moraine - skirts the northeast corner of Macomb County trending southeast but which then pivots almost 90° to continue in a southwesterly direction. It disappears just before reaching the northern shore of Lake St. Clair.

Grosse Isle Moraine - is a continuation of the Emmett Moraine and covers the entire island of Grosse Isle and the adjacent mainland to the west of the island.

<u>Port Huron Moraine</u> - is the youngest moraine that is encountered within the study area and is part of the Port Huron morainic system which borders the southern shore of Lake Huron in both Michigan and Ontario. Only a tiny portion of this moraine system is found in the Michigan study area running through its type locality, the city of Port Huron.

APPENDIX II

ANALYTICAL RESULTS

CRNOKRAK (1990) SAMPLES (Michigan):

SAMPLE #	COUNTY	TOWNSHIP	SECTION/ TOWNSHIP/ RANGE	STATIC HEAD (m)	THICKNESS (m)	$\delta^{18}O$ (°/oo)	E.C. (µS/cm)
1	OAKLAND	OAKLAND	4/4N/11E	288.3	91.44	-9.89	506
3	OAKLAND	OAKLAND	19/4N/11E	274.3	89.0	-9.03/-9.34	371
4	OAKLAND	OXFORD	14/5S/10W	309.4	>91.5	-9.91	333
8	OAKLAND	OAKLAND	26/4N/11E	264.3	89.31	-9.74	345
9	OAKLAND	ADDISON	15/5N/11E	294.8	96.0	-10.34	401
10	OAKLAND	ADDISON	30/5N/11E	302.3	108.8	-9.97	578
12	OAKLAND	PONTIAC	21/4N/10E		91.44	-9.78	362
13	OAKLAND	AVON	28/3N/11E	234.4	66.14	-11.67	941
14	OAKLAND	ORION	33/4N/10E	295.7	122.83	-10.64	950
15	OAKLAND	BLOOMFIELD	24/2N/10E	219.7	67.36	-10.23/-10.14	685
16	OAKLAND	BLOOMFIELD	31/2N/10E	240.2	66.45	-9.61	474
17	OAKLAND	W.BLOOMFILED	27/2N/9E	264.9	92.66	-9.76	493
18	OAKLAND	FARMINGTON	21/1N/9E	228.0	55.17	-9.60	494
19	OAKLAND	SOUTHFIELD	27/1N/10E	191.0	34.75	-12.99	708
20	OAKLAND	LYON	35/1N/7E	285.0	92.96	-9.88	390
21	OAKLAND	NOVI	29/1S/7E	278.0	112.8	-9.15	358
22	OAKLAND	FARMINGTON	25/1N/9E	195.1	42.06	-9.53	531
23	McCOMB	BRUCE	32/5N/12E	264.7	72.54	-10.02	538
24	McCOMB	ARMADA	18/3N/13E	235.5	35.97	-10.12/-9.88	1350
25	McCOMB	WASHINGTON	34/4N/12E	207.9	49	-9.77	1239
26	McCOMB	WASHINGTON	10/4N/12E	241.4	60.05	-9.69	502
27	McCOMB	RAY	19/4N/13E	199.7	43.28	-9.94	2380
28	McCOMB	SHELBY	27/3N/12E	192.0	45.7	-10.03	1073
29	McCOMB	McCOMB	33/3N/13E	180.8	51.51	-15.86	6320
30	McCOMB	STERLING	18/2N/12E	194.2	31.55	-9.84	4340
31	McCOMB	STERLING	33/2N/12E	177.7	56.48	-16.06	2000
32	McCOMB	ERIN-LAKE	6/1N/13E	172.8	47.55	-15.54	3480
33	McCOMB	CLINTON	35/2N/13E	175.9	33.83	-11.86	1023
34	WAYNE	LIVONIA	19/1S/9E	198.1	42.06	-9.44	576
35	WAYNE	NORTHVILLE	2/1S/8E	239.0	64.01	-9.28	585
36	WAYNE	CANTON	5/2S/8E	212.1	53.95	-9.54	852
37	WAYNE	CANTON	19/2S/8E	213.0	36.58	-9.50/-9.26	1027
38	WAYNE	VAN BUREN	23/3S/8E	190.2	36.88	-9.55	480
39	WAYNE	CANTON	13/2S/8E	197.8	24.38	-9.4	486
40	WAYNE	LIVONIA	4/1S/9E	197.8	59.74	-9.27	650
41	WAYNE	SUMPTER	7/4S/8E	184.7	29.57	-7.84	306
43	WAYNE	BROWNTOWNE	20/4S/10E	181.4	11.28	-12.16	2480
44	MONROE	BERLIN	19/5S/10E	176.5	10.36	-8.49	1590
45a	MONROE	ASH	31/5S/9E	181.9	10.97	-8.09	921
45b	MONROE	ASH	31/5S/9E	181.9	10.97	-8.32	1020

46	MONROE	ASH	5/5S/9	180.7	7.32	-9.86/-10.22	1896
47	MONROE	FRENCHTOWN	27/6S/10E	177.4	2.13	-8.01	3360
48	MONROE	FRENCHTOWN	18/6S/8E	176.2	2.74	-8.73	963
49	MONROE	FRENCHTOWN	24/6S/8E	183.2	4.57	-8.46	1255
50	MONROE	EXETER	4/5S/8E	186.6	15.85	-9.12	603
51	MONROE	EXETER	19/5S/8E	185.9	13.72	-8.14/-7.94	1078
52	MONROE	EXETER	33/5S/8E	186.2	5.79	-8.74	1494
53	MONROE	LASALLE	9/7S/8E	186.5	4.88	-7.92	1847
54	MONROE	LASALLE	33/7S/8E	183.8	3.35	-8.47	917
55	MONROE	LASALLE	24/7S/8E	175.6	10.67	-8.11	1834
56	MONROE	RAISINVILLE	28/6S/8E	184.4	7.01	-7.63	1773
57	WAYNE	HURON	1/4S/9E	185.7	27.43	-9.48	2950
58	ST. CLAIR	COTTRELLVILLE	4/3N/16E	176.5	39.32	-12.69	2670
59	ST. CLAIR	CHINA	4/4N/16E	177.7	36.88	-13.79/-13.69	3950
60	ST. CLAIR	ST. CLAIR	2/5N/16E	175.6	44.50	-15.48	1690
61	WAYNE	ROMULUS	10/1N/8E	194.8	30.48	-6.48	603
62	WAYNE	GROSSEILE	30/4S/11E	172.9	16.46	-12.90	2090
63	WAYNE	GROSSEILE	557	177.1	4.88	-8.70	1900
64	ST. CLAIR	CLYDE	32/7N/16E	194.5	57.0	-10.23	756
65	ST. CLAIR	PORT HURON	5/6N/17E	167.6	44.5	-10.32	697
66	ST. CLAIR	PORT HURON	19/6N/17E	173.4	37.49	-14.82	1298
67	ST. CLAIR	CHINA	23/4N/16E	169.1	38.71	-16.63/-17.05	4160
68	ST. CLAIR	ST. CLAIR	2/5N/16E	173.7	40.84	-12.15	766
69	ST. CLAIR	COTTRELLVILLE	20/3N/16E	170.7	44.20	-16.34	1293
70	ST. CLAIR	CLAY	734/3N/16E	171.6	43.89	-16.63	2200
78	WAYNE	DEARBORN	14/2N/10E	175.6	>23	-9.22/-9.07	977
79	WAYNE	DEARBORN	14/2N/10E	-	>23	-9.34	775

CRNOKRAK (1990) SAMPLES (Ontario):

SAMPLE	COUNTY	TOWNSHIP	WELL#	EAST/NORTH	STATIC HEAD (m)	THICKNESS (m)	$\delta^{18}O$ ($^o/_{oo}$)	E.C. ($\mu S/cm$)
71	KENT	HARWICH	33-7891	28620/88928	177.12	38.40	-18.46	1467
72	KENT	CHATHAM	33-1071	97500/02500	177.4	19.51	-9.33/-9.18	690
73	KENT	CHATHAM	33-5511	06100/08150	179.5	19.8	-9.64	1640
74	KENT	DOVER	33-2287	85650/91750	176.5	19.2	-8.50	1180
75	LAMBTON	WARWICK	34-4434	23443/61609		10.36	-10.42	508
76	LAMBTON	MOORE	34-4111	95930/48440	185.8	45.11	-17.25	1496
77	LAMBTON	MOORE	NWR	88100/35650	179.8	44.5	-17.59	3160

NWR - NO WELL RECORD

ESSEX COUNTY (RM - MacGregor,1985 ; BC - CRNOKRAK, 1987):

SAMPLE #	TOWNSHIP	WELL #	EASTING/NORTHING	STATIC HEAD (m)	THICKNESS OF TILL (m)	$\delta^{18}O$ (°/oo)	δD(°/oo)	TRITIUM (TU)	E.C. (µS/cm)
RM-1	MAIDSTONE	1823	48625/81200	177.6		-15.6	NA	6	2629
RM-2	ROCHESTER	2645	65040/77725	179.2	31.4	-14.2	NA	17	5696
RM-3	TILBURY WEST	2999	77240/77425	175.6	38.1	-15.8	-114.0	26	5367
RM-4	TILBURY WEST	3278	71270/74400	178.0	35.7	-15.06	NA	0	1972
RM-5	MERSEA	2547	73980/69225	180.1	34.4	-15.4	NA	7	1205
RM-6	ROCHESTER	4252	60760/73240	185.6	39.0	-9.2	NA	10	405
RM-7	COLCHESTER NORTH	NWR	37500/70660	175.0	30.2	-13.7	-101.2	0	1314
RM-8	SANDWICH WEST	2968	33880/75400	181.4	27.4	-15.3	NA	12	2191
RM-9	ANDERDON	4099	29237/70721	174.0	20.4	-12.8	-94.5	20	1862
RM-10	MALDEN	NWR		176.2	11.0	-7.8	-58.7	68	1534
RM-11	COLCHESTER SOUTH	4637	42500/61020	181.7	24.4	-9.3	NA	17	3067
RM-12	GOSFIELD NORTH	4470	50820/62800	189.9	35.4	-9.6	NA	0	931
RM-13	GOSFIELD NORTH	4281	60300/67560	188.7	31.7	-9.5	NA	13	310
RM-14	MERSEA	NWR	66538/64509	192.9	33.5	-9.9	NA	0	383
RM-15	MERSEA	2568	70540/58125	177.1	25.6	-8.7	-59.0	0	438
RM-16	GOSFIELD SOUTH	4496	62680/55480	185.6	34.1	-9.6	NA	0	701
RM-17	COLCHESTER	NWR	40770/49620	180.1	30.2	-9.1	NA	6	537
RM-18	MALDEN	2137	31160/55920	174.7	15.8	-9.7	NA	1	1862
BC-1	GOSFIELD NORTH	3693	63180/69970	186	18	-9.78	NA	25	375
BC-2	MAIDSTONE	4342	58058/79550	183	24	-10.77	-74.7	33	568
BC-3	ROCHESTER	NWR	60450/73200	185	36	-9.18	-64.5	23	493
BC-4	ROCHESTER	NWR	63160/74400	186	31	-9.76	-68.0	6	650
BC-5	ROCHESTER	4189	59340/70520	186	33	-10.15	-70.2	6	474
BC-6	ROCHESTER	NWR	68030/75300	179	35	-8.33	-59.5	41	1880
BC-7	TILBURY WEST	3008	68950/72400	182	30	-13.64	-100.0	54	2090
BC-8	TILBURY WEST	3012	72100/72150	183	35	-14.67	NA	13	1120
BC-9	ROCHESTER	NWR	63400/74400	181	31	-10.07	NA	6	565
BC-10	TILBURY WEST	NWR	75250/74180	174	41	-16.87	NA	45	3110
BC-11	TILBURY WEST	3036	73600/76950	174	39	-15.93	-116.2	6	4000
BC-12	TILBURY WEST	3045	76900/77350	178	36	-16.09	-116.8	18	4500
BC-13	MERSEA	NWR	74500/69175	179	36	-16.98	NA	7	838

BC-14	TILBURY NORTH	4620	76600/81400	173	34	-8.25	NA	49	595
BC-15	TILBURY WEST	3026	72210/69510	181	35	-14.35	-59.5	6	1100
BC-16	TILBURY NORTH	NWR	78500/78660	171	39	-15.98	NA	6	2650
BC-17	TILBURY NORTH	2933	80000/73110	179	43	-16.66	NA	22	3900
BC-18	ROCHESTER	NWR	67750/79850	174	39	-15.93	NA	6	4300
BC-19	ROCHESTER	NWR	67550/69780	182	30	-7.24	-52.0	57	315
BC-20	TILBURY WEST	3278	71270/74400	179	36	-15.68	NA	8	1222

NWR - No Well Record

NA - Not Analyzed

KENT COUNTY (Baxter, 1987):

SAMPLE #	TOWNSHIP	WELL #	EASTING/NORTHING	STATIC HEAD (m)	THICKNESS OF TILL (m)	$\delta^{18}O$ (°/oo)	δD (°/oo)	TRITIUM (TU)	E.C. (µS/cm)
JB-1	HARWICH	2792	15220/98480	187.5	21.0	-10.94	-81.2	20	514
JB-2	HARWICH	3057	15225/03475	182.6	21.6	-9.62	-59.0	14	467
JB-3	CHATHAM	NWR	06240/13580	178.0	12.2	-9.88	-64.6	<6	867
JB-5	HOWARD	3486	33510/99355	189.0	66.75	-9.06	NA	<6	334
JB-6	ORFORD	NWR	39065/04285	180.3	68.0	-9.68	NA	<6	402
JB-7	ORFORD	NWR	34800/11130	186.5	66.4	-12.60	NA	<6	927
JB-8	ZONE	NWR	26100/19280	197.9	29.6	-9.97	NA	<6	529
JB-9	TILBURY EAST	NWR	84515/76025	177.7	35.4	-16.32	NA	<6	2000
JB-10	ROMNEY	NWR	91220/70930	173.9	45.7	-17.41	NA	<6	1200
JB-11	ROMNEY	4410	82140/68510	181.1	48.2	-16.11	NA	10	4210
JB-12	RALEIGH	NWR	99970/75470	178.0	41.1	-15.77	NA	<6	790
JB-13	RALEIGH	3752	93550/82735	178.3	27.7	-17.19	NA	8	2000
JB-14	DOVER(ED)	2257	85150/92275	175.0	21.6	-8.46	NA	8	1700
JB-15	DOVER(ED)	2393	97825/00390	173.1	19.5	-8.53	NA	<6	980
JB-16	DOVER(ED)	2038	94560/92860	175.0	18.6	-8.33	-63.4	17	560
JB-17	RALEIGH	3851	02675/83350	176.8	30.0	-12.04	NA	10	910
JB-18	HARWICH	NWR	13685/82730	171.3	50.0	-17.94	-133.6	<6	1200
JB-19	HARWICH	3817	06520/90380	171.9	28.0	-15.24	-111.5	12	4000
JB-20	HARWICH	5156	20850/97350	175.8	31.7	-14.98	-109.4	<6	570
JB-21	HARWICH	7091	27840/89680	177.0	41.1	-18.22	NA	13	1800
JB-22	HARWICH	5662	21755/86475	175.8	38.4	-18.17	-133.9	<6	1800
JB-23	HARWICH	3035	10000/97845	179.2	18.9	-9.76	NA	<6	395
JB-24	DOVER	NWR	97265/00760	175.9	22.3	-10.14	-72.3	<6	680
JB-25	CHATHAM	5407	89915/09300	173.1	20.9	-10.10	NA	11	1800
JB-26	CHATHAM	5332	82625/19900	171.6	33.2	-13.89	-102.4	<6	2300
JB-27	CHATHAM	NWR	96130/71260	173.6	16.2	-9.40	NA	<6	2200
JB-28	CHATHAM	1080	01720/07450	175.3	17.4	-9.34	NA	<6	710

NWR - No Well Record

NA - Not Analyzed

LAMBTON COUNTY:(SS - SCOTT,1986 ; WILLIAMS, 1988)

SAMPLE #	TOWNSHIP	WELL #	EASTING/NORTHING	STATIC HEAD (m)	THICKNESS OF TILL (m)	$\delta^{18}O$ (°/oo)	δD (°/oo)	TRITIUM (TU)	E.C. (µS/cm)
SS-1	BOSANQUET	3860	33300/84740	179.6	55.2	-15.73	NA	<6	992
SS-2	BOSANQUET	4016	32900/77625	202.2	12.8	-10.18	NA	55	450
SS-3	BOSANQUET	139	20120/76550	201.1	23.5	-13.49	NA	<6	1645
SS-4	PLYMTON	4395	15858/69564	200.4	25.6	-13.15	NA	6	877
SS-5	WARWICK	3041	24250/69795	217.6	27.4	-9.58	NA	11	382
SS-6	WARWICK	3029	30075/69325	227.4	30.5	-9.88	NA	17	375
SS-7	WARWICK	2940	32325/61545	233.4	39.6	-10.48	NA	19	360
SS-8	WARWICK	2924	23500/61650	213.7	10.4	-9.14	NA	55	601
SS-9	PLYMPTON	3979	14675/61410	211.3	33.2	-11.10	NA	<6	476
SS-10	PLYMPTON	1955	06900/61550	186.4	33.2	-16.26	NA	8	616
SS-11	SARNIA	2452	97100/59510	179.8	31.7	-14.84	NA	12	590
SS-12	SARNIA	2468	93760/59620	174.5	37.8	-10.47	NA	33	692
SS-13	MOORE	1713	86540/48980	183.3	43.0	-17.14	NA	13	1228
SS-14	MOORE	3954	95880/48780	187.0	43.6	-15.22	NA	22	674
SS-15	ENNISKILLEN	1160	01480/48145	192.3	33.5	-7.45	NA	95	436
SS-16	ENNISKILLEN	1179	14945/47560	199.2	17.4	-10.74	NA	<6	443
SS-17	BROOKE	528	24390/46825	215.2	19.8	-10.83	NA	18	551
SS-18	BROOKE	533	32575/47595	210.9	13.4	-9.12	NA	<6	531
SS-19	BROOKE	404	25395/39200	207.0	16.5	-8.44	NA	13	430
SS-20	BROOKE	403	23445/39345	205.4	18.0	-8.65	NA	8	410
SS-21	ENNISKILLEN	1048	14050/39800	199.3	18.3	-9.88	NA	<6	499
SS-22	ENNISKILLEN	1034	08049/40505	191.7	18.3	-10.68	NA	6	410
SS-23	MOORE	4296	95040/37720	182.9	30.5	-13.54	NA	7	1067
SS-24	MOORE	1504	87560/37960	173.4	46.6	-7.24	NA	58	409
SS-25	SOMBRA	2862	87100/34010	179.2	46.6	-16.60	NA	<6	1824
SS-26	SOMBRA	2870	96860/32390	180.7	29.0	-11.92	NA	6	1249
SS-27	DAWN	699	05140/32000	193.5	17.7	-9.94	NA	6	566
SS-28	DAWN	874	14900/31975	199.0	23.2	-9.31	NA	6	413
SS-29	EUPHEMIA	1391	23425/31345	192.7	16.8	-9.31	NA	6	420
SS-30	EUPHEMIA	1318	18935/22470	186.5	14.3	-9.07	NA	6	425
SS-31	DAWN	879	15850/22625	189.0	10.4	-8.57	NA	6	406
SS-32	DAWN	706	06200/23175	188.7	12.5	-10.10	NA	6	580
SS-33	SOMBRA	2702	96640/24020	180.7	21.9	-13.40	NA	17	1086
SS-34	SOMBRA	2686	85940/24110	174.6	37.8	-13.37	NA	<6	1309
SS-35	SOMBRA	2677	78420/23580	172.8	52.4	-17.42	NA	6	1064
SS-36	WALPOLE ISLAND	3369	77230/15780	172.8	36.06	-15.17	NA	8	1015
US1	CHINA	NWR	-	-		-14.78	NA	<6	890
US2	CHINA	NWR	-	-		-16.42	NA	23	1760
US3	ST. CLAIR	NWR	-	-		-16.19	NA	<6	1258
US4	PORT HURON	NWR	-	-		-12.78	NA	6	999

4-85*	SARNIA	DOW-GTC	-	-	34.5	-14.81	-109.2	<6	2210
6-85*	SARNIA	MOE-MOOR E1	-	-	30.3	-13.85	NA	<6	3300
7-86*	SARNIA	CN#7	-	178.01	41.14	-15.9	NA	<.8	600
2-87*	SARNIA	MSMW 2-87	-	176.67	45.11	-16.86	-121.1	2.9	2675
5-87*	SARNIA	MSMW 5-87	-	184.15	37.49	-15.93	NA	<6	1775
8-87*	SARNIA	MSMW 8-87	-	186.29	36.42	-16.03	NA	<6	1700
9-87*	SARNIA	MSMW 9-87	-	187.14	36.58	-15.67	NA	<6	910
10-87*	SARNIA	MSMW 10-87	-	186.94	36.88	-15.53	NA	9	680
14-87*	SARNIA	MSMW 14-87	-	184.66	39.93	-15.7	NA		1710

NWR - No Well Record

NA - Not Analyzed

MACOMB AND ST. CLAIR COUNTIES (Erdmann, 1987):

SAMPLE #	COUNTY	TOWNSHIP	SECTION #/ TOWN #/ RANGE#/	STATIC HEAD (m)	THICKNESS OF TILL (m)	$\delta^{18}O$ (‰)	δD (‰)	TRITIUM (TU)	E.C. (μS/cm)
RE-1FW	ST. CLAIR	COTTERVILLE	4/3N/16E	-	41	-16.38	-119.2	35	3450
RE-2FW	ST. CLAIR	IRA	2/3N/15E	177.1	36.1	-12.01	-86.3	14	1700
RE-3	MACOMB	CHESTERFIELD	1/3N/14E	NR	NR	-9.79	-67.8	<6	310
RE-4	MACOMB	CHESTERFIELD	4/3N/14E	NR	NR	-9.42	-65.6	46	600
RE-5	MACOMB	MACOMB	3/3N/13E	NR	NR	-9.11	-65.4	19	990
RE-6	MACOMB	RAY	22/4N/13E	NR	NR	-9.42	NA	<6	1700
RE-7	MACOMB	LENOX	20/4N/13E	NR	NR	-9.12	NA	16	930
RE-8FW	ST. CLAIR	CASCO	18/4 /15E	196.9	56	-8.93	NA	62	630
RE-9FW	ST. CLAIR	CASCO	23/4N/15E	181.9	47.7	-11.30	-82.8	17	2900
RE-10FW	ST. CLAIR	CHINA	15/4N/16E	179.1	36.6	-15.63	-114.0	28	2000
RE-11FW	ST. CLAIR	CHINA	4/4N/16E	178.6	39.3	-12.81	NA	42	1300
RE-12	ST. CLAIR	CASCO	1/4N/15E	-	44.8	-9.82	NA	<6	1000
RE-13	ST. CLAIR	COLUMBUS	31/5N/15E	NR	NR	-10.14	-73.1	42	410
RE-14	MACOMB	RICHMOND	33/5 /14	NR	NR	-8.73	NA	36	950
RE-15	MACOMB	ARMANDA	32/5 /13	NR	NR	-9.72	NA	37	525
RE-16	MACOMB	ARMANDA	17/5 /13	NR	NR	-9.36	NA	60	800
RE-17	MACOMB	RICHMOND	17/5 /14	NR	NR	-8.90	NA	6	530
RE-18	ST. CLAIR	COLUMBUS	19/5N/15E	NR	NR	-9.01	NA	23	35
RE-19	ST. CLAIR	COLUMBUS	24/5N/15E	NR	NR	-9.15	NA	<6	770
RE-20FW	ST. CLAIR	ST. CLAIR	15/5N/16E	178.3	46	-11.27	-83.7	<6	3350
RE-21	ST. CLAIR	KIMBALL	35/6N/16E	NR	NR	-8.95	NA	<6	360
RE-22	ST. CLAIR	WALES	36/6N/15E	NR	NR	-9.22	NA	11	680
RE-23FW	ST. CLAIR	WALES	33/6N/15E	215.8	71	-9.03	NA	<6	470
RE-24	ST. CLAIR	RILEY	33/6N/14E	NR	NR	-8.60	NA	14	590
RE-25	ST. CLAIR	BERLIN	28/6N/13E	NR	NR	-9.09	NA	8	600
RE-26	ST. CLAIR	BERLIN	11/6N/13E	NR	NR	-9.25	NA	13	630
RE-27	ST. CLAIR	RILEY	17/6N/14E	NR	NR	-9.09	NA	56	940
RE-28FW	ST. CLAIR	WALES	17/6N/15E	213.4	73	-9.56	NA	17	1100
RE-29	ST. CLAIR	KIMBALL	10/6N/16E	NR	NR	-8.58	-61.6	34	760
RE-30	ST. CLAIR	WALES	11/6N/15E	NR	NR	-9.61	NA	12	480
RE-31	ST. CLAIR	CLYDE	32/7 /16	NR	NR	-8.95	NA	56	620
RE-32	ST. CLAIR	KENOCKEE	33/7 /15	NR	NR	-9.60	NA	19	450
RE-33	ST. CLAIR	EMMETT	36/7N/14E	NR	NR	-9.12	NA	21	1200

NA: Not Analyzed

FW: Freshwater aquifer wells (remainder of wells are completed in drift).

NR: Not representative of the freshwater aquifer (since thse wells are completed in drift)

APPENDIX III

SDSFM PROGRAM LISTING

```
10      OPTION BASE 1
20      COM /A/ Vol(200),Cint(200),Cfin(200),Vij(200),Vout(200),Vin(200)
30      COM /B/ INTEGER Ncells,REAL Cin(0:1000) 31    COM /C/ C(0:1000,200)
40      FOR I=1 TO 8
50      ON KEY I LABEL "",8 GOTO Repete
60      NEXT I
70      ON KEY 1 LABEL "ENTER VALUES",8 CALL Enter_values
80      ON KEY 2 LABEL "SAVE VALUES",8 CALL Save_values
90      ON KEY 3 LABEL "SOLVE",8 CALL Solve
100     !ON KEY 4 LABEL "PRESENT",8 CALL Pres_res
110     ON KEY 5 LABEL "EXIT",8 GOSUB Rtn
120 Repete:  !
130         LOOP
140         END LOOP
150 Rtn:!
160     END
170     !_____


_____
190     SUB Enter_values
191     OPTION BASE 1
200     COM /A/ Vol(200),Cint(200),Cfin(200),Vij(200),Vout(200),Vin(200)
210     COM /B/ INTEGER Ncells,REAL Cin(0:1000)
220     COM /C/ C(*) 230    FOR I=1 TO 8 240    ON KEY I LABEL "",9 GOTO Repete
250     NEXT I
260     ON KEY 1 LABEL "ENTER VALUES",9 GOSUB Enter_values
270     ON KEY 2 LABEL "EDIT VALUES",9 GOSUB Edit_values
280     ON KEY 3 LABEL "LIST VALUES",9 GOSUB List_values
290     ON KEY 4 LABEL "RETURN",9 GOSUB Rtn
300 Repete:  !
310         LOOP
320         END LOOP
330 Enter_values:!
340         CLEAR SCREEN
350     INPUT "HOW MANY CELLS IN THE MODEL",Ncells
360     FOR I=1 TO Ncells
370     CLEAR SCREEN
380     PRINT "ENTER VOLUME OF CELL ";I
390     INPUT Vol(I)
400     PRINT "ENTER INITIAL CONCENTRATION OF CELL ";I
410     INPUT Cint(I)
420     PRINT "ENTER FINAL CONCENTRATION OF CELL ";I
430     INPUT Cfin(I)
500     NEXT I
510 RETURN
520 List_values:!
530         CLEAR SCREEN
540         Incl=0
550     Nsets=Ncells/16+1
560     FOR I=1 TO Nsets
580     PRINT TABXY(1,1);"CELL #";TAB(8);"VOLUME";TAB(15);"IN.
CON";TAB(23);"FIN. CON";
590     FOR J=1 TO 16
600     Incl=Incl+1
601     IF Incl>Ncells THEN GOTO 660
610     PRINT TABXY(1,3+J);Incl;TAB(9);DROUND(Vol(Incl),5);TAB(18);Cint(Incl);TAB(
26);Cfin(Incl);
620     NEXT J
630     DISP "HIT CONTINUE TO VIEW NEXT SET"
640     PAUSE
650     NEXT I
660 RETURN
670 Edit_values:!
680     INPUT "ENTER CELLS NUMBER TO EDIT (0-WILL EXIT)",Cnum
```

```
690   IF Cnum<>0 THEN
700   CLEAR SCREEN
710   PRINT "1) VOLUME OF CELL ";Cnum;" =";Vol(Cnum)
720   PRINT "2) INITIAL CONCENTRATION OF CELL ";Cnum;" =";Cint(Cnum)
730   PRINT "3) FINAL CONCENTRATION OF CELL ";Cnum;" =";Cfin(Cnum)
740   PRINT "4) RECALCULATE CONCENTRATIONS"
750   PRINT "5) REMOVE THIS CELL"
751   PRINT "6) ADD CELL AFTER THIS CELL"
760   PRINT "7) ESCAPE TO NEXT EDIT"
770   PRINT "8) WILL EXIT EDIT"
780   PRINT
790   PRINT
800   PRINT "ENTER NUMBER OF VALUE TO EDIT FOR CELL ";Cnum
810   INPUT Cedit
820   SELECT Cedit
830   CASE 1
840   INPUT "ENTER NEW CELL VOLUME ",Vol(Cnum)
850   CASE 2
860   INPUT "ENTER NEW INITIAL CONCENTRATION",Cint(Cnum)
870   CASE 3
880   INPUT "ENTER NEW FINAL CONCENTRATION",Cfin(Cnum)
881   CASE 4
882   CALL Convertcon(Cint(*),-1)
883   CALL Convertcon(Cfin(*),-1)
884   CASE 5
885   INPUT "ARE YOU SURE (Y/N)",An$
886   IF An$="Y" THEN
887   FOR I=Cnum TO Ncells-1
888   Vol(I)=Vol(I+1)
889   Cint(I)=Cint(I+1)
890   Cfin(I)=Cfin(I+1)
891   NEXT I
892   Vol(Ncells)=0
893   Cint(Ncells)=0
894   Cfin(Ncells)=0
895   Ncells=Ncells-1
896   END IF
898   CASE 6
899   FOR I=Cnum+2 TO Ncells+1
900   Vol(I)=Vol(I-1)
901   Cint(I)=Cint(I-1)
902   Cfin(I)=Cfin(I-1)
903   NEXT I
904   INPUT "WHAT IS THE NEW CELL VOLUME",Vol(Cnum+1)
905   INPUT "WHAT IS THE NEW INITIAL CONCENTRATION",Cint(Cnum+1)
906   INPUT "WHAT IS THE FINAL CONCENTRATION",Cfin(Cnum+1)
907   Ncells=Ncells+1
909   CASE 7
960   GOTO 680
970   CASE 8
980   GOTO 1020
990   END SELECT
1000  GOTO 700
1010  END IF
1020 RETURN
1030 Rtn:!
1040 SUBEND
1050 !
```

```
1060 SUB Save_values
1070 OPTION BASE 1
1080  COM /A/ Vol(200),Cint(200),Cfin(200),Vij(200),Vout(200),Vin(200)
1090  COM /B/ INTEGER Ncells,REAL Cin(0:1000)
1091  COM /C/ C(*)
1092  ALLOCATE Slime(0:1000)
1100  DIM Cat$(100)[16]
1110  FOR I=0 TO 9
1120  ON KEY I LABEL "",9 GOTO Repete
1130  NEXT I
1140  ON KEY 1 LABEL "SAVE" GOSUB Sav
1150  ON KEY 2 LABEL "RETREIVE" GOSUB Rtv
1160  ON KEY 3 LABEL "RETURN" GOSUB Rtn
1170 Repete: !
1180         LOOP
1190         END LOOP
1200 Sav: !
1210  CLEAR SCREEN
1220  INPUT "WHAT FILENAME WOULD YOU LIKE",File$
1230  CAT TO Cat$(*);NAMES
1240  FOR I=1 TO 100
1250  IF Cat$(I)=File$ THEN PURGE File$
1260  NEXT I
1270  CREATE BDAT File$,1,64*256
1271  CREATE BDAT File$,Ncells,8008
1280  ASSIGN @P1 TO File$
1281  ASSIGN @P2 TO CHR$(179)&File$
1290  OUTPUT @P1,1;Ncell::Vol(*);Cint(*);Cfin(*);Vij(*);Vout(*);Vin(*)
1310  ASSIGN @P1 TO *
1311  FOR I=1 TO Ncells
1312  MAT Slime= C(*,I)
1313  OUTPUT @P2,I;Slime(*)
1314  NEXT I
1320 RETURN
1330 Rtv:!
1340  CLEAR SCREEN
1350  CAT TO Cat$(*);NAMES,COUNT Cownt
1360  Nsets=Cownt/64
1370  Incl=0
1380  FOR I=1 TO Nsets
1390  FOR J=1 TO 4
1400  FOR K=1 TO 16
1410  Incl=Incl+1
1420  PRINT TABXY((J-1)*20+1,K);Incl;")";Cat$(Incl)
1430  NEXT K
1440  NEXT J
1450  DISP "HIT CONTINUE TO VIEW NEXT SET"
1460  PAUSE
1470  NEXT I
1480  INPUT "ENTER THE DESIRED FILE NUMBER",Filenum
1490  ASSIGN @P1 TO Cat$(Filenum)
1500  ENTER @P1,1;Ncells;Vol(*);Cint(*);Cfin(*);Vij(*);Vout(*);Vin(*)
1520  ASSIGN @P1 TO *
1530 RETURN
1540 Rtn:!
1550 SUBEND
1560 !
```

```
1570  SUB Convertcon(A(*),Sn)
1580  OPTION BASE 1
1590  IF Sn=1 THEN
1600  MAT A= A/(1000)
1610  MAT A= A+(1)
1620  MAT A= A*(2.E-6)
1630  END IF
1640  IF Sn=-1 THEN
1650  MAT A= A/(2.E-6)
1660  MAT A= A-(1)
1670  MAT A= A*(1000)
1680  END IF
1691  SUBEND
1700  !_____

1710  SUB Solve
1720  OPTION BASE 1
1730  COM /A/ Vol(200),Cint(200),Cfin(200),Vij(200),Vout(200),Vin(200)
1740  COM /B/ INTEGER Ncells,REAL Cin(0:1000)
1741  COM /C/ C(*)
1744  REDIM C(0:1000,Ncells)
1750  ALLOCATE C1(0:1000),C2(0:1000)
1760  INTEGER I,J,K,L
1770  CLEAR SCREEN
1780  DISP CHRS(131);"SOLVING"
1781  FOR I=0 TO 550
1782  Cin(I)=-16*(1-I/1100)
1783  NEXT I
1786  FOR I=551 TO 800
1787  Cin(I)=-8
1788  NEXT I
1789  FOR I=801 TO 1000
1790  Cin(I)=-2*(I-801)/200-8
1791  NEXT I
1792  CALL Convertcon(Cin(*),1)
1793  CALL Convertcon(Cint(*),1)
1800  CALL Convertcon(Cfin(*),1)
1810  MAT C(0,*)= Cint(1:Ncells)
1820  Cin1=(-8/1000+1)*2.0E-6
1860  Tim=1000
1861  Errr=.00001
1862  MAT Vij= (0)
1863  MAT Vout= (0)
1864  MAT Vin= (0)
1870  FOR I=1 TO Ncells
1880     SELECT I
1890     !
1900     !*** FIRST CELL IN THE SYSTEM ***
1910     !
1920     CASE 1
1930     !*** DETERMINE VIN DIRECTLY FOR CELL No. 1***
1940     Vin(1)=-Vol(1)/Tim*LOG((Cin1-Cfin(1))/(Cin1-Cint(1)))
1950     !*** DETERMINE CONCENCTRATION HISTORY FOR CELL No. 1***
1952     !C(0,1)=Cint(1)
1960     !FOR J=1 TO 1000
1970     !C(J,1)=Cin*(1-EXP(-Vin(1)*J/Vol(1)))+Cint(1)*(EXP(-Vin(1)*J/Vol(1))) 1980
!NEXT J
1990     Vij(1)=Vin(1)
1991     MAT C1= Cin
2000     CALL Euler(1000,C1(*),C2(*),2,1)
2001     Vijnew=Cfin(1)/C2(1000)*Vij(1)
2003     REPEAT
2004  Vijold=Vij(I)
```

```
2005  Vij(I)=Vijnew
2006  Inc=I
2007  CALL Euler(1000,C1(*),C2(*),2,I)
2008  Delta1=(Cold-Cfin(I))
2009  Delta2=(C2(1000)-Cfin(I))
2010  Vijnew=(Vijold-Vijnew*Delta1/Delta2)/(1-Delta1/Delta2)
2011  Cold=C2(1000)
2012  Rasio=Cold/Cfin(I)
2013  UNTIL Rasio<1+Errr AND Rasio>1-Errr
2014  Vin(1)=Vij(1)
2015  MAT C(*,1)= C2
2016  FOR J=550 TO 1000
2017  Cin(J)=(-8/1000+1)*2.0E-6
2018  NEXT J
2020     PRINT "DONE FIRST CELL"
2021     !
2022     !*** INTERMEDIATE CELLS***
2023     !
2024     CASE 2 TO Ncells
2025     !*** ASSUME NO RECHARE OR DISCHARGE ***
2026     Inc=1
2027     MAT C1= C(*,I-1)
2028     CALL Euler(1000,C1(*),C2(*),0,I)
2029     IF C2(1000)/Cfin(I)>1+Errr THEN GOTO Itervij
2030     IF C2(1000)/Cfin(I)<1-Errr THEN GOTO Itterrecharge
2031     IF C2(1000)/Cfin(I)<1+Errr AND C2(1000)/Cfin(I)>1-Errr THEN GOTO
Nextone
2032 !
2033 Itterrecharge: !
2034 ! 2035            Rasio=C2(1000)/Cfin(I)
2036               Vincnew=1/Rasio*Vij(I)
2037               Cold=C2(1000)
2038  REPEAT
2039  Vincold=Vin(I)
2040  Vin(I)=Vincnew
2041  CALL Euler(1000,C1(*),C2(*),1,I)
2042  Delta1=(Cold-Cfin(I))
2043  Delta2=(C2(1000)-Cfin(I))
2044  Vincnew=(Vincold-Vincnew*Delta1/Delta2)/(1-Delta1/Delta2)
2045  Cold=C2(1000)
2046  Rasio=Cold/Cfin(I)
2047  UNTIL Rasio<1+Errr AND Rasio>1-Errr
2048  Vij(I)=Vin(I)+Vij(I-1)
2049  GOTO Nextone 2050 !
2051 Itervij:!
2052 !
2053               Rasio=C2(1000)/Cfin(I)
2054               Vijnew=Vij(I)*(1-Rasio)
2055               Cold=C2(1000)
2056               Vijold=Vij(I)
2057  REPEAT
2058  Vijold=Vij(I)
2059  Vij(I)=Vijnew
2060  Inc=I
2061  CALL Euler(1000,C1(*),C2(*),2,I)
2062  Delta1=(Cold-Cfin(I))
2063  Delta2=(C2(1000)-Cfin(I))
2064  Vijnew=(Vijold-Vijnew*Delta1/Delta2)/(1-Delta1/Delta2)
2065  Cold=C2(1000)
2066  Rasio=Cold/Cfin(I)
2067  UNTIL Rasio<1+Errr AND Rasio>1-Errr
2068  Vout(I-1)=Vij(I-1)-Vij(I)
2069  Vij(I-1)=Vij(I)
```

```
2070  GOTO Nextone
2071  !
2072 Nextone: !
2073        MAT C(*,I)= C2
2074        END SELECT
2075  NEXT I
2076  DISP CHR$(128);"DONE"
2077  CALL Convertcon(C(*),-1)
2078  CALL Convertcon(Cfin(*),-1)
2079  CALL Convertcon(Cint(*),-1)
2080  PRINT "CFINAL"; "C1000"; TAB(40) ;"Vij(I)" ;TAB(60) ;"Vin(I)" ;TAB(80)
;"Vout(I)"
2081  FOR I=1 TO Ncells
2082  PRINT Cfin(I) ;TAB(20) ;C(1000,I) ;TAB(40) ;Vij(I) ;TAB(60); Vin(I)
;TAB(80);Vout(I);TAB(100)
2083  NEXT I
2084  PAUSE
2085  GINIT
2086  OUTPUT 705;";IN"
2087  PLOTTER IS 705,"HPGL"
2088  CLEAR SCREEN
2089  GCLEAR
2090  GRAPHICS ON
2091  Tit$="O18 CONCENTRATION Vs. TIME"
2092  X$="TIME IN YRS"
2093  Y$="O18 CONCENTRATION"
2094  MOVE 65,93
2095  CSIZE 2.5,.6
2096  LORG 5
2097  CLIP OFF
2098  DIM Tit$[80],X$[80],Y$[80]
2099  LABEL Tit$
2100  MOVE 65,15
2101  CLIP OFF
2102  LABEL X$
2103  MOVE 15,55
2104  LDIR PI/2
2105  CLIP OFF
2106  LABEL Y$
2107  LDIR 0
2108  CSIZE 2,.6
2109  FOR I=1 TO Ncells
2110  MOVE I*18,10
2111  PEN I
2112  CLIP OFF
2113  LABEL "CELL No. ";I
2114  NEXT I
2115  CSIZE 2.5,.6
2116  VIEWPORT 20,110,20,90
2117  Mx=MAX(C(*))
2118  Mn=MIN(C(*))
2119  FRAME
2120  CSIZE 2,.6
2121  WINDOW 0,1000,Mn,Mx
2122  FOR I=1 TO Ncells
2123  PEN I
2124  MOVE 0,C(0,I)
2125  FOR J=0 TO 1000
2126  DRAW J,C(J,I)
2127  NEXT J
2128  NEXT I
2129  MOVE 0,Mx
2130  LORG 8
```

```
2131 CLIP OFF
2132 LABEL DROUND(Mx,2)
2133 MOVE 0,Mn
2134 CLIP OFF
2135 LABEL DROUND(Mn,2)
2136 LORG 6
2137 LABEL "0"
2138 MOVE 1000,Mn
2139 LABEL "10000"
2140 SUBEND
2141 !
```
_____

```
2142 SUB Euler(INTEGER Nit,REAL C1(*),REAL C2(*),INTEGER Typ,Inc)
2143 OPTION BASE 1
2144 COM /A/ Vol(200),Cint(200),Cfin(200),Vij(200),Vout(200),Vin(200)
2145 COM /B/ INTEGER Ncells,REAL Cin(0:1000)
2146 COM /C/ C(*)
2147 MAT C2= (0)
2148 SELECT Typ
2149 !
2150 !
2151 CASE 0
2152 C2(0)=Cint(Inc)
2153 V=Vij(Inc-1)
2154 Vij(Inc)=V
2155 FOR I=1 TO Nit
2156 C2(I)=C2(I-1)+(V*C1(I)-V*C2(I-1))/Vol(Inc)
2157 NEXT I
2158 !
2159 !
2160 CASE 1
2161 C2(0)=Cint(Inc)
2162 V=Vij(Inc-1)
2163 Vre=Vin(Inc)
2164 Vowt=V+Vre
2165 FOR I=1 TO Nit
2166 C2(I)=C2(I-1)+(V*C1(I)+Vre*Cin(I)-Vowt*C2(I-1))/Vol(Inc)
2167 NEXT I
2168 !
2169 !
2170 CASE 2
2171 C2(0)=Cint(Inc)
2172 V=Vij(Inc)
2173 FOR I=1 TO Nit
2174 C2(I)=C2(I-1)+(V*C1(I)-V*C2(I-1))/Vol(Inc)
2175 NEXT I
2176 !
2177 !
2178 END SELECT
2179 SUBEND
```

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1965   Born in Winnipeg, Manitoba, Canada on September 29

1987   Received the Degree of Bachelor of Applied Science in Geological
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