1972

Geothermal Measurements In A Sedimentary Basin

Alan Stephen Judge

GEOTHERMAL MEASUREMENTS IN A SEDIMENTARY BASIN

by

Alan Stephen Judge

Department of Geophysics

Submitted in partial fulfillment
of the requirements for the degree of
Doctor of Philosophy

Faculty of Graduate Studies
The University of Western Ontario
London, Canada
Nov. 1971

ABSTRACT

A regional heat flow survey of Southern Ontario and part of the Southern Peninsula of Michigan was conducted. Variations of temperature with depth below the earth's surface were collected in 31 boreholes from which regional temperature gradients and their variation between and within geological formations could be examined. Thermal conductivity measurements have been made for as many of the geological formations as cores were available. These totaled some 800 samples. Variations of conductivity were examined within and between geological formations across the region. The effects of inter-variations of composition, porosity, density and thermal conductivity were investigated in several formations. Heat flows across the region were calculated and the effects of disturbing factors investigated. A completely cored borehole on the campus of the University of Western Ontario allowed detailed investigations to be made of the return to equilibrium of a borehole after completion of drilling and of the effect of recent surface temperature changes on subsurface temperatures. Possible variations of surface temperatures in the past were considered and some estimates made of the effects induced in the subsurface by the advances and retreats of the Pleistocene ice-sheets. The measured heat flows were found to be similar to those of the exposed shield rocks of Grenville age which also form the basement below the Palaeozoic sediments of Southern Ontario.

ACKNOWLEDGEMENTS

The author would like to thank Dr. A.E. Beck for originally suggesting the project and his continual supervision and encouragement during its prolonged continuance. Thanks also go to Dr. J.H. Sass, for ably acting as supervisor during Dr. Beck's absence on sabbatical leave, and to Dr. Winder for his revelations into the geology of Southern Ontario and Michigan. Dr. B. Brigham assisted in introducing the author to the wealth of information contained in the Ontario Well Data File. Mr. J. Carson and Mr. Z. Logis of the Geophysics Dept. assisted in laboratory and field measurements.

A very great deal of the information included here, was possible only by the tremendous assistance of the Ontario Department of Energy Resources Management, in particular Mr. D. Sharp and his Inspectors, Imperial Oil Enterprises, Mr. J. Pounder, the Michigan Consolidated Gas Co., Mr. K. Lundy, Consumers Power Co., Mr. J. Duane, Consumer Gas Co., Mr. B. Wallace & Dr. G. Theokritoff. It really must be stressed that only with the assistance of each of these organisations and companies was the project at all possible.

Finally, grateful thanks must go to Dr. A.M. Jessop of the Earth Physics Branch both for permission to use results collected by the Heat Flow group and for the patience that he has shown while this work was being completed.

The author also acknowledges with thanks the award of an Ontario Fellowship during his years at the University of Western Ontario.

INDEX SECTION

TABLE OF CONTENTS

Appendices

LIST OF TABLES

LIST OF FIGURES

Appendices

CHAPTER I

GENERAL INTRODUCTION

I-1 INTRODUCTION

I-1-1 Importance of Geothermal Studies

In any attempt to understand the earth's history and its present physical state, a knowledge of its thermal conditions is of prime importance, since the earth is essentially a heat engine. Discussions on the thermal history are quite numerous; to cite a few - Holmes (1915), Jacobs & Allan (1954), MacDonald (1959) and Clark & Ringwood (1964). The classic paper by Holmes (1915) was the first to include the heat production by natural radioactivity and thus to end the arguments that Kelvin (1864) had raised in the previous century. More recent developments, principally in seismology, have led to better information as to the plausible stucture and conditions of the interior. Any viable theory, however, must utilize the surface geothermal flux as a boundary condition. This point is well illustrated by the first oceanic heat flow measurements by Petterson (1949) and Bullard (1954). The similarity of results from continental and oceanic areas led to a reconsideration of the radioactive isotope distribution in the crust and mantle by Verhoogen (1956). Utilizing this fact and Tilton & Reed's (1963) analysis of the distribution of radioactive elements in dunite, Ringwood & Clark (1964) were able to deduce that "the most important conclusion emerging from our discussion of the radioactivity of the upper mantle is that a heat flow of 0.5 μcals/cm^2 is derived from depths greater than 400 km". Such a conclusion is of very great importance, since the rheological conditions and phase transformations which may occur in the earth's interior depend largely on the temperature distribution with depth.

More recently a combination of heat flow and heat production measurements by Roy et al (1968) and Lachenbruch (1968) have enabled more stringent limits to be placed of the heat flux from the lower crust and mantle. It also indicates that the heat contribution from the mantle could vary by a factor of three beneath different continental geological provinces.

I-1-2 Distribution of Heat Flow Measurements

The two thousand or so terrestrial heat flow results reviewed by Lee & Uyeda (1965), Simmons & Horai (1968) are very unevenly distributed with large gaps existing in Asia, Africa, South America and Antarctica. Over 85% of the results available were made in the oceans, reflecting the greater simplicity in organization and method. However, many of these measurements are suspect due to insufficient regard for such important factors as sea-bottom topography, sediment thickness, stability of water-sediment interface, behaviour of water-logged sediments under ocean bottom conditions or adequate sampling of thermal conductivity. Many of the few land stations are also somewhat suspect, since measurements are usually made in pre-existing boreholes or mines in areas that may be anomalous. Only recently have programmes of drilling special holes for heat flow measurements been commenced e.g. Jessop (1964).

I-1-3 Propagation of Temperature

A temperature perturbation may be propagated in one of three ways, these being conduction, convection and radiation. In the mantle most or all of these processes may be active (Clark, 1957, Vening Meinesz, 1962). However, in this section it will suffice to restrict comment on those processes which might be active in the crust. Radiative transfer only becomes important at temperatures greater than 1000°C and can therefore be neglected. Mass transfer is probably important in certain parts of the upper crust as water migration certainly does occur, evidenced by the results from Eakring (Bullard & Niblett, 1951) and Pechelbronn (Claude, 1952), but only under specialized geological conditions. However, such conditions are important in active geothermal areas and may be important in very permeable sedimentary rocks or in faulted areas. Thus, most of the heat flow through the earth's surface, some 10^{28} ergs/yr at present, is conducted to the surface, a fact which, in itself, leads to problems which must be recognized in the use and interpretation of heat flow data in comparison with seismic or electromagnetic interpretations. The velocity of propagation of a conducted heat pulse of a period of 10^6 years in crust-like materials is of the order of 10^{-7} cms/sec. A

disturbance 10^6 years ago at the mantle-crust boundary would now be becoming observable at the surface. Equally well, changes in the surface temperature affect the rate of escape of heat and a new equilibrium state is not reached for a long period, eg., the maximum disturbance due to the ice ages is to be found at a depth of about 1.5 km. Possibilities arise that here is a tool for direct measurement of thermal conditions in the past.

I-2 PREVIOUS GEOTHERMAL STUDIES

General studies of the thermal conditions of the crust for scientific purposes have been undertaken for many years now, initially under the auspices of the British Association for the Advancement of Science (1882), in the 1920's by Van Orstrand (1934) in the United States and in more recent times by the Russians (U.S.S.R. Coll. 1957 & 58). The majority of these collections were merely of temperature logs taken in boreholes, plus a few results in mine drifts. Even once reasonably accurate measuring equipment was developed and the necessity of leaving a borehole for a period of time after drilling to attain equilibrium with the surrounding rocks was appreciated, very little was done other than temperature logging. Hence a great deal of information exists on thermal gradients, but, owing to a lack of data on the thermal conductivities of the formations encountered, the use to which this information may be put is limited. Examples of the qualitative value of gradient measurements may be found in Rozychi's collection for Poland (Rozychi, 1948):

Cool areas close to old shield in North East and in the Carpathian Mts.

Warm areas close to Hercynian and Kimmerian folding

Increased temperatures in vicinity of Tertiary volcanics.

For the United States, Van Orstrand quoted the following results (Van Orstrand, 1935):

Low gradients in Appalachians, New Mexico and Texas

High gradients in Wyoming, East Texas, Kansas and Oklahoma

Irregular gradients in the Rocky Mountains.

Whilst these results are interesting, they may merely reflect regional variations in the mean thermal conductivity of near surface rocks. In a highly localized survey, temperature differences could point out such structural features as a salt dome (Strong,1934), although on a continental scale correlations are of little value.

Modern measurements of terrestrial heat flow through the surface layers began with Benfield (1939) and Bullard (1939), both of whom realized the importance of thermal conductivity measurements on representative rock samples. To date, most of the continental measurements have been made in boreholes in mining areas or in mines and tunnels (Misener et al, 1951, Clark & Niblett, 1956, Howard & Sass, 1964, Roy et al, 1968) ie., in metamorphic and igneous regions. Such areas have been most popular, since holes are more numerous, cores are usually taken for assay purposes and the thermal conductivities are less variable than those of sediments.

I-3 THESIS OBJECTIVES

Sedimentary basins occupy some 40% of the earth's emergent land surface. Their distribution compiled from several sources is shown in FIG I-1. Much of the continental shelf regions are also a sedimentary sequence and, because of the thermal disturbance of the unconsolidated sediments by active ocean currents, shallow oceanic probe techniques will probably be dropped in favour of oil and gas exploration holes; thus they may be added to regions considered here. To get a reasonable spacial coverage of heat flow measurements, basin measurements must be included. The genesis of basins and their persistence is a problem requiring solutions which might be aided by heat flow measurements.

Objectives in this work, then, were to study some of the problems of heat flow work in sedimentary regions, to build suitable equipment for temperature and thermal conductivity measurement, to determine what factors may disturb the equilibrium heat flow and what techniques might be used to investigate these factors.

To study the thermal conditions in any region the methods used must depend on the information and material available, but there are roughly three approaches which, in order of increasing accuracy, are:

1) Use of temperature gradients from boreholes without core,
2) Use of temperature gradients in boreholes without core but in areas in which representative core is available from geological horizons which may be correlated with horizons present in the borehole, and
3) Conductivity measurements and temperature gradients derived

3) Conductivity measurements and temperature gradients derived from the same borehole.

In the present work the possibility of using the second alternative with high accuracy has been investigated, and one hole in the third category has been examined in detail to determine some of the factors which might upset measurements and make single measurements in a region unrepresentative of the region. The usual heat flow technique is to consider a single borehole to be representative.

As a result of the variations of heat-flow down one borehole it was possible to examine surface temperature changes in the recent past which was then extended to look for similar effects of larger magnitude in the more distant past.

Southwestern Ontario presents an ideal opportunity to study sub-surface thermal conditions. There is a good system of concession roads which permits access to within one mile of any borehole, and the core is stored in convenient major centres. It is the home of the North American oil industry with the first producing oil field at Oil Springs in Enniskillen township and is still producing from some eighty or more oil and gas fields widely distributed through the area. Drilling still continues. Although the policies of the Ontario Department of Energy Resources and Resources Management (ODERM) require that holes be plugged before abandonment, temperature data can still be obtained in observation wells, suspended wells and old wells. At present ODERM is conducting a programme of plugging abandoned holes and some of these boreholes have been logged for temperature by the writer. Reasonable large amounts of core were available for thermal conductivity sampling. Several complete diamond drill sections were available with which measurements on less complete sections were compared. This permitted an extensive study of the spatial conductivity variations in given formations and between different formations. Finally, it was possible to build up average formation conductivities and their variation, together with regional variations of temperature, gradient and heat flow. Comparisons were then made to look for structural effects. The logging of several holes in the Michigan Basin provide end-members for the work.

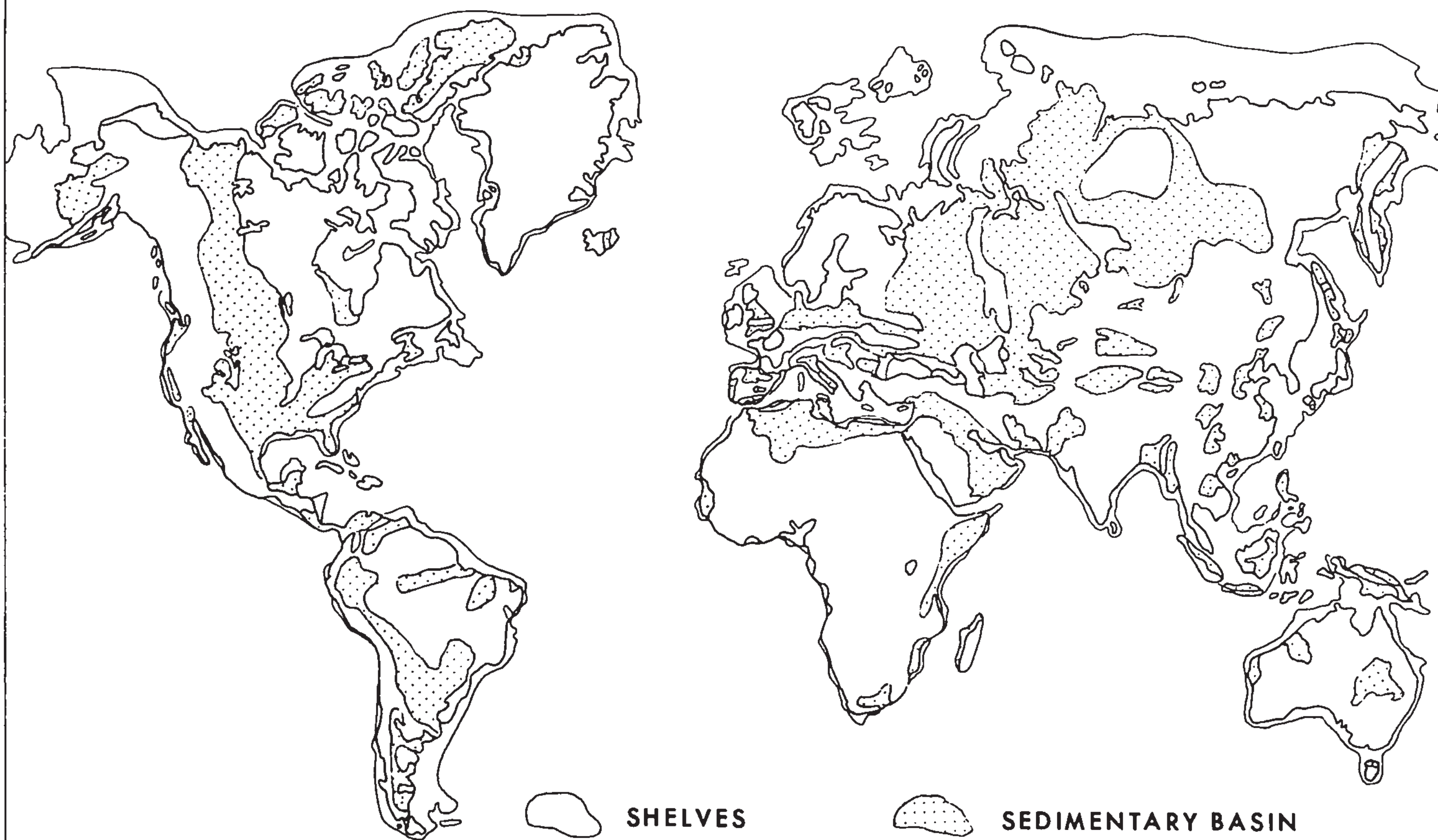


FIG. I-1

SEDIMENTARY BASINS AND CONTINENTAL SHELVES

(Adapted from Fairbridge (1957))

CHAPTER II

SOUTHWESTERN ONTARIO

PHYSICAL AND GEOLOGICAL DESCRIPTION

II-1 GENERAL DESCRIPTION

Southwestern Ontario is the peninsula of land lying between Lakes Huron to the north and Erie and Ontario to the south, and bounded by the St. Clair River and the United States border in the west and by the emergent Precambrian Canadian Shield in the east. It lies roughly between latitudes 42^{o}N and 45^{o}N and longitudes 79^{o}W and 82^{o}W. In land area it occupies some 68,000 sq. km and, if similar geology is taken as a guide-line, it could be said to cover a further 39,000 sq km beneath the Great Lakes. Topographically it is divided into three sections: in the west and east the land is between 150 and 300 m above sea-level, in the centre the land rises gradually to a maximum of 540 m at the south end of Georgian Bay, the highest points being close to the Niagara Escarpment, and a lowland around the shores of Lake Ontario averaging 90 m. The Niagara Escarpment, running across the peninsula northwest to southeast from the top of the Bruce Peninsula to Niagara Falls, is a carbonate cuestra resulting from the differential erosion of resistant Lockport dolomites and the soft underlying strata.

The Southern Peninsula of Michigan lies directly to the west of Lake Huron and the Southwestern Ontario area. It extends some 320 km to the eastern shores of Lake Michigan and a distance of 400 km north to south. Elevations range from 200 m on the eastern and western sides to 460 m in the northern central area. FIG II-1 shows the area in which the study was carried out on a map of North America.

II-2 GEOLOGICAL HISTORY

The Southwestern Ontario sedimentary basin is not actually a separate basin but forms a part of the flanks of both the Appalachian and Michigan basins (see FIG II-2). Important factors in the deposition of the sedimentary strata have been the two major tectonic features of the basement, the Algonquin arch dipping from the exposed Shield to the north and east in a southwesterly strike at approximately 6.6 m per km, and the Findlay arch which crosses beneath Lake Erie from the south and enters the western extremity of the peninsula. Since Late Cambrian times the Algonquin arch has shown a positive structure, evidenced by the truncation of even the Cambrian sediments against it. However, the Findlay arch shows little or no rise until the Middle and Upper Silurian. It seems that the arches have, in general, acted as hinges in basin movement. Between them is a depression known as the Chatham Sag which is faulted at its northern end against the Algonquin arch. This in itself may explain the apparent independent movement of the two arches. The basement is composed of metamorphosed sedimentary rocks of the Grenville series and is similar in nature to the rocks exposed on the emergent Shield in the northern counties of southern Ontario (Stockwell, 1965). Overlying this is a sedimentary sequence shown in FIG II-3, ranging in age from Upper Cambrian to probable Mississipian (adapted from Michigan Dep't of Conservation, 1964). As the figure indicates, this sequence, composed of limestones, dolostones and shales for the most part, is divided into some 30 formations ranging in thickness from 1.5 to 460 m. In the eastern area these sediments reach a maximum thickness of 610 m around Toronto and get progressively thicker to a maximum of 1,525 m around Sarnia and beneath Lake Erie, with a dip of approximately 2^{o}. Further to the west and south in the central parts of the sedimentary basin the thicknesses reach 4,575 m.

The sedimentary sequence in Michigan is similar to that of Southwestern Ontario, ranging from Lower Cambrian to Mesozoic in age. More recent portions of the sedimentary sequence have been preserved as shown in FIG II-3.

Structurally the Michigan Basin is roughly circular in shape with sediment thicknesses varying from 1,220 to nearly 4,575 m over the portion considered in this work. Very few holes have penetrated to the Precambrian basement, but it is believed that it is divided into Grenville and Central Province series, the Grenville front striking northeast to southwest through the city of Flint (McLaughlin, 1954). Boundaries to the basin are the Wisconsin arch in the west, emergent Shield in the North, Kanakee arch in the south and the Findlay and Algonquin arches in the east. FIG II-4 shows a section through the basin.

Oil and gas occurs in large quantities in Southwestern Ontario. Production amounted to one million barrels of oil and sixteen billion cu.ft. of gas in 1964, recovered from most of the Palaeozoic formations. The Devonian reservoirs are mostly related to the underlying slump structures, the Middle Silurian reservoirs are bioherms and reef structures, excepting in the Niagara gas fields, whereas the Middle Ordovician production is from dolomitized fracture zones. Reservoir characteristics may have quite a bearing on the temperature and heat flow data and will be discussed for some areas in greater detail in later chapters. Salt is also of commercial importance in the area, where it is found in sub-units of the Salina with thicknesses up to 215 m.

Oil and gas production is also important in Michigan amounting to ten million barrels of oil and twenty-eight billion cu.ft. of gas in 1964. Production horizons include some of the Mississipian and post-Mississipian formations, such as the Michigan formation sandstones, not found in south-western Ontario. Salt, gypsum and coal are other commercially exploited minerals in the area.

The entire area was covered by the Pleistocene ice-sheets which on their retreat deposited a blanket of boulder clay of maximum thickness 90 m in Southwestern Ontario but reaching a thickness of 300 m in the Michigan peninsula. Hough (1963) has suggested that the Great Lakes are a result of glacial scouring of river valleys. Lake Michigan is the deepest of the lakes bounding the region with depth of 290 m, while Lake Erie is the shallowest with a deepest point of only 60 m.

FIG. II-1 STUDY AREA AS PORTION OF NORTH AMERICA

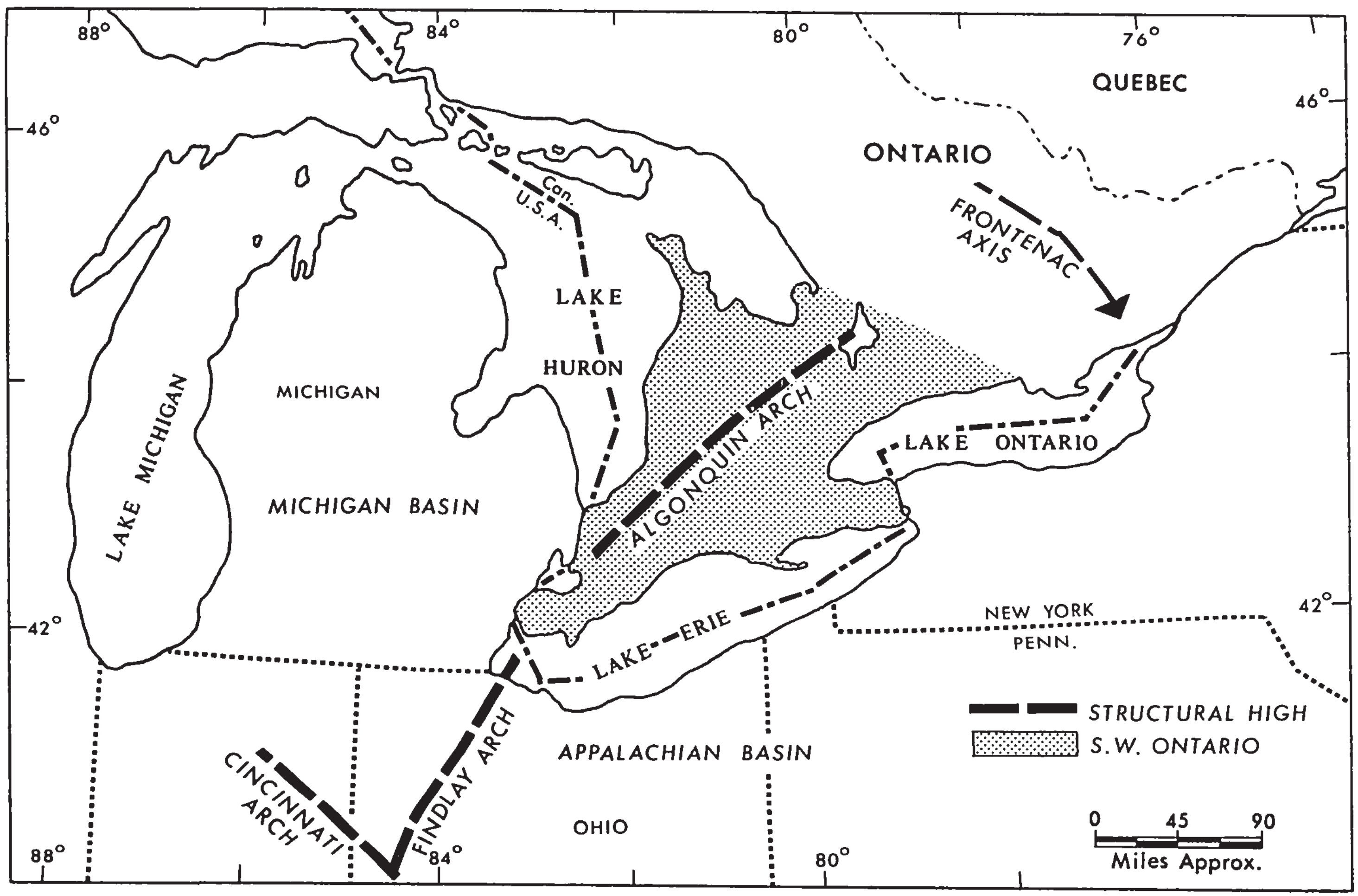


FIG. II-2
MAJOR STRUCTURAL FEATURES
OF EASTERN GREAT LAKES

FIG. II-3 STRATIGRAPHIC SEQUENCE IN SOUTHERN ONTARIO AND MICHIGAN

ERA	PERIOD	EPOCH	SERIES	GROUP	FORMATION
MESOZOIC	JURASSIC	L	KIMERIDGIAN		Red Beds
PALAEOZOIC	PENNSYLVANIAN	L	CONEMAUGH		Grand River
		E	POTTSVILLE		Saginaw
	MISSISSIPPIAN	L	MERAMECIAN	Grand Rapids	Bayport Michigan
		E	OSAGIAN		Marshall
			KINDERHOOKIAN		Coldwater
					Sunbury Berea Bedford Antrim
	DEVONIAN	L	CHATAUQUAN		
			SENECAN	Traverse	Squaw Bay Alpena Bell
		M	ERIAN		Rodgers City Dundee
			ULSTERIAN	Detroit River	Lucas Amhurstburg Sylvania
					Bois Blanc
		E			Garden Island
	SILURIAN	L	CAYUGAN	Bass Island	
				Salina	G F E D C B A
		M	NIAGARAN	Guelph	
		E	ALEXANDRIAN	Cataract	Cabot Head Manitoulin
	ORDOVICIAN	L	CINCINNATIAN	Richmond	Queenston Utica
		M	MOHAWKIAN	Trenton Black River	
		E	CHAZYAN		St. Peters
			CANADIAN	Prairie Du Chien	Shakopee New Richmond Oneota
	CAMBRIAN	L	ST. CROIXAN	Lake Superior	Trempealeau Munising
		M			Jacobsville
		E			
PRECAMBRIAN					

CHAPTER III

TEMPERATURE MEASURING EQUIPMENT

III-1 TEMPERATURE MEASUREMENT

A large variety of different methods have been used in attempts to log temperatures in boreholes. Many of the methods of temperature measurement are listed in NBS Monograph 27 and supplements (1961, 63, 67) and their advantages and disadvantages are discussed by Misener & Beck (1960) and Beck (1965). In the oil industry the traditional method, which is still used to measure bottom hole temperatures during other logging procedures, is the maximum thermometer. The maximum thermometer presents the most reliable means of measuring borehole temperatures, though at the same time it is less precise and more tedious and time-consuming than other techniques. The classic reference to this instrument and its use is Van Orstrand (1924). Accuracies of 0.1°C are obtainable with due care, but temperature inversions are not recorded. In the oil industry, the Amerada vapour pressure thermometer is also used, giving an accuracy of 0.3°C (Anglin, 1964). Similar accuracy is obtained with the bimetallic strip used by Marshall et al (1963) in hot spring studies. Well-logging companies operating in Southwestern Ontario use a thermistor tool with an accuracy of 0.3°C for a continuous log.

The thermistor sensing element is today the most common one in terrestrial heat flow studies and in terms of such factors as weight of equipment, cost, reliability and speed of operation, the thermistor was chosen as the sensing element for this work. Temperature measurements were made with a three-lead lightweight cable wound on a Sharpe GR 4000 portable winch. The thermistor was sealed in a pressure-tight probe and a modified Wheatstone Bridge circuit used to measure the resistance of the sensor. Both the shunt compensated relay type and the more usual 3-lead series compensation type bridges were used. Bridges, cables and probes are discussed in greater detail in Appendix III. Certain problems exist in the stability of thermistors which are discussed in Appendix A-II. Calibration of the sensors against a platinum resistance thermometer is described in Appendix A-II-5. Results from the calibration were printed out by the computer as tables of resistance versus temperature at 10 ohm resistance intervals. Thermistor calibrations were checked at two or three points before and after each field trip.

Resistance readings were taken at 30 m intervals in boreholes other than the U.W.O. one. Probes have time constants of 0.3 minutes or less in water. At bottom-hole the probes were generally left for 15 minutes to check for drifts in the readings. It was usual also to recheck one depth, usually 300m in deep boreholes in the return direction. Total on-site time for a 915 m borehole is about 4 hrs. if the hole is water-filled.

A programme was set up during the course of this work to make temperature measurements in wells as part of the ODERM well abandonment operation, which required equipment to be available within 24 hours of a telephone message.

III-2 DEPTH MEASUREMENT

Measurements were normally taken from the top of the casing, and the height of this above ground level noted. Positions of the 30 m markers were initially good to better than 1 cm as checks have shown. Permanent stretch of the light-weight cable may result in depth errors of 3 m in the probe position in deep boreholes. This stretching is essentially a property of new cables being used for the first time. However further stretching probably results from the cable being wound back on the reel under tension after running a temperature survey of a deep hole. Frequent checks of the distance between markers provide an adequate check on this factor. Elastic stretching is the major cause of uncertainty in estimating errors of depth measurement. Usually the bottom of the well appears to be a little shallower than the total depth given in the drillers records.

This problem can be overcome by noting the relationship between readings on the pulley counter and the appearance of the 30 m (100 ft) markers on the cable. As more of the cable disappears down the hole, the apparent distance between markers seems to become greater. At depths of 915 m, the stretch is about 0.6 m in the uppermost 30 m, but will be progressively less at greater depths. Since the cables are generally marked under slight tension, the total errors in depth may amount to -6 m at 915 m, and a slight positive error at very shallow depths (representing an error of 1.5% in 30 m gradients at depth, or 0.5% on the overall gradient). The above, of course, assumes liquid filled boreholes; in gas filled boreholes the effect would be very much greater. However since pulley counter readings are noted for each depth marker, a correction may be made. Misener & Beck (1960) deal with the theoretical aspects of estimating cable stretch.

Several types of well-head equipment were taken into the field, since there is little standardisation in industry on well-head arrangements. The most useful one was a cylinder 15 cm (6 in) in diameter which clamped over the casing and which had a pulley with an attached footage counter

III-3 RELIABILITY OF RESULTS

III-3-1 Temperature Measurement

The relay bridge decade was originally calibrated against the Pye standard bridge and so no thermistor interchange errors should arise; however, the 3-lead series compensated bridge was also used in the field. Comparisons of 0.005% L & N standard resistances on this bridge gave errors as high as 7.5 ohms in 50 kohms. An error curve was prepared and the required difference subtracted from the reading to give the absolute value. Where the thermistor had not been originally calibrated on the cable, the difference in the series resistance of the cable leads had also to be allowed for. This is usually 1 or 2 ohms for the light-weight cable and less for the heavier grade. Calibration accuracy was 0.003°C, as discussed in Appendix A-II-5, so that the deciding factor on overall accuracy then depended on thermistor drift. Most of the field work was carried out in one- or two-day trips. Since calibrations were checked before and after each field trip and drifts allowed for on a linear correction, errors should not be larger than 0.005°C. The longest trips were those to Michigan where a week passed between thermistor calibration checks. A very stable thermistor was chosen for this work, and drifts were not greater than 0.01°C between calibrations.

Summarizing, it is probably true to claim an accuracy of 0.005^{o}C between sets of data and 0.003^{o}C between individual readings. Absolute accuracy should be of the order of at least 0.01^{o}C. Occasionally very pronounced fluctuations were observed in boreholes, probably due to convective overturn. Such accuracy cannot be claimed for these horizons.

In temperature gradient measurements this may lead to a maximum error of twice the accuracy, ie., 0.006^{o}C. Since the lowest gradients, excepting inversions, measured were of the order of 0.20^{o}C over a 30m (100ft) interval the maximum error is about 3%. However most gradients will contain much smaller errors, and averaging over several intervals will reduce errors further.

III-3-2 Depth Reliability

Once the depth readings had been corrected for the discrepancy between the positions of the 100 ft (30m) markers on the cable and the apparent depths of these markers as recorded on the pulley counter, the total depth error in the borehole should be only a few metres. However the pulley and cable are calibrated against each other over the first 500 ft (150m) of the hole to establish the pulley count per hundred feet of cable. Any slippage of the cable over the pulley will increase the errors considerably. With care the error due to elastic cable stretch is quite small. Two larger sources of error are: the positioning of each 100 ft marker at the top of the casing in such a way that electrical connection to the measuring circuit may be made, and the accuracy of positioning the 100 ft markers in the original marking of the cable. The combined errors here may be as high as 6 in (15cm), which in a gradient measurement may mean a maximum depth error of one foot in 100 feet, or 1%.

III-3-3 Combined Errors

The maximum combined errors in the gradient measurements are expected to be about 4% in 100ft sections with very low temperature gradients. In zones with higher temperature gradients the errors will be proportionately smaller and where temperature profiles have been measured in detail such as the UWO hole, smoothing over several data intervals improves the accuracy further still.

CHAPTER IV

MEASUREMENT OF THERMAL CONDUCTIVITY

IV-1 THE MEASUREMENT OF THERMAL CONDUCTIVITY

The second factor to be determined in the estimation of terrestrial heat flow is that of the thermal conductivity of the rocks in the section. Of course the most desirable way of doing this would be to make 'in-situ' measurements. However the present probes give only a radial measurement ie., the thermal conductivity perpendicular to the direction of flow of the heat. Recent experiments by Wright & Garland (1968) also give an indication of axial values. Such methods only yield an accuracy of 10% even with fairly elaborate computerised curve-fitting techniques and long periods of measurement at each location in the borehole. The advantages and disadvantages of 'in-situ' measurements have been discussed at some length by Beck, Anglin & Sass (1971). Core samples were readily available in Southern Ontario and therefore for the purposes of this work it was decided that the divided bar presented the simplest and most accurate method of determining thermal conductivity. A few measurements using line source methods were made for comparison purposes, and are discussed in Chapter XI.

IV-2 DIVIDED BAR

Introduction

The divided bar has a very lengthy history beginning with the experiments of Clement and Peclet (1841) and Landsberg (1953). Modern systems stem directly from a modification of the Forbes bar suggested by Lodge (1878) and first utilised by Lees (1892). These methods are now widely used as the standard technique for measurement of thermal conductivity and thus large numbers of variations exist. Perhaps the most common in heat flow usage is the Beck system (Beck, 1957) in which the reservoirs at either end are maintained at constant temperatures, quartz discs are used to determine the thermal conductivity of the brass of which the bar is made and the rocks are compared to this.

IV-2-1 Divided Bar Used

It was felt that the above system led to a very long stack and hence to fairly great heat losses. For this reason it required special manufacture with an included guard ring to eliminate the losses. If the comparison standard in each measurement were of a higher resistivity the stacks could be made shorter and rather simpler.

Several bars were constructed as described in Appendix A-IV, the final one of which consisted of stacks composed of fused silica discs sandwiched between thin brass discs as shown in FIG IV-1. The effect of environment on the bar and its calibration are also described in Appendix A-IV, as are the calibration methods and reliability of the values.

IV-2-2 Interlaboratory Comparison

It was felt that since the divided bar is a relative instrument some very real systematic differences might exist between the values calculated on bars at different institutions. Since there are variations in the standards, a good method of standardising bars is to exchange samples. For this reason some samples of Misener et al (1951) were remeasured on the UWO bar and some specimens were measured both at UWO and at the Dominion

Observatory. The table below shows these comparisons, based on UWO as correct.

Leney (1956)	Misener (1951)	U.W.O.	Dominion Observatory	No. samples compared
		0	+2%	20 sets
	<±2%	0		5 sets
<5%		0		indirect

TABLE IV-1 INTERLABORATORY COMPARISONS OF THERMAL CONDUCTIVITY

At the time of these comparisons it was not known that the discs prepared and measured by Leney (1956) still existed. Therefore, to compare the divided-bars, his Detroit River conductivities were compared with values determined for the same formation in this work.

IV-2-3 Accuracy of Thermal Conductivity Measurements

At axial pressures of greater than 400 psi, with reservoir baths at ±5^{o}C on either side of the ambient temperature and with the centre point kept to within o.5^{o}C of ambient, it is believed that the divided bar is capable of giving thermal conductivities of discs of thickness approximately 0.2 to 1.5 cm and thermal resistance between 50 and 400 sec^{o}C/cal to ±2%. The bar was probably capable of detecting conductivity differences of 0.3% with very careful use.

IV-3 ANALYSIS OF THREE-DISC SETS

Standard procedure for the measurement of thermal conductivity of rocks on a divided bar is to use sets of three or four discs cut from the same specimen but of differing thicknesses. The least squares computed conductivity is used as the representative value. The major reason for this has been the uncertainty of contact resistance between disc and bar surface. It was felt that with a liquid contact and an axial pressure the resistance should be very small and show little variation from rock to rock. Should this be true, then the use of single discs would enable more detailed sampling of a geological section for a given expenditure of time.

To this end 30 sets of three discs apiece were prepared from a wide variety of different geological horizons in the southern Ontario section. Samples of thicknesses 4, 8 and 12 mm were processed in the same manner as that described for single discs in VI-4. Results were programmed on the IBM 7040 to write out computed conductivity and interpret the contact resistance on a least squares fit. Initially the latter was assumed to represent the contact resistance and the mean and standard deviation of the values found. These were $1.1 \pm 7.5(\text{mcals/cmsec}^{o}\text{C})^{-1}$, compared with the standard quartz calibration, a very large standard deviation indeed. Reasons for this are to be found, not, apparently, in the nature of the contact as might be expected, but rather in the conductivity contrasts between individual discs in a set and thus is a reflection of compositional differences. For example, FIG IV-2 shows the effect on the least squares intercept and the conductivity of altering each of the disc conductivities by 5 or 10% in sequence for initial values of .005 and .015, thus covering the range of most samples encountered in this work.

In the light of this, the disc sets giving the largest intercepts were examined to see whether, indeed, they did show large compositional variations. Set IV 3168 showed a 50% higher thermal conductivity in the thinnest disc with respect to the other two discs. Observation of all three discs under the microscope showed that, whereas the 8 and 12 mm discs contained traces of pyrite, a mineral with a high thermal conductivity, the 4 mm disc contained about 10% and the grain size was sufficient to cause 'parallel' paths. Since the conductivity of the 4 mm disc is larger than that of the other two, the thermal resistivity is lower, thus producing a large negative intercept. This is not always the case, however, as is shown by set M3A1700 which shows a large positive intercept, again caused by variations in the disc composition. The 4 mm disc contains dolomite interspersed with gypsum, whereas in the thicker discs bands of anhydrite, having a much higher thermal conductivity than gypsum, begin to appear. A final case is cited, that of RAGA2699, which behaves similarly to the first. The thinnest disc has quartz-filled fractures running through it, effectively creating 'parallel' paths since quartz has a high thermal conductivity, although the proportion of such paths is less in number than in the earlier set.

Varying the thermal conductivity in each disc of a hypothetical three disc set by 5%, one disc at a time,gave apparent contact resistances of between +8.0 and -6.0 (FIG IV-2). If all of the measured three disc sets which show conductivity contrasts of greater than ±5% between discs were now rejected, the resulting intercepts could be said to indicate a fairly constant resistance if all of the data lie within the limits of the model. FIG IV-3 shows that this is in fact so, the maximum deviations being -6.0 and +7.0 leading to a mean of 3.7 ± 1.5. Also, according to FIG IV-2, the standard deviations should get smaller as only those discs with a smaller conductivity contrast are considered. If only those samples with a conductivity variation of less than ±1% are considered, leaving seven sets to be included, the mean becomes + 0.3 ± 0.5 with reference to the quartz contact resistance. Thus it would seem that the contact resistance is probably a little greater than that of the quartz calibration curve. The rock sets included various lithologies, ie. granites, limestones, dolomites and shales, and thus the contact resistance quoted above may be considered as representative of such lithologies. Since it seems to show little change, a constant value is used throughout this work.

Finally, if single discs are to be used, it is necessary to consider what minimum thickness must be used to ensure that the sample is representative of the section. Obviously, as has been discussed previously, the 4 mm disc cannot be considered to be representative. Now limits are placed on the maximum thickness by the thermal loss conditions of the bar, particularly the ones of smaller diameter. For presentation on the ratiometer, an ideal thickness lies between 8 and 10 mm; this would ensure that all the conductivities met with in the sedimentary section would lie within the most sensitive portion of the ratiometer. Perhaps the easiest way to determine whether there are significant conductivity variations in this region would be to compare the 8 mm disc with the 12 mm one. The 8 mm disc shows a mean which deviates from that of the 12 mm disc by less than 1% with a spread of 5%. Any thickness chosed between those limits will then be at least as representative of the section as the 12 mm disc.

This section has shown that the contact resistance between the bar and discs of different types of sedimentary rock receiving the same preparation is sufficiently uniform and small, for a liquid contact and axial pressure, that single discs may be used. Further, that a single disc may be used for conductivity measurements with a thickness of 8-10 mm and may be as representative of the section as the 12 mm disc. It also points out the dangers of conductivity estimation using three disc sets of varying thickness, because of conductivity variations from disc to disc and the importance of grain size in the thinnest disc. It would be more representative to use three disc sets of discs of the same thickness. Since the lowest resistivities likely to be encountered would be about 50 sec^{o}C/cal, the remaining uncertainty in contact results in errors of ±1% in the thermal conductivity of higher conductivity rocks.

IV-4 PREPARATION AND MEASUREMENT OF SAMPLES

The details of this section are presented in Appendix A-V. The original core was reduced if necessary to one of the standard bar diameters and then discs one cm thick produced for measurement on the divided bar. All specimens were soaked in water before measurement in an attempt to simulate the in-situ condition as closely as possible, and then the thermal conductivity measured on the divided bar described previously. A computer programme was written to determine the thermal conductivity of each disc including a diameter correction where the disc was slightly oversize or undersize in relation to the diameter of the bar itself. Densities and porosities were also calculated for each disc. The final output gave thermal conductivities, densities and porosities measured at temperatures of between 22 and 26^{o}C.

IV-5 RELIABILITY OF THE THERMAL CONDUCTIVITY RESULTS

In section 2-3 and in Appendix A-IV-5 the reliability of the bar itself was discussed and ±2% decided upon as the error limits even at the extremes of thermal conductivities measured. The use of a constant contact resistance for single discs where in fact the values are variable should be encompassed within these limits for well-prepared sample surfaces. Thus the thermal conductivities of the discs should be accurate to ±2% at the worst in relation to each other. However their absolute values may be a further 2% in error due to uncertainties in the thermal conductivities of the standards.

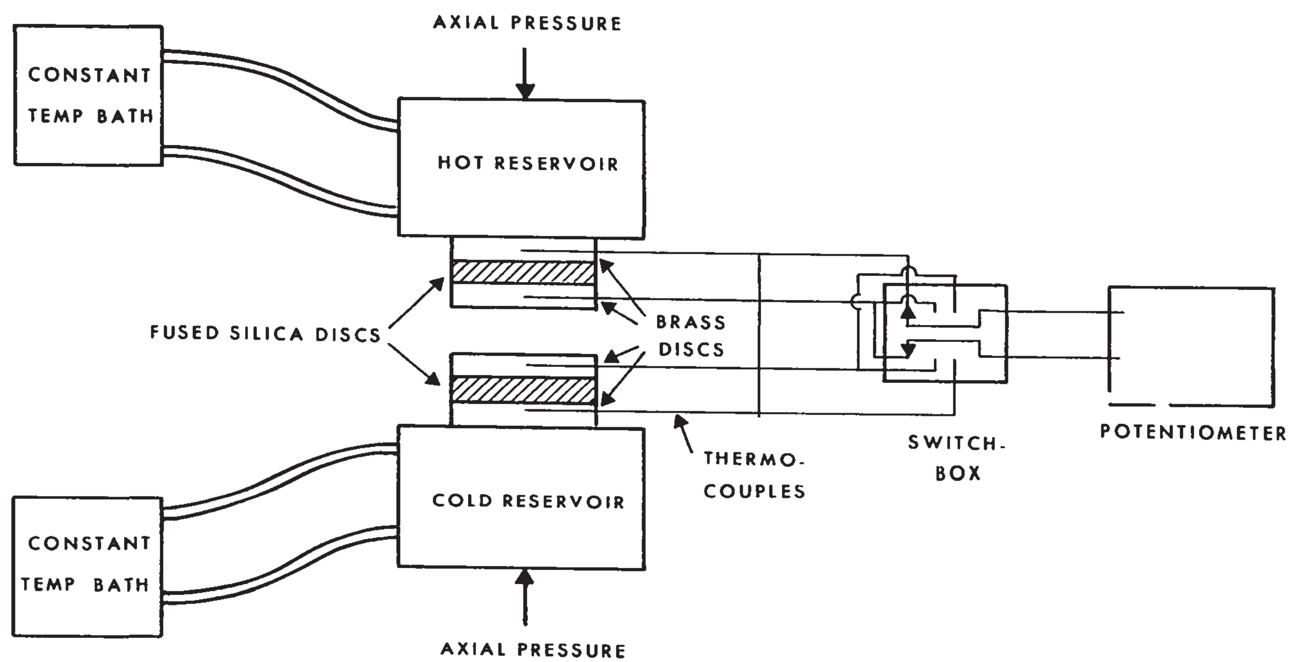


FIG. IV-1 CONFIGURATION OF DIVIDED—BAR

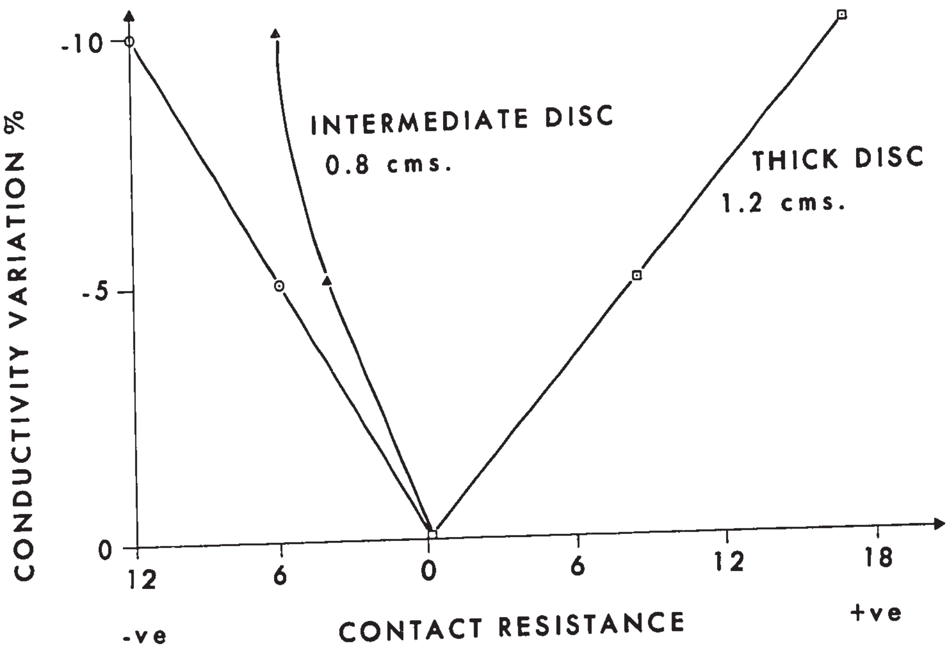


FIG IV-2 APPARENT CHANGE IN CONTACT RESISTANCE CAUSED BY CONDUCTIVITY CONTRASTS OF 5 TO 10% IN EACH OF THE DISCS

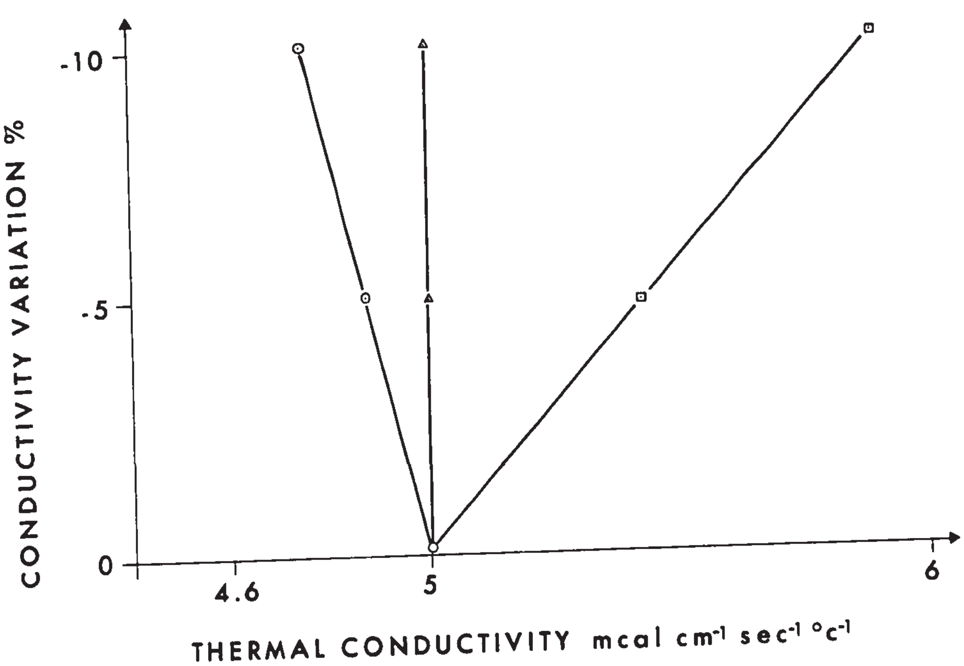


FIG IV-2 RESULTING APPARENT CHANGE IN LEAST SQUARES CONDUCTIVITY

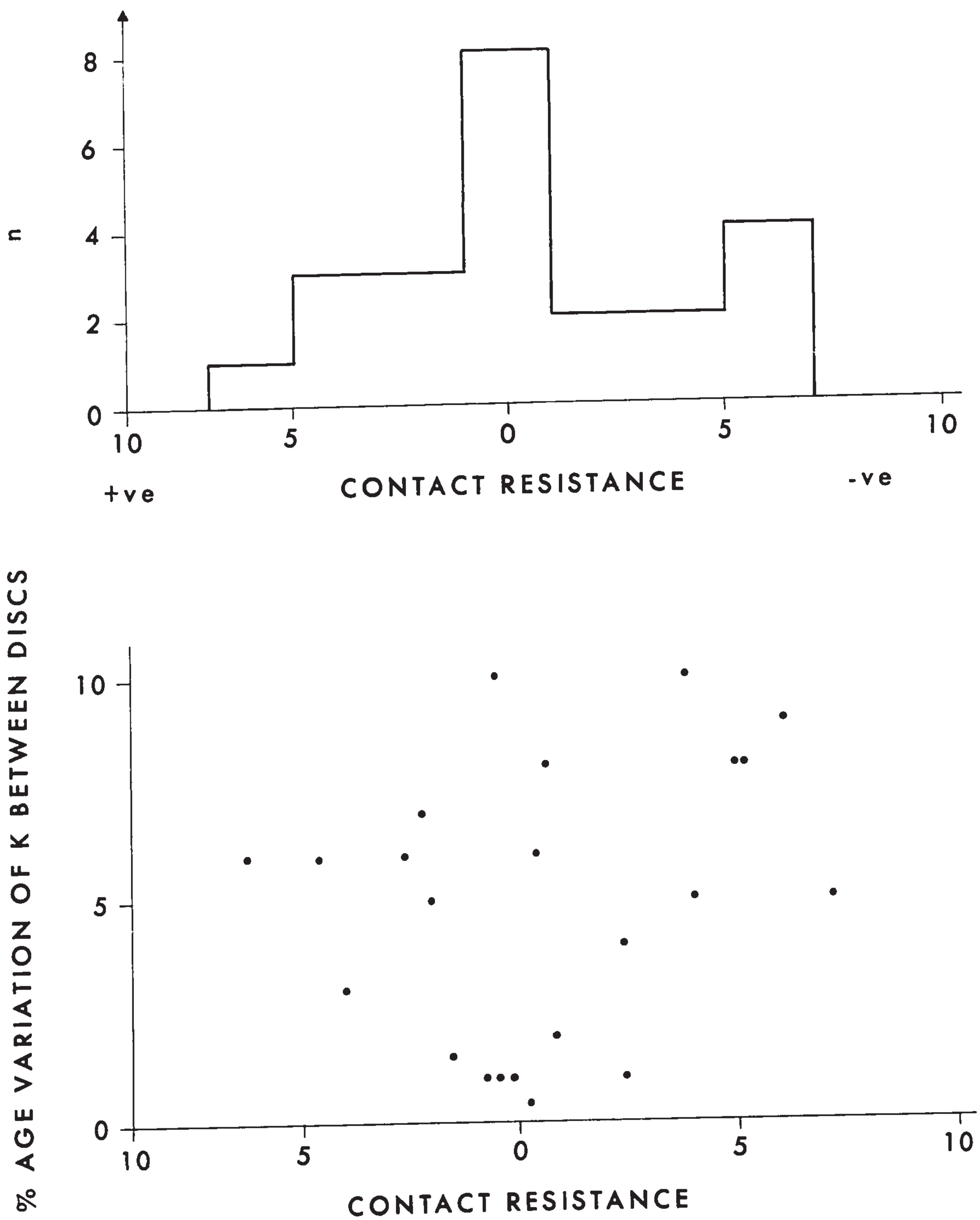


FIG IV-3

CONTACT RESISTANCE OF THREE-DISC SETS FOR CONDUCTIVITY VARIATION OF ±5% BETWEEN DISCS

CHAPTER V

SAMPLING

AND

RELATION OF DISCS TO BULK CONDUCTIVITIES

V-1 AVAILABILITY OF SAMPLES

Unlike mining areas in metamorphosed regions where drill holes are generally cored for assay, holes drilled in sedimentary areas are either not cored at all or are cored for porosity and permeability studies in potential oil or gas producing sections. These alone would not be complete enough or in sufficient quantity to do a very complete analysis of thermal conductivity variations. However, in the past six years several companies, notably Imperial Oil, Midrim Mines and U. S. Steel, have done a certain amount of diamond drill coring in stratigraphic test holes and it is these complete sections which have formed the basis of this work, together with the U.W.O. hole drilled by Canadian Longyear for the Dominion Observatory. These holes, shown in FIG V-1, are as follows: Enniskillen 20-5 (Chatham) to the Salina A-1 Subunit, U.W.O.-DOM. OBS. No. 4 (London) to the Meaford, MacGillivary 5-19 (Chatham) to Rochester, Plympton 1-3 (Chatham) to A-1, Argor No. 1 (Chatham) into Guelph, Midrim Mines No. 3A and 4(London) to A-1, U.S. Steel No. 1 (Ottawa) to Cambrian, and International Salt Company Harwich (Lansing)to Salina. As well as these, results were available for Delray well No. 2 (Lansing) to A-2 from Leney (1956), for C.N.E. well (unknown) to Black River from Misener et al (1952), for G.S.C. No. 1 to Black River and G.S.C. No. 2 (Ottawa) from Jessop & Judge (1971). The town name in brackets after the well name refers to the core storage location. Appendix A-VII details all boreholes in Southern Ontario and Michigan from which complete cores were used. Both Leney's and Jessop's bars have been tied in to the U.W.O. bar to determine whether significant differences existed in the measured thermal conductivities. Misener's C.N.E. samples have been lost, but some of his northern Ontario samples have been remeasured. Results are mentioned in Chapter IV.

Each of the stratigraphic test holes were sampled every 20-25 ft, fairly representative samples about 6 inches long of each section being chosen.

Partial core material was also available, stored in Chatham, Ottawa and Lansing, covering certain limited horizons, namely the Delaware-Detroit River, A_2, A_1, Guelph and Trenton-Black River. These cores are not completely representative since oil companies usually core in potential oil and gas producing structures. Only very rarely is the complete horizon cored (eg. in the Guelph the most porous horizons are often in the top 60 ft, hence this is most frequently cored as a potential reservoir). Most of the latter type of core is 4 inches in diameter and therefore must be over-drilled to bar size. Usually $1^3/8$ inch cores were taken from the centre where oxidation and salt leaching were not so far progressed. This latter precaution is very important where core has been in storage for up to 15 years under less than ideal conditions.

Finally, in a few instances quarry and exposure samples have been collected from outcrop areas. However, little reliance ought to be placed on these because of the extent of weathering and the fact that near-surface samples are in the region of active ground water circulation and hence alteration may be extensive.

V-2 SELECTION OF GEOLOGICAL HORIZONS

Naturally, the major governing factor is that of availability, followed by the need for measurements in horizons for which temperature gradients were obtainable. The study of conductivities was twofold in purpose; namely,

(a) to study the vertical and horizontal variation of conductivities,

(b) to obtain formation conductivities for heat flow determination,

with the emphasis not always on (b). For example, over 200 samples of the Detroit River formation have been measured. Generally, this formation is at quite shallow depths in southern Ontario and outcrops half-way across the peninsula. Hence, for (b) it is not the most valuable horizon, but for (a) it is ideal since much coring has been done looking for small oil and gas fields which are very cheap to bring into production. In description, the formation is quite similar throughout the area and thus useful estimates may be made on the value of core descriptions for extrapolating thermal conductivities. The spatial variation in thermal conductivity for different formations will not, in general, be the same, of course, since it is dependent on facies variations, or, in other words, on the environmental variations at the time of deposition and the geological history of the region since that time. For both (a) and (b) thin beds which pinch out are not very suitable and hence, the Clinton-Cataract formation and the Cambrian are not studied in very great detail. The former is compared in the U.W.O., P.W. 2, and U.S.S. No. 1 holes. Very shaley horizons are ruled out by the fact that they are not potential oil or gas producers and hence are rarely cored. These horizons include the Rochester, Queenston and Meaford-Dundas. It is unfortunate that these are not more frequently cored, since they are fairly uniform throughout the region. Three wells have been used to compare the Queenston: U.W.O., P.W. 2 and U.S.S. No. 1. Diamond drill cores have enabled a fairly intensive study of the section to be made from the Delaware to Salina A2 or A1. Partial cores have added to the Salina A2 and A1 data. Below this, the Guelph has been compared in U.W.O., U.S.S. No.1, McG 5-19 and Douglas No. 1, with partial sections in other holes. Trenton-Black River rocks have been compared for U.S.S. No. 1, C.N.E., G.S.C. 1 and 2, Colst. 2 with numerous sections, especially in the Colchester area.

V-3 COMPARISON WITH 'IN-SITU' POROSITIES

A fairly important problem is to determine to what extent the saturation can be considered to be similar to natural conditions. This has been ascertained on the basis of comparing the effective calculated porosity for the discs with porosity data obtained for complete cores soon after drilling from the same formation. The table below shows that the range of porosities is similar, suggesting that, in fact, the apparent porosities determined herein are significant. When it is recalled that for thermal conductivity studies very thin vuggy horizons are not sampled, whereas these are the horizons that are very important for oil or gas production, the similarity of porosities is very encouraging.

Delaware - Detroit River

A - Typical laboratory measurements (U.W.O.)

Well	Porosity Range %
Enniskillen 20 - V	1.4 - 6.0
Plympton 1 - III	1.8 - 12.4
U.W.O. No. 1	2.1 - 12.6
Midrim No. 3A	0.9 - 20.3

B - Typical industrial measurements (Core Labs., Hycalab.)

Sarnia 2 - 6A	1.5 - 13.8
Plympton 6 - II	1.0 - 13.0
Enniskillen 18 - I	2.7 - 28.5

Table V-1 COMPARISON OF POROSITIES OF DISCS WITH INDUSTRIAL MEASUREMENTS

V-4 HOW REPRESENTATIVE ARE THE DISCS?

Since cores of a diameter as little as 7/8" have been used, it is important to consider to what extent these discs are representative of the complete geological section. The section in the U.S. Steel hole show that for Trenton-Black River the sets of discs of 7/8" and 1-3/8" diameters agree well with each other although these discs show a rather coarse make-up. Several 4" cores of Guelph reef have been measured with a needle probe and later cored out to 1-3/8" and sets of discs were cut. Results of these comparison measurements showed differences of less than 5%. Several 6" lengths of the Detroit River formation from the U.W.O. hole have been measured, using a line-source method (Jaeger & Sass, 1964), by Sass and Anglin. These show similar conductivities for these lengths in comparison with the disc measurements although the extent to which the samples were saturated is in doubt. Thus, the results from the 4" diameter core down to a 7/8" diameter disc with a thickness of 1 cm would seem to agree within 5%. If the rock were a coarse-grained crystalline granite or pegmatite, these comparisons would probably no longer be true. Studies of the latter are at present in progress (Jessop, personal communication).

Generally good agreement is obtained, but it must be recalled that 1 cm discs are being used as representative of 6 m sections of rock. The validity of this assumption must be analysed. It is obvious that, if only one or two samples were chosen arbitrarily, very great errors might result; however, samples are chosen in conjunction with the geological log and selected to be representative as far as the unaided eye can distinguish. Where conductivities deviate considerably from the mean for a horizon, the reason is usually readily apparent under the microscope and the problem then is to decide what length of core on either side of the sample is affected. One way to look at the deviations is to take several well-sampled sections and look at then statistically. One of the formations showing the largest variations of thermal conductivity throughout its length is the Detroit River, the reasons for these variations not always being completely obvious to the naked eye. Below are shown the means calculated for single discs taken for one depth and for a 30 cm offset from this depth.

U.W.O. Section Detroit River 60-90m.

Set I	Set II	
7.8 ± 1.0	7.9 ± 1.0	: Average of ten depths

Set II displaced 30 cm down section from Set I.

Table V-2 CONDUCTIVITY COMPARISONS IN U.W.O. BOREHOLE

As can be seen from results in the above table, the averages are to within 2% of each other, whereas the individual discs in each section show maximum variations of up to 50%. This suggests that the sampling techniques used give errors no greater than 5% if they are used in conjunction with geological logs and a reasonable careful eye. Beck, Anglin & Sass (1971) have reported 'in-situ' conductivity measurements in some sections of the U.W.O. hole. Their results are compared to the divided bar measurements on single discs made by the writer (FIG V-2). Allowing for a 10% error in the 'in-situ' probe, the results are very similar. When conductivity interpolations are made with a lithological description of the core, the results are even better.

In general, one would estimate that the discs are representative of the borehole environment to ±5%. Should this be true, then accuracies of heat flux determination ought to be better than 10%. But this will be better revealed by a detailed examination of the conductivity results.

V-5 EFFECT OF 'IN-SITU' PRESSURE & TEMPERATURE

The measurements of thermal conductivity were made in a range of temperature between 22 and 26°C and an axial load of about 70 bars. However the 'in-situ' temperatures discussed in Chapter VI vary from 7°C at the surface to almost 82°C at the base of the Palaeozoic section in the Michigan Basin. This encompasses depth ranges of 0 to 3000 m. corresponding to lithostatic pressure differences of 0 to 300 bars. Clark (1966) lists data on the variation of conductivity with both pressure and temperature. The pressure effect is very small amounting to less than 2% on compressing the water-soaked rock to as much as 700 bars in the case of limestones and dolomites with porosities up to 13%. The maximum possible effect can be derived by determining the matrix conductivity of the porous rock which would occur beyond the crushing strength. Between 0 and 100°C the resulting decrease in conductivity may be as much as 22% of the conductivity at 0°C. Values derived from Clark are given below for a variety of rock types.

Writing the results in terms of the conductivity normalised to 0°C gives:

$$k_T/k_o = 1 + \alpha T$$

where k_T is the conductivity at T$^{\circ}$C above 0; then typical α's are given in the table:-

Rock Type	Coefficient $\alpha \times 10^3$
Sandstone Dolomite Pure Limestone	2.2
Carbonaceous Limestone	1.5
Slate	1.0
Granites	0.7 - 1.4

Table V-3 VARIATION OF k WITH TEMPERATURES 0 TO 100°C

Above 100°C the coefficient α decreases somewhat.

V-6 USE OF CHIPS AND SHORT CORED SECTIONS

In many instances the boreholes drilled in sedimentary regions are either not cored or are only cored at very porous layers, thus although it is possible to make quite accurate temperature measurements, it is not possible to obtain thermal conductivities. This aspect has certainly been the major reason for the small number of heat flow measurements in sedimentary regions. Although in Southwestern Ontario there were sufficient cored holes available to take a gross look at the conductivity variations across the region, any more detailed work will need some other method. All rotary and cable-tool drilled holes produce chips of rock and these are collected for geological analysis of horizons. Thus a method of making thermal conductivity determinations on the well-chips could be a substitute for the lack of core material. The simplest methods would use a mixture of the chips and a liquid of similar thermal conductivity to that of the rock. The measurements could be made with a needle probe, if large quantities of sample are available, or with small cells on the divided bar if only small amounts can be obtained. Of course the major disadvantage of this method is that the conductivity of water is only a quarter that of the average rock and there are no other liquids at room temperature which are suitable. However many fairly unstable inorganic salts exist which melt at fairly low temperatures not much above room temperature. A few of these were tried but problems arose with degassing the liquids, preventing dissociation of material and a tendency towards being hydroscopic. The technique used was to make up discs of the mixture of the salt and of rock chips suitable for divided bar measurements. Yet a further technique is to embed the dhips in a thermal conductivity epoxy and thus make discs suitable for use on the divided bar. The experiences with each of these methods are related in Appendix A-VI.

The aspect of recreating porosity and saturation in a rock from the chip results is also discussed in this Appendix. With the aid of density and porosity logs from the original borehole this is not difficult.

However the chips collected are not completely representative of a horizon since they are always heavily contaminated with chips from higher levels in the hole. Sobzchuk, Weber & Roots (1970) examined the use of chips to determine rock densities which are required to interpret gravity results in the arctic. The results of measurements on one hole, shown in FIV V-3 illustrate the point well. The chip densities were those determined on the drill-chips recovered from the hole whereas 'Nafe & Drake's' densities and'Woollard's' densities are determined from down-hole geophysical logs. Only very gross measurements can be attempted using drill chips.

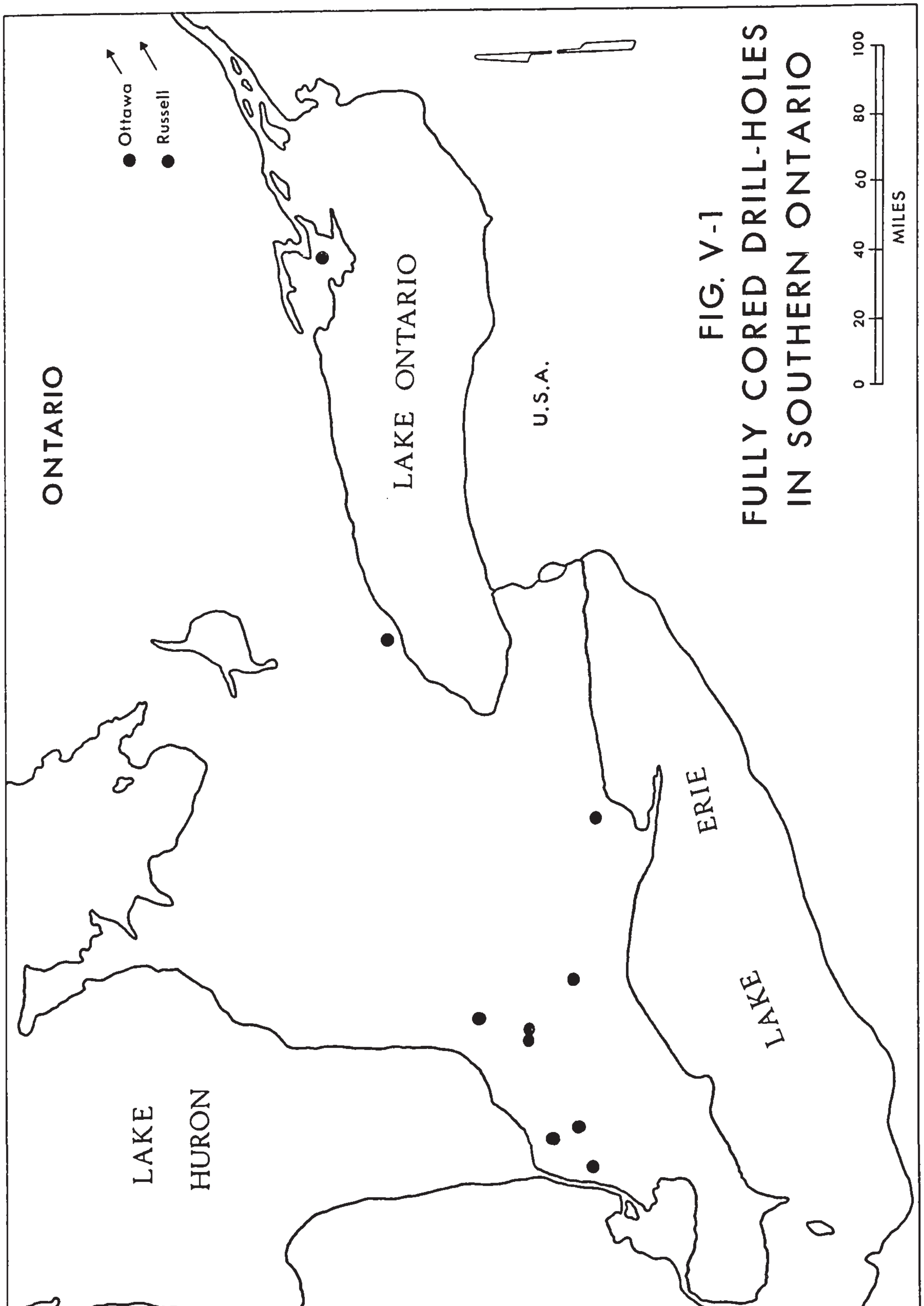


FIG. V-1
FULLY CORED DRILL-HOLES
IN SOUTHERN ONTARIO

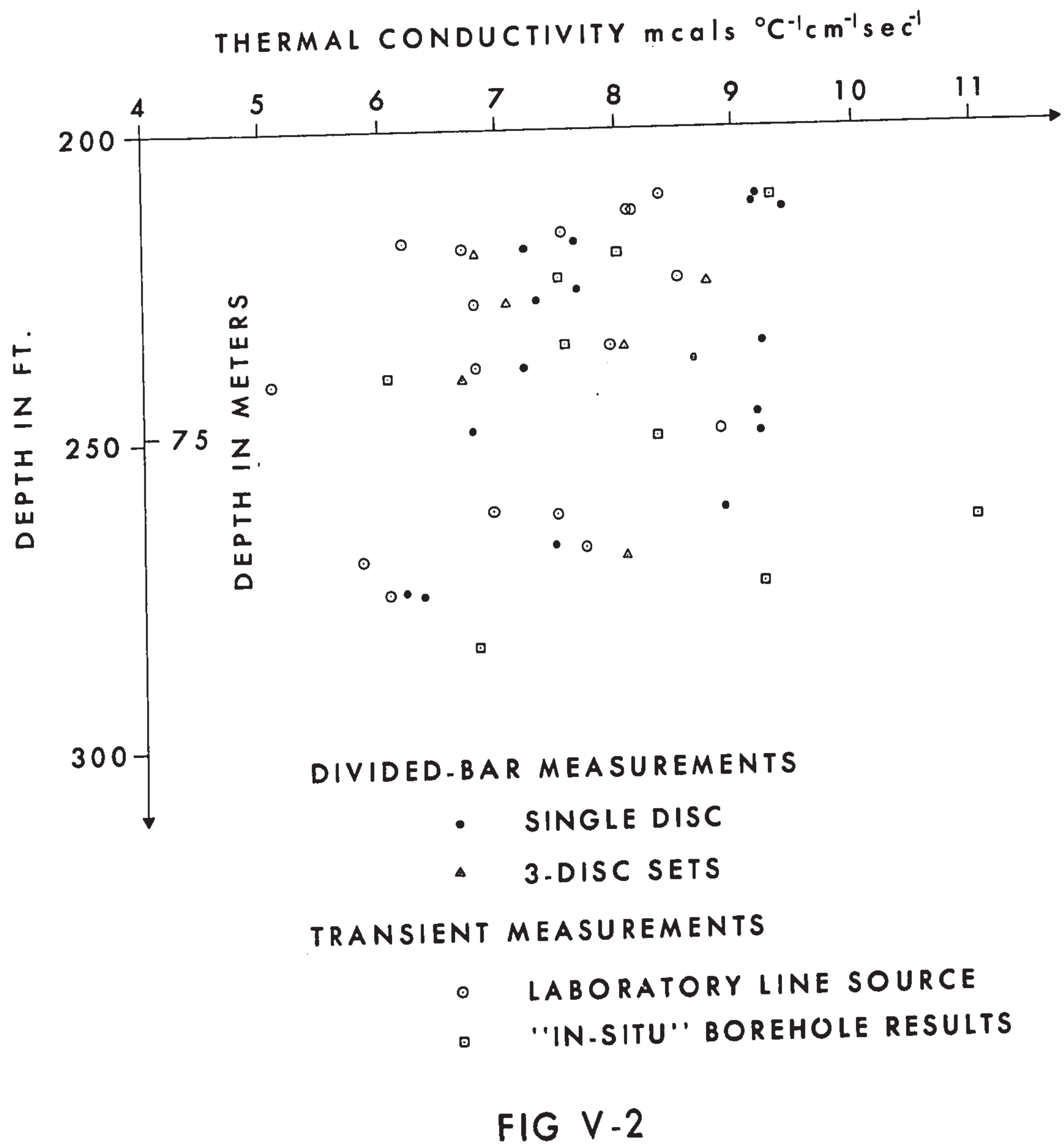


FIG V-2

COMPARISON OF LINE SOURCE, IN-SITU AND DIVIDED-BAR CONDUCTIVITIES INTHE LONDON HOLE

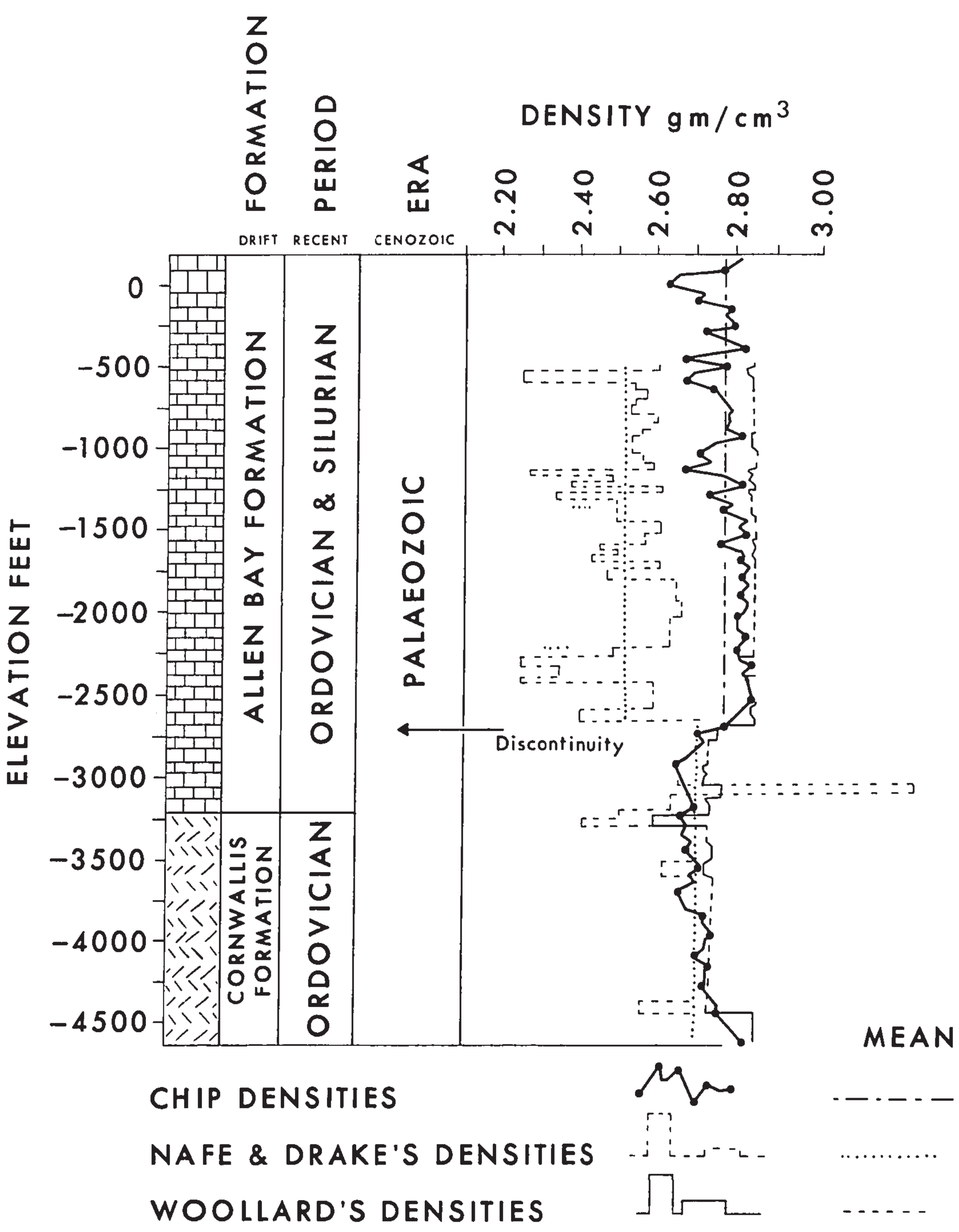


FIG. V-3 COMPARISON OF CHIP DENSITIES WITH WELL LOG DENSITIES IN A BOREHOLE

CHAPTER VI

THERMAL FIELD IN THE REGION

VI-1 AVAILABILITY OF HOLES

ODERM regulations require that holes be plugged immediately after drilling unless suspension for further development work is requested. If production is not expected, companies are loathe to request suspension and the casing is withdrawn and the hole plugged while the drill crew is still on the site. Measurement of temperature at this time has very little significance in rotary-drilled holes unless the history of the drilling disturbance is known. Bottom hole temperatures in diamond-drill or cable-tool holes probably would be within 0.1°C of the virgin rock temperature a few days after circulation is stopped. Here is a possible method of expanding the numbers of holes, by giving logging equipment of a very simple nature to drillers and having them log after a weekend's halt in drilling. Potentially, for this survey the most likely wells for use are: those suspended for further development where a period of several months could be allowed to pass after cessation of drilling, old producing wells now shut down, observation wells close to fields, and wells drilled before plugging regulations came into effect. The requirements decided upon for the purposes of this survey were framed in a letter to all oil companies and organizations which might possess or know of suitable wells. Extracts from the letter are quoted below:

"Since lightweight electrical cables are being used, we cannot tackle wells under pressure. However, if the well-head pressure is less than 40 lbs/sq in, and the well can be blown down and stop flowing, measurements the following day are little disturbed. For disturbed wells in general the following time spans should be followed:

a) recently drilled hole, log at least one month after completion,

b) old spent producer for several years' production, log at least same period after shut-down,

c) short-term producer, log at least several months after shut-down,

d) cleaned-out hole, log after period of few days or more, depending on estimated disturbance."

To assist in the location of holes, a hole-seeker programme was written to search the ODERM well-file tapes for wells which might be utilized, namely Observation, Suspended, Temporarily Abandoned, Oil Show Abandoned, Gas Show Abandoned, and Discovery and Abandoned. Constraints applied were that the well was drilled since 1950, was over 305 m deep, and had no record of casing being pulled or the hole abandoned. One useful well showed up, the rest merely had the plugging data missing on the well-cards, and the one observation well would have been too heavily disturbed.

In an attempt to determine what time interval passed between permit issue and completion of drilling, one year's ODERM drilling sheets were examined. It was hoped that these statistics might assist in the location of interesting wells for possible suspensions. Resultant histograms are shown in FIG VI-1. These suggest average drilling rates of 30m/day if drilling begins as soon as the permit is issued. However, there are a large number of wells which must have been drilled phenomenally quickly. Reference to the Figure will illustrate this. It also shows the preponderance of Salina wells with the peak in the 610-762 m histogram.

Time intervals suggested then are 2-6 weeks after permit issue for a check on well status dependent on total depth. For the purposes of this initial work, it was felt that this type of check would be too time-consuming for other than a long-term project, and so the idea was put aside.

In Michigan, gas storage fields have been operating for several years, providing convenient brine observation wells around the storage fields. Many of these boreholes are fitted with floats and are cased down to the storage horizon. Water levels are read every week in these wells, and so we were able to choose the ones furthest from the active field with the least disturbance. If the well is linked to the active field, water levels change between time of maximum pressure in the fall and maximum depletion in the spring. The wells will be disturbed eventually by the conducted heat from the hot gas pumped into the field. Disturbance propagation is shown in FIG VI-2 after various intervals of time.[1] Thus, with no gas breakthrough, the wells beyond this propagation interval should represent the true rock temperatures. Where possible, several logs were made in the same area to check local consistency of results.

The most knowledgeable people to approach for well information in Ontario are the ODERM Inspectors. They know where the holes are, but they do not know what the pressure might be at the well head, whether the well will blow-down or whether children have dropped lumps of wood or the like, down the well. This material drops to the fluid surface and blocks the well. Generally, it was found that, if a well is blocked at some point, it remains blocked although the blockage may be moved down several hundred feet. However, included in the field equipment

[1]Calculated using Carslaw & Jaeger (1959) p. 60.

was 2000 ft of 1/8" aircraft wire with a piece of 6 ft steel stock for pounding at blockages.

All together, some 30 holes were logged in the course of this survey, out of over 100 holes that were investigated. If the extra wells investigated by individual Inspectors are included, the useful proportion would be much less. Saul et al (1962) managed to log 3 out of 80 wells investigated in the St. Lawrence Lowlands, which gives this work a good average in comparison. It would seem apparent that future surveys in sedimentary areas must tackle shut-in gas fields, which, of course, would require much heavier equipment, power winches, etc.

A fact which bodes well for the future is that, because of the new Trans-Canada pipeline through Michigan, many of the older, almost exhausted, fields are being converted, or considered for conversion, to storage fields. If the pattern in Michigan is followed elsewhere, this should provide some observation wells suitable for heat flow determination.

VI-2 DESCRIPTION OF TEMPERATURE LOGS

The objective of this section is to divide the boreholes logged into regional groupings. For each of these regions a brief description of topography, geography and geology, both lithological and structural, will be given. Where these boreholes are in areas of active or past oil production, this aspect is discussed, since such activities could have substantially changed the temperature field. Finally, the temperature logs are discussed in each individual case and any regional trends or similarities pointed out. FIG VI-3 shows the area of the individual regions.

Most numbers in the text are given in the c.g.s. system. To convert to the English system of units the following conversion factors may be useful:

1 metre $\equiv$ 3.28 feet	$1^{o}C/km \equiv 0.055^{o}F/100ft$
1 kilometre $\equiv$ 0.62 miles	1 hectare $\equiv$ 2.47 acres

Gas pressures and production figures are left in the conventional English units.

VI-2-1 <u>Region 1</u> <u>Eastern Ontario</u>

Temperature results have also been obtained at three sites in Eastern Ontario and since this is the nearest of the regions to the outcropping shield it is treated as Region 1. All of the measurements were made by members of the Dominion Observatory including the author. The equipment used is very similar to that used for Southwestern Ontario so no large discrepancies are likely to exist. As a major difference between most of the sites in Southwestern Ontario and these, it is important to realise that all of the Eastern Ontario ones were drilled for stratigraphic information. The Russell and Picton holes were drilled by the Geological Survey of Canada in 1964 as stratigraphic test-holes in the Ordovician, the Franktown hole by the Observatory section searching for meteorite craters in 1962. Finally the Ottawa borehole was drilled in 1962 as part of a programme to make heat flow measurements at each of the seismic stations in Canada. The borehole sites are shown in FIG VI-4, all being within the Eastern Ontario sedimentary basin. The hole drilled at Picton by the Geological Survey of Canada should in terms of distance be incorporated into this section although geologically it belongs to the Michigan Basin sediments. Other than the Picton hole the data are included here to give a much wider picture of thermal conditions across Southern Ontario.

The Eastern Ontario basin is described by Sanford (1962) as having an areal extent of 11,650 sq km, and is bounded by the Ottawa River and the Frontenac axis along which the Precambrian outcrops. Maximum thickness of sediments is about 914m and the basin is mostly composed of sandstones, limestones, dolomites and shales of Upper Cambrian or Ordovician age. Faulting has considerably altered the original flat lying beds. The topography of the region around the borehole is fairly gentle (termed 'weakly broken' by the Ontario Department of Lands and Forests), varying between 76m and 152m above sea-level. To the north and west of the region the land rises in an exposed Precambrian dome to heights of 518m. Overburden varies between deep loams and clays of glacial origin to over 130m in thickness and Lacustrine deposits from post-glacial lakes.

Comparison of the temperature logs in FIG VI-5 shows very large differences in both temperatures and temperature gradients between the sites. Final logs were made three years after borehole completion. The surface temperature intercepts lie between 7.5°C and 9.3°C with all of the holes expect Picton showing a shallow inversion in the temperature curves, although the Picton site does show curvature in the upper portion.

Both the highest near surface temperature and the lack of inversion in the latter hole probably reflect its position close to Lake Ontario and the moderation of climate which results. Only the Ottawa hole penetrates into the Precambrian granites far enough to determine a temperature gradient, which is 14.1°C/km. The Franktown hole as described by Jessop (1968) penetrates 268 m of Precambrian dolomites and limestones with a mean gradient of 9.0°C/km. The lower section of the Russell hole and that at Picton show similar temperatures and gradients. The Trenton gradient at Picton is 24.6°C/km. It is not possible to determine a Trenton gradient in the 183 m of Ordovician sediments in the Ottawa borehole because of disturbances due to water-flows (Jessop & Judge, 1971). A small gas field with production from the bioclastic limestones in the Sherman Falls and Kirkfield formations of the Trenton Group was operated at Picton between 1937 and 1940 although the only commercial production was about 1,000 Mcf produced from 14 wells in 1940.

VI-2-2 Region 2 Toronto Area

This region covers Toronto and vicinity and the area as far as Guelph to the west. Two new holes were logged and temperature measurements were also available for the C.N.E. well in Toronto as reported by Misener et al (1951).

New temperature logs were obtained in a borehole in Chinguacousy township and to a depth of 610 m in a hole near the village of Morriston in Puslinch township. These are shown in FIG VI-6 together with an industrial log from Esquesing township. Locations are shown in FIG VI-7. The former borehole was originally drilled to 598 m early in 1962 and subsequently deepened to 610 m in November 1962, and finally to 755 m in September 1963 using a diamond drill. Until the time of logging in 1966 the borehole remained shut-in and at the time of logging still had no well-head pressure on it. Temperature measurements were conducted prior to the final abandonment of the hole by ODERM as part of their plugging programme. It is believed that this borehole was thermally very stable and the temperatures are probably among some of the most reliable measurements made in southern Ontario. The formations encountered in the borehole range from the Ordovician Meaford-Dundas shales to Precambrian gneisses, with the hole penetrating 213 m into the latter. Some temperatures were also obtained in a further hole which penetrates the basement. This hole was drilled close to the village of Morriston to a total depth of 1762 m in a hunt for Precambrian oil, and as a result penetrated 1054 m of the basement rocks. Unfortunately the borehole probe broke down on each of the several visits to this site and no complete log could be made. However Consumer Gas Co. had a temperature log made ten months after completion of the hole which agrees to 0.25°C with the measured temperatures in the section of the hole between the water-table and the probe break-down.

The Chinguacousy hole was drilled on the very flat plain to the east of Toronto on which Malton airport is also situated and as such there is less than 10 m topographic variation for some miles. However the Morristown hole although having little topographic relief around it is on the Niagara Escarpment some 65 m higher than the former. Drift cover is 15.2 m and 21 m respectively.

Water level in the Morriston hole is at 533 m and below this the mean temperature gradient to the top of the Precambrian is 25.6°C/km. Through the top section of the basement the gradient reduces to 12.8°C/km between 732 and 1341 m. Below this it averages 9.8°C/km, in a zone of a suspected small up-hole water flow, but which must await a more accurate survey for a complete explanation. The temperature gradient through the Precambrian section in the Chinguacousy hole is 15.4°C/km, which is 20% higher than that at Morriston. Trenton formation gradients may be compared between Chinguacousy and the C.N.E. borehole. At the former the mean gradient through the Trenton-Black River Group is 18.0°C/km, compared with 16.1 in the latter. Now the latter well is close to the shoreline of Lake Ontario and thus if the mean annual temperature of the lake bottom is warmer than the surrounding land, the borehole gradients must be increased slightly as discussed in VI-4 to obtain the equilibrium values.

Surface intercept temperatures of 6.5°C and 9.3°C in these holes are to be compared with mean annual ground temperatures of 10.2°C at Toronto. (Aston, 1969)

An industrial log obtained in the region is also shown in FIG VI-6. The temperatures given by it are much higher than might be reasonably expected although the Trenton gradient of 21.0°C/km is reasonable in comparison with other locations.

VI-2-3 Region 3 Niagara Peninsula

A unique opportunity arose to look at the possible use of bottom-hole temperatures as a structural indicator. Consumer Gas of Toronto had opened large numbers of old wells in the Niagara peninsula in the early 1960's. In each one before final abandonment they had first measured bottom-hole temperatures. Commercial well-logs were also available for four holes with total depths of 914 m or more. The positions of these latter holes are shown in FIG VI-8. Each of the logs was made some years after completion of the original hole and although it cannot be said with certainty that the boreholes were in equilibrium since they were reopened and cleaned out prior to these logs being made, the disturbance caused by this type of operation should be fairly small. Examination of temperatures and gradients suggests that, in fact, the temperatures and gradients are quite consistent and thus the logs are likely equilibrium ones. The formations penetrated by the boreholes range from dolomitic limestones of the Bass Island to the Cambrian sandstones. In the northernmost hole the Queenston forms the bedrock.

The temperature logs show, in FIG VI-9, wide variations in temperature close to the surface with similar variations in the surface intercept from 7 to 13^{o}C. However at a depth of 975 m the temperature spread between the three most southerly holes is less than 1^{o}C. At intermediate depths the holes show a progression of temperature from north to south across the peninsula with the highest in the north. The considerably higher temperatures at shallow depth in the most northerly of the holes reflects the shales close to the surface where the other holes have a section of Guelph or

Salina carbonates. The temperature gradient at the base of the Meaford-Dundas which is much greater than those in any of the other holes is believed to represent compositional differences. However if all formation gradients in this borehole are greater than the others this would suggest non-equilibrium conditions. Comparison of least squares gradients in the holes gives very similar gradients in the two inland wells; 17.7°C/km and 19.7 between Crowland and Bertie.

When the formation gradients are compared the two Bertie township wells have values of 17.7 and 19.0°C/km, the Crowland well 16.4 and the Louth 14.8 in the Trenton-Black River. It is interesting to note that this is a progression of higher gradients towards the south. No geological material is available to show why this might be so. Caley (1940) describes the major structural feature of the Niagara Peninsula as monoclinal with the Silurian formations dipping southward across the region at an average rate of 5.3m/km and this dip is shared by the Ordovician formations. In general the Paleozoic formations have been so little deformed that they appear flat-lying at outcrop. Sanford (1961) gives no lithological data for the Peninsula although his diagrams do show a thinning of the Trenton formation from north to south.

Comparison of the Meaford-Dundas gradients suggests much higher gradients in the Louth, the northernmost well,which is close to the area of the greatest thickening in the formation, as described by Sanford (1961). This may coincide with a far greater shale content. Queenston and shallow formations yield similar gradients in the other boreholes whereas in the Louth hole the former formation composes the bedrock.

As mentioned above over 400 bottom-hole temperatures were available for this region, mostly between depths of 183 and 305 m below ground-level and mostly in Bertie, Wainfleet, Humberstone, Caistor and Crowland townships. Most of the holes are old gas producers, and the shallow ones of the depths mentioned penetrate to the Clinton, Grimsby or Whirlpool horizons which provide the bulk of the production. The data was collected as a general part of the survey for transfer of some producing areas to storage. Caley (1940) has pointed out that the structure is not the principal controlling factor in gas accumulation in this region. It is probable that porosity, permeabilitytraps and differential cementation form the major part in accumulation. The reservoir rocks range from sandstones to dolomitic limestones. Koepke and Sanford (1965) place all of the production in Bertie, Humberstone, Crowland and Willoughby townships together as the Welland field which, since its discovery in 1889, had produced to the end of 1964 some 51,000 Mcf of gas over a region of 37,300 hectares with an initial pressure of 525 psi. This initial pressure however varies greatly from well to well. Plotting the bottom temperatures versus depth shown in FIG VI-10 shows a variation of 10°C. If these temperatures are plotted on a regional map several very interesting features appear. The number of points are not well enough spaced to make definitive statements and the results are fairly scattered as might be expected in this type of survey. However several zones of consistent temperature patterns appear which must be related to the local rock types as well as to production. Comparison with topography also seems to show some similarities. Results are shown in FIG VI-11. It is possible that such localised studies would, with more care taken in temperature measurements and a few limited cores for conductivity measurements reveal very useful information in the form not only of reservoir temperatures but also indirectly of geology and structure. But this must await greater industrial interest.

VI-2-4 Region 4 London Area

A single borehole was logged in the vicinity of the Willey Pool in Dunwich township, Elgin County. This pool, with production from a thin section of sands and dolomites in the Cambrian, was discovered in March 1965 and since that time had produced 376,000 barrels of oil and 500 Mcf of gas to September 1968 with bottom pressures of 1400 psi. The pool is located in the north-west corner of Dunwich township some 32 km south-west of London.

Before the main pool was discovered some production was obtained from well 16-I but salt-water encroachment shut off production. The well logged here is 17-I drilled the same year but closed at 1006 m with cement; since this time it has been used as a calibration hole by well-logging companies. It lies some 3.2 km west of the producing field and hence of the major structure. The region is on a positional high in the Precambrian on the south-east flank of Algonquin Arch, of which the producing field occupies 405 hectares. The positions of the boreholes drilled in the field and the contours on the Precambrian basement are shown in FIG VI-12 taken from Hancock (1967). Controlling features appear to be faulting through the Palaeozoic section. The post-Salina section is much increased in thickness by solution of salt and slumping.

In general salt does not occur within 13 - 16 km of the producing pool.

During the early programme of the Dominion Observatory a borehole was drilled by the D.E.M.R., Dominion Observatory, as a programme to make heat-flow measurements at each of the seismic observatories across Canada. This borehole was drilled on the campus of the University of Western Ontario, and although expected to reach 610 m, it had to be abandoned at 594 m. Since the hole was completely cored it was studied in great detail and is written up in Chapters X and XI. For the purposes of this section general comparisons only are given. Very little deep drilling has been done of this order around London; the deepest being at Crumlin about 13 km east of the city.

A further well was obtained in South Walsingham township about 64 km south-east of London. The well had been sitting unknown for upwards of 40 years and was logged prior to plugging by the ODERM inspectors. Over the years some oil leakage had occured from shallow depths and into a creek.

Topographic aspects are as follows: the J. Brown well in South Walsingham lies amid heavy bush below a small ridge beside a creek, whereas both the U.W.O. and Dunwich wells are on cleared land, the former on a cleared slope leading down to a creek in one direction and the Thames river in the other, the latter on higher flat land about 0.8 km from the Thames river. The borehole temperatures are shown in FIG VI-13 and show quite varied intercept temperatures. Both the Dunwich and Brown wells show no near surface inversions in the temperature curves, whereas the U.W.O. hole shows a very distinctive one. Respective surface intercept temperatures are 8.7°C and 10.8°C. Why these differences should occur is difficult to understand. In Chapter XII the near-surface inversions are explained as climatic effects which might be expected to affect most boreholes.

The formations penetrated in the Brown well are from the Dundee to the Clinton with 99 m of drift at the surface. In the Dunwich well the formations encountered are from the Dundee to the Trenton with 74 m of drift. The London well penetrated from the Dundee to the top of the Meaford-Dundas with 32 m of glacial drift. Relatively little geological material has been written up for this region outside of that of Caley (1943) which needs updating. Few holes have been drilled in the vicinity of London itself as mentioned before. However the stratigraphic information available has been gathered together and is discussed in Chapter XI. Since a fairly large number of conductivities were measured in this region the geological description is left to be dealt with in detail in Chapter VII, except to point out that the formations occur at greater depths in the Dunwich well which is down dip of the other two wells.

Temperatures are highest in the Brown well and lowest in the U.W.O. well with a 3°C spread at a depth of 427 m. Only the Dunwich well is deeper than 610 m and at bottom-hole (1006 m) reaches 26°C. Comparison of the temperature gradients in the three holes shows a higher least squares gradient in the Dunwich hole, which remains so even over the upper 610 m section. The comparative gradients are 11.4°C/km at Dunwich, 9.5 at South Walsingham and 7.9 at London. Formation gradients vary from nearly 29.5°C/km in the Meaford-Dundas shales for the Dunwich hole to lows of less than 6.6°C/km in the Salina section of the U.W.O. hole.

The Queenston shows gradients of 16.1°C/km at Dunwich and 15.7 at U.W.O. Comparison of Guelph gradients shows both the Dunwich and Brown wells with 7.5°C/km whereas U.W.O. is less than 6.6.

A fairly large number of bottom-hole temperatures were available for the area around the Brown well. In cases where two measurements have been taken of bottom-hole temperature, the later one is used here. Thus three wells in Lot 9, Conc.B give bottom temperatures between 14.7 and 14.9°C, the well in 5-B logged here gives 14.7°C and a well in 18-II gives 14.7°C at depths of between 415 and 421 m. A further well in 9-B gave bottom temperatures of only 11°C at 418 m on two successive runs. Thus the majority of the holes give similar results to the Walsingham well.

VI-2-5 Region 5 Northwest Section

A total of five wells were logged in this region. Three of them were suspended wells in Becher, Kimball and Bickford Pools, the other two were suspended wells in the active region of waste disposal around Sarnia.

The temperature curves are shown in FIG VI-14 and the drill-hole locations in FIG VI-15. It is immediately apparent that the Sarnia results can be dispensed with, since the temperature gradients down the holes are similar in both holes and are double those in the holes further south. Disposal is generally where the B-salt has been removed which would be at about 488 m. Thus it can be said that temperatures in this immediate area have been highly disturbed and probably are some 5°C greater than the normal equilibrium values. It would be of very great use to make temperature measurements at intervals in this whole region as a means of tracing the pollution as it spreads through the fractures in B-salt and possibly in other horizons. However this was not attempted here. Darton (1920) has recorded some temperatures at Port Huron which may better reflect equilibrium temperatures in the Sarnia region.

Discussing the other holes from north to south, the first is a hole in the region of the Kimball field. Like other fields in this area the Kimball-Colinville field is a dolomite pinnacle reef in the Guelph and Salina A. It occupies 672 hectares in area in the centre of Moore township, and the main reef is approximately 73 m in vertical extent at a depth of 610 m. Originally discovered in 1947 the field had an initial well-head pressure of 910 psi and since that time it has produced 91,000 barrels of oil and 34,300 MMcf as of the end of 1964. Since the log was made, the field has been converted to gas storage. The hole logged here lies almost 3.2 km south-east of the main reef and about 1.6 km north of the fault zone in the Guelph formation. It had been a small gas producer. At the time of logging, it had less than 40 psi well-head pressure. Thirteen km to the south and slightly to the west lies the Bickford Pool which, like Kimball, is a pinnacle reef in the Guelph and Salina A-1 and again is dolomite. It is somewhat smaller, occupying only 310 hectares in the north-west corner of Sombra township, but has a vertical extent of 110 m. Originally discovered in 1950 with an initial pressure of 988 psi, it had produced some 1160 barrels of oil and 15,115 MMcf of gas at the end of 1964. The hole logged is 752 m south of the main reef and was a suspended gas show. Since the measurements, this field has likewise been converted to gas storage. Sixteen kilometers east and south of the Bickford field lies the East Becher field. This is the smallest of the three, only 65 hectares, and although production is from the Guelph dolomites the structure is a small anticline. The field was discovered in 1946, had an initial pressure of 921 psi and by the end of 1964 had produced no oil and 799 MMcf of gas. Once again the suspended hole logged lay off the main producing structure, 3.2 km to the east, and had been shut-in for some years.

The temperature curves shown in FIG VI-14 suggest that temperatures above 305 m are highly unreliable. These measurements were made in the air column so this is perhaps not surprising. Each hole was cased into the Guelph and the water levels in the holes reflected the hydraulic head in that formation. Surface intercept temperatures are a little over 12°C so that the probable real ground temperature will be lower, since the shallower formations have a higher thermal conductivity, as is shown in Chapter VII. It is apparent in the lower section that the temperatures between the holes differ by only 1°C with the lowest temperatures at the most northerly field. An attractive speculation would be that the temperature differences reflected deep structure but a possible reason lies in the gas production of the nearby fields. As gas is produced it cools the surrounding reservoir and the lowest temperatures here are in the region of the field with the greatest production, ie. the Kimball field. Probably more holes will need to be logged before definite conclusions can be drawn. The temperature gradient in the stable sections are very similar: 9.2°C/km in Kimball, 7.5 in Bickford and 9.8 in Becher. The relation between these gradients and the relative components of the Salina section will be dealt with in Chapter VIII.

VI-2-6 Region 6 South-West Peninsula

Two holes were logged and two industrial logs were available from suspended holes in this region which borders the region of the Wayne County borehole of Region 8. The new measurements were in a long-suspended hole in the Colchester oil-field and in a suspended well in Gosfield North township in Essex County. When first drilled in 1957 the Gosfield hole had oil and gas shows and was shut in at that time pending future development. At the time of logging in 1967 a well-head pressure of 50 psi existed which was released several hours before temperature logs were made. A second hole was suspended under similar circumstances a distance of 30 m away. Unfortunately the latter was blocked at a shallow depth and thus no comparison could be made. No information is available in the literature on the region excepting that oil shows occurred at the base of the Salina and gas shows in the Cobourg formation of the Trenton Group. The old gas field at Kingsville occurs some 19 km south of these boreholes and it initially produced some oil but as it was developed became a very important Guelph gas producing field until flooded. The borehole in the Colchester oil-field was completed early in 1959 and was the wildcat that discovered the Colchester Trenton oil field. After the tests were completed the hole was suspended pending future development work. Temperature measurements were run in late 1963 when the borehole had only a few pounds gas pressure at the surface. To judge by the problems of sinking a probe for the measurements the hole was filled with a very heavy grade of crude oil. Production from the field has steadily decreased from 1961 partially because of recovery problems with the thick heavy oil; by 1965 only five wells were active, producing 28,252 barrels - only a quarter of the 1961 total. Most of the production in the field is from the Cobourg and Sherman Falls formations usually from dolomitised zones within the formation. Sanford (1961) describes the structure of the field as a north-plunging structure with a corresponding synclinal structure on the west side. The dolomitised zones begin in the Black River then transgress up section and up dip to the more porous Trenton formation where they are more intense. The nature of this secondary dolomitization suggests intensive local faulting. Sanford has constructed a cross-section of the field from west to east which is shown together with the position of the logged borehole IHS-1 in FIG VI-16.

The Malden pool is a potential gas producer from the Salina A-1. Discovered in 1946 it has been shut-in ever since, although 14 potential wells have been completed, and never used. The producing reservoir is at a depth of 255 m in a dolomite reef structure with a porosity of 10%.

A comparison of the borehole temperatures between the North Gosfield and the Colchester S. holes shown in FIG VI-17 shows very large differences in temperature gradients although temperatures are similar. The latter hole being only .8 km from the shoreline of Lake Erie certainly contributes to a disturbed gradient as discussed in section VI-6. However it would be useful to have further data from the region as a comparison. Several temperature logs made in suspended boreholes were available from Imperial Oil Ltd. and these are also shown in FIG VI-17 together with the measured temperatures. The log in the #25 hole is less reliable than that in the #75 hole as the former was an up-hole log. However the general character of the temperatures and gradients is similar to those of the Gosfield hole. Surface intercept temperature was about 14°C at Gosfield and extrapolating the curve for Colchester would give a similar result. However the least squares intercept temperatures are considerably lower; 9.3°C for Colchester and 10.6°C for Gosfield with corresponding least squares temperature gradients of 13.8°C/km and 10.8°C/km. The mean annual surface temperature at Harrow between 1960 and 1968 was 10.3°C which gives reasonable agreement (Aston, 1969). Formations penetrated by the wells are between 12 and 33 m of drift lying on weathered Detroit River in all except the Gosfield hole in which Dundee forms bedrock. Below this is the usual sequence of Palaeozoic beds with the boreholes penetrating to the Trenton and to the Cambrian in the case of one of the Malden holes. Temperature gradients in the Trenton in Malden #75 give 20.0°C/km which is high in comparison with the other boreholes.

VI-2-7 Region 7 South-East Michigan

Two holes were logged in this region which directly bounds that of southern Ontario. A set of temperature results was also available for the short hole logged by Leney (1956) at the Wayne County airport near Detroit. The holes logged for this work were rather deeper and lay about 48 km west of Detroit in the Northville field in Washenaw County and 64 km north at the Muttonville gas field in Macomb County. At the time of logging neither of the fields was producing since the former was awaiting conversion to a storage field and the latter to go into production as a Salina gas field.

The Northville field originally began production in 1949 as a Niagaran (Guelph) gas producer with a bottom hole pressure of 1470 psi. To the end of production in 1964, the field had produced 3,410,000 Mcf. In 1954 several holes were drilled to the Trenton formation and to the end of 1964 gas production was 10,310,000 Mcf. Both zones were converted to casing-head status by the development of oil production between 1961 and 1963. Hole #106, the one in which temperature logs were made, was originally completed to 396 m in 1954. It was extended to 1341 m in 1964 to penetrate the Trenton formation and serves as an observation well on that zone. Because of the lightweight cable used the temperature logs could only be completed to a depth of 1936 m. At the time of logging, the hole had never produced and had not been opened for two years, during which time it had accumulated only a few pounds well-head pressure. The topography in the area is very flat with surface cover varying from open fields to scrubby woodland, this particular site being in the woodland. However it is an area with a great deal of building, both residential and industrial being completed.

On the initial planning the hole to be logged in the St. Clair area north of Detroit was one in the Belle River Mills Storage area but gas pressure had broken through into this hole and it could not be blown down. Instead a hole was used in the Muttonville field which had only been completed for six months. This hole was suspended pending fracture into the shut-in gas field. There was no pressure on the well. Although this field is at present shut-in, the discovery well drilled several hundred metres away did break down at the time of drilling and caught fire. However this escape was controlled within a few days and is thus unlikely to have caused much temperature disturbance to the field excepting in the vicinity of that hole. The effect at depth would be a cooling one since the behaviour would be a Joule expansion. Topography is very flat, with a surface cover of coarse grass and a wooded area to the north.

Drift thicknesses vary from 64 m in the Muttonville hole to 128 m at Northville; however below this the former contains shales of the Sudbury and Bedford formations, both of which are missing at the latter. The Antrim shales form the bedrock in the Northville hole and only the lower 12 m are present of the formation. However the Muttonville hole contains 56 m of it. Below this the geological descriptions are similar with the limestones and shales of the Traverse, the limestones of the Dundee and the dolomites and limestones of the Detroit River. The Sylvania sandstone is present only at the Northville, and the Bois Blanc and Bass Island only at the Muttonville hole. Both contain much salt in the Salina below depths of about 563 m of thickness 152 to 183 m. The Guelph formation occurs at a depth of 878 m at Muttonville compared with 1044 m at Northville.

Structurally the two regions are very different since the Northville field lies on an anticline which shows structural closure on all of the formations, Ordovician upwards and is probably reflected in the Precambrian as well. The Muttonville field is on a small Silurian reef typical of the St. Clair region and indeed of the north western region of southern Ontario. These reefs are typically one hectare in extent and as much as 90 m high. Kidoo (1962) and Jodry (1969) have written about the reefs of St. Clair and Macomb counties and list of the order of 25 so far discovered.

Since the temperature gradients in the two holes are substantially different, the logs are best dealt with one at a time. However they are shown together with Leney's results in FIG VI-18.

The Northville hole shows a near-surface inversion in the temperature log with a minimum of between 76 and 91 m. Below a depth of 244 m where this effect has disappeared and the shales of the Antrim formation passed, the temperature log is quite uniform with only small fluctuations of the gradient varying between 9.8°C/km in the Sylvanian sandstone, and in some dolomitic sections of the Salina, to 13.1°C/km in the sections of the Detroit River and the shalier portions of the Salina formation.

At the Muttonville site rather more of the Devonian shales are present and the temperature curve is quite steep down to 335 m and the top of the Detroit River. Close to the surface the temperature log shows a curvature which would probably be turned into an inversion in a region of lower undisturbed gradients. In the upper part of the Detroit River there is a region between 335 and 366 m of almost constant temperature probably corresponding to a zone of water movement. Below this level the gradients vary between 6.6 and 9.8°C/km through the Silurian formations with again a corresponding lowering of the gradient with increased proportions of dolomite and anhydrite in the rocks.

Comparing the two logs with each other and with the shallow log from Detroit, the temperatures are highest in the Muttonville hole by a maximum of 2.7°C although due to the higher temperature gradients at depth in the Northville hole, the temperatures appear to converge. A projection of the Muttonville log would suggest that the well temperatures may be similar at about 1006 m. Projecting the Detroit hole together with measurements made in a salt mine nearby at a depth of 427 m suggests a similar temperature gradient near Detroit to that some miles north at Muttonville in similar formations. However the temperatures differ by 3^{o}C probably partially due to lack of any Devonian shale formations in the Detroit section.

VI-2-8 <u>Region 8</u> <u>Southwest Michigan</u>

Three holes were logged in the Overisel gas storage field a little east of the village of Overisel, about 13 kilometres southeast of the city of Holland and about 19 kilometres from the shore of Lake Michigan. Each of the holes is an observation well lying at the edge of the main structure; each was drilled in 1958 as part of the programme of defining a gas field producing from the Salina A-2 carbonates and marked as a dry hole. The field was discovered in 1956 and went into production in 1957, producing 14,650,000 Mcf of natural gas out of originally estimated reserves of 67,000,000Mcf by 1960, when it was removed from production and converted to storage. It began operation as such in March 1960. Original field pressure was 1364 psi, which had fallen to 979 psi by the time of conversion. Storage has been slowly built up to 28.200.000 Mcf with a slight overpressure, and annual injection and withdrawal had been stabilized by 1964 as 12,000,000 Mcf.

Earliest production from the large palmate-shaped structured field extending over some five square kilmeters was in 1938 from the Traverse limestone at a depth of 533 m. Petroleum was recovered from the crystalline dolomite at the top of the Traverse formation. with a 3 m pay zone. The Salina gas zone discovered in 1956 averaged 15 to 18 m at a depth of 808m. In this section the reservoir is a light tan to buff, finely crystalline dolomite interbedded with anhydrite. Porosity in the pay section is intracrystalline with very little solution porosity and ranges from 2.6 to 22.2%. Permeability of the dolomite ranges from 0,1 md to 29.0 md with permeable lenses separated by lower permeability beds or by anhydrite. Interstitial water content is high, averaging around 29%.

Most of the oil and gas fields of southwest Michigan occur along elongate structural trends, most of which seem to be salt-cored anticlines over lenses of 30 - 60 m of Salina A-1 salt. and the Overisel field is no exception with a lens of 46m of salt on the Niagaran. The effect of this structure persists at least as near to the surface as the Coldwater red-rock and Traverse highs are right over the A-2 high. Although few wells have been drilled through to the Niagaran, Ells (1967) believes the surface to be essentially flat. How these structures originated is a matter for conjecture. One theory suggests differential removal of salt by solution and indeed one core description from Section 27 shows solution brecciation characteristics. Solution brecciation characteristics have been described in many wells in southern Ontario and Michigan (Landes, 1948, Sanford, 1965). However "salt-pillars" have been described by Trusheim (1960) as migration phenomenon.

FIG VI-19 shows the temperature logs and positions of the boreholes with respect to the main storage area, and FIG VI-20 shows the wells with respect to cross-sections of the main structure.

Beneath 152m of glacial drift lies the Coldwater formation consisting of light or bluish-grey shales with beds of calcareous sandstone and dolomitic shale. Much of the drift appears to be derived from weathering of this formation and thus the top can be very hard to pick. In the Overisel region the Marshall formation appears to have been eroded away, although it occurs to the east and north. It would seem that Overisel with morainic hills around it lies on an old post-glacial lake bed. Riggs (1938) states that the first continuous bed in Allegan County is the Coldwater limestone, a rather cherty shale. In the wells logged here, the depth of this horizon varies from 103 m in 0-157 to 204 m in 0-162, and 180 m in 150. 0-157 is also at a higher elevation than the other two and sits on a moraine side, suggesting that the moraine deposition may have been structurally controlled. However, the Coldwater red, marking the base of the formation and consisting of red argillaceous limestone or shale, occurs at close to 274 m in all three wells. Below this the wells penetrate 107-167 m of greenish-grey shale of the Ellsworth formation, beneath which are the similar shales of the Antrim, succeeded by the shaly limestone and dolomites of the Traverse formation. Below this the limestone of the Dundee is not always found in the subsurface. The transition to the dolomites of the Detroit River is very abrupt. Not all of the subdivisions of the Detroit River are present in Southwestern Michigan, and the zones of salt generally found in the central basin are absent. The highest subdivision of the series is that of the Lucas formation which consists of 91 m of dolomite and anhydrite. Below the Lucas, the Amherstberg of the more eastern and northern regions is absent over much of southwest Michigan. Although the subdivisions of the Detroit River are not identified in the core descriptions, it would seem from Landes (1951) that the Amherstberg pinches out to the east of Overisel, as does the Bois Blanc, leaving the Lucas dolomite sitting directly on the dolomites of the Bass Island. At the base of the Devonian rocks there is usually an inconformity, where further south the Devonian is deposited directly on the Niagaran beds. In the Overisel region the logs do not generally subdivide the Salina above the A2 carbonite; however, the descriptions indicate a succession of limestones, shales and dolomites with large amounts of anhydrite present. The A2 consists, in the upper part, of dolomite with limestone beneath. In the Overisel Field this unit is fractured and then infilled with anhydrite. The gas pay occurs within this zone. Below this zone the A1 salt thickens beneath the structural high. Few holes have penetrated very much deeper than this, and all that may be said of the Niagaran is that it is a brown dolomite of undisclosed thickness. The temperature logs of the boreholes are shown in FIG VI-19.

Below 180 m, borehole #162 is warmer than #150 by 0.6°C and warmer than # 157 by 1.8°C. The average gradients above 488 m are 20.7°C/km and below 488 m, 8.9°C/km. This break corresponds to the depth of the top of the Traverse formation which is the first limestone formation in the series.

In more specific terms, only #162 shows the surface inversion. All three wells show the same temperatures at 91 and 122 m, probably due to water movement in a sandy zone that covers most of the area. At 152 m the major difference in temperature in each of the wells becomes finalized and is constant from that depth down, with local fluctuations especially in the shallower formations. Well #175 shows a zone of constant temperatures at 335 and 366 m, which may be due to water flows.

VI-2-9 Region 9 Central and Northwest Michigan

In the central and western regions of Michigan, comprising approximately the Counties of Lake, Osceola, Mecosta and Newago, temperature logs were made in six holes. Five of these holes were observation holes around so-called 'shoestring' gas fields, and the sixth was a much deeper hole extending into the Detroit River Formation, and is used as an observation well on the Loreed field. For convenience of discussion these holes are divided into two groupings, namely, the three Marion holes of the Consumer Power Company and the Big Rapids-Reed City holes of Michigan Consolidated Gas Company. Water levels were checked in the Marion holes, and the Austin hole of the latter grouping, at monthly intervals.

Central Michigan varies quite widely in surface characteristics but generally tends to be heavily forested and fairly sandy in the northeast with the proportion of farmland increasing to the south and west. Topographic variation is generally not large and the land is generally about 305-366 m above sea level.

At this stage it is appropriate to divide the holes into the two groupings for further description.

Marion Area

The holes selected for logging are observation wells on the fringes of three gas storage fields, Cranberry Lake, Winterfield and Riverside. All three fields were originally put into production in 1941 as gas producers from the Michigan Stray Formation. The original field pressure in each case was about 540 psi, which fell to final pressures of between 146 and 417 for the various fields before production ended. In 1947 the

Cranberry Lake and Winterfield fields were converted to storage at a pressure about 10% over the original field pressure. The Riverside field followed in 1952, making a total storage capacity in the three fields of 40 Mcf. As in the case of the Overisel well discussed in the previous section, the chosen wells were ones showing little or no annual disturbance to the water column, ie., not connected directly to the reservoir and as far away from the reservoir as was physically possible. In this case the wells were rather farther from the reservoir than in the case of Overisel, which partially offsets for the greater length of time used as storage fields. The surface features are somewhat varied since the Cranberry Lake hole is in a large wheat field on very flat farm land, the Riverside hole is in a small clearing in deciduous woodland and the Winterfield hole is on the edge of an open, very gently, grassy slope.

Comparison of the geological series in each of the three holes reveals the variation of drift thickness to be fairly small, only about 20 m between the holes, which makes the surface of the bedrock very flat when the elevation differences are taken into account. The geological sequence consists of the sandstones and siltstones of the Saginaw Formation below which is the 'Triple Gypsum' of the Michigan Formation composed of alternate sequences of anhydrite and shale. Both of the above formations thicken considerably to the south and west away from the central basin. The 'Brown Lime' section of the Michigan consists of dolomites and shales with dolomite stringers. Beneath this is the Michigan Stray sandstone which is the pay and storage zone for the 'shoestring' gas fields. Beneath the Stray and at the base of the Michigan Formation is a sequence of dolomites, sandy shales and shaly limestones. The Marshall Formation below, at depths varying from 457 to 518 m beneath the surface, consists of a grey medium-grained sandstone similar to the Michigan Stray sandstone.

Temperatures are shown in FIG VI-21, and VI-23 showsthe approximate location of the fields. As can be seen, the Winterfield hole shows the highest temperatures, followed fairly closely by the Riverside hole with only 0.2^{o}C difference in the lower sections. whereas the Cranberry Lake hole is considerably cooler, ie., about 1.5^{o}C over the same sections. In each of the holes temperatures converge above 152 m where the boreholes penetrate glacial tills rather than bedrock. The explanation of this must be sought either in rapid deposition of the surface layer and insufficient time for it to achieve equilibrium, or in common water movement in the area. These possibilities are discussed further in Section VI-4, so, for the moment, the problem is posed of similar temperatures in the top 152 m and an inversion at 122 m in borehole in which the first 183 m is through unconsolidated deposits. The least squares surface intercept temperatures are very similar at 7.8^{o}C. Each of the holes penetrates similar geological horizons, namely the Saginaw and Michigan Formations. These consist of fairly complex interbedding of sandstones, shales, dolomites and anhydrites. The temperature gradients in these formations are between 17.3 and 19.7^{o}C/km in the Saginaw, and in the Michigan vary between 23.0 and 26.2^{o}C/km.

Big Rapids - Reed City Area

As in the case of the Overisel and Marion wells, the three holes in which temperature measurements were made are salt-water observation wells close to three gas storage fields. The Austin and Billingsly holes are relatively shallow since the storage horizon is the Michigan Stray, similar to the Marion fields. The former field was discovered in 1934 and converted to storage in 1941. By 1964 it had become a major field of the Michigan Consolidated Gas Company and was over-pressured by approximately 50%, making a total storage capacity of 13Mcf. Rather a different case is presented by the Billingsly field which is at the edge of a small sub-field discovered in 1947 and linked into the Goodwell field several years ago for storage purposes. The third of the holes is a rather longer one drilled to the Reed City zone of the Detroit River and has been converted from a dry hole to an observation well on the Loreed gas storage field. This field was the Reed City Monroe oil field and was converted to gas storage in 1963. It is a fairly small field at present with only 5 Mcf capacity.

The region is covered by a fairly uniform thickness of drift, about 183 m, and is uniform in topography with all three holes being collared between 305 and 335 m above sea level. Both the Austin and Billingsly holes contain red shales forming the bedrock, which are probably Permo-Carboniferous Kimmeridgian 'red-beds'. Beneath these, and forming the bedrock in the Bregg well area are the shales of the Saginaw Formation. 'Parma' sandstone occurs only in the Billingsly well in which the Saginaw is very much thinner than in the other two wells. The Michigan Formation consists of 90 m of limestones, shales with some sandstone sections and anhydrite in some sections. In both the Austin and Billingsly holes the sandstones of the Stray Formation occur but this is represented in the Bregg well by only a sandy limestone. While the Billingsly hole bottoms in the

Stray and the Austin in the white sandstones of the Marshall Formation, the Bregg hole continues through the bluish shales of the Coldwater and the red limestones at its base with a total thickness of 183 m, through 61 m of interbedded limestone and shale of the Sudbury Formation, 183 m of grey shales of the Ellsworth and Antrim. The Traverse Formation at a depth of 896 m consists of dark coloured limestones.

The temperatures for each of these three wells are shown in FIG VI-22. Again the curves show a temperature inversion close to the surface, but, as is not the case in the Marion holes, the curves remain offset by about the same amount as they do in the deeper portions of the holes. These holes are much longer distances apart and thus perhaps it is not surprising that the inversion curves are displaced. The offset in the deeper sections of the holes amounts to 0.8^{o}C between the respective curves with an apparently increasing general temperature towards the east and south, a pattern similar to that evidenced in the Marion holes. Both the Billingsly and Austin fields have similar least squares intercept temperatures; however, that of the Bregg well is considerably lower even when only the sections to the same formation as the other two are considered. Whereas the Austin hole is on rolling, sandy, open heathland and the Billingsly is on fairly flat, open pasture with some forested lots, the Bregg hole is on the north side of gently rising land beside a small creek, and this difference of environment might explain the differences in temperature.

The temperature gradients in the Saginaw are quite varied, ranging from 19.7 to 24.6^{o}C/km, reflecting compositional differences. Similar variations occur in the Michigan Formation with the overall mean lower than in the Saginaw. In the Bregg hole the gradients are relatively high through the shales of the Coldwater, Ellsworth and Antrim Formations with averages of 36.1°C/km, and the gradient falls off to below 19.7°C/km in the Traverse limestones.

VI-3 REGIONAL TEMPERATURES AND TEMPERATURE GRADIENTS

It is of particular use to the oil and gas industries to have a reasonable idea of what temperatures might be expected during the drilling of a borehole and also what reservoir temperatures might be expected, since this has a great effect on the viscosity of oils and on the formation of hydrates both of which affect production rates considerably.

With some large fields in southern Ontario being converted to storage reservoirs these factors become even more important. For this reason a short section is included here which is a discussion of temperatures at depth across the entire region covered in these surveys.

A Temperatures

Temperatures at depths of 151, 305 and 610 m below ground level across the region are shown in FIG VI-24 which also shows a temperature profile and geological map across the line AB. As can be seen in the diagram, the temperature pattern at different depths shows a considerably different distribution. This distribution changes as formations of a different thermal conductivity succeed each other in the geological sequence. Very few deep temperature observations exist for the region. Probably the deepest hole in this area was Brazos-Sun-Superior State Foster #1 drilled at Rose City in Ogemaw County in 1964 where bottom-hole temperatures were made 6-8 hours after circulation ceased. These temperatures were 53°C at 2692m in the Guelph and 64^{o}C at 3743 m, 82^{o}C at 3958 m in the Cambrian Mt. Simon formation. This hole did not reach Precambrian basement. Temperatures are obviously very disturbed but in such a way that they will be slightly lower than the true formation temperatures. Thus we do have some lower boundary temperatures for the base of the Michigan Basin. Several long boreholes were logged in the course of this work giving the following temperatures:-

23.8^{o}C	at Northville	in the Guelph	1036 m
36.2^{o}C	at Loreed	in the Traverse	1006 m
26.0^{o}C	at Wylie	in the Trenton	1006 m

plus the Morriston industrial log:-

33.2^{o}C	at Morriston	in the Precambrian	1768 m

Comparing these results shows that temperature differences of as much as 15^{o}C exist across the region at any one depth below ground level which must be due to either very great changes in heat flow or to a variable sedimentary sequence, since surface temperatures vary by only a few degrees. The latter is the most likely explanation since the series contains formations ranging from very friable shales of very low thermal conductivity and hence in which the temperature gradients will be very high, to rock salt in which the reverse is true.

B <u>Temperature Gradients</u>

A further way of presenting temperature data for the oil industry is in terms of a linear relationship between bottom-hole or reservoir temperature and mean surface temperature. This has been done by

Van Orstrand (1935) and Darton (1920) in the past and on a regional basis for the U.S. Gulf Coast by Nichols (1947) and is still commonly done today, eg., Alberta Oil & Gas Conservation Board (1967). This method takes no account of either changes in thermal properties of the rocks at intermediate depths or of the presence of near-surface temperature inversions in many boreholes. Probably of far more use are the temperature gradients and variability within each formation since this may then be used to build up a truer picture of the temperature changes to be expected at any depth. These are shown in the Table below. Using these figures plus a geological and a mean annual ground temperature a temperature profile for an area can be constructed which should be a reasonable approximation. The variation of heat flow in the region, which if large may upset these figures, is discussed in Chapter VIII.

Table VI-1

TEMPERATURE GRADIENTS ($^{o}C/km$) IN SOUTHERN ONTARIO AND MICHIGAN

Formation Name	Range	Mean	Stnd Dev.	No. of Holes	Extent (km)
Precambrian	12.8 - 15.4	14.1	1.1	3	400
Cambrian	-	8.5	-	1	-
Trenton-Black River	14.8 - 24.6	18.4	2.6	12	670
Collingwood	30.0 - 45.0	36.7	4.3	5	600
Meaford-Dundas	19.7 - 31.5	26.1	3.6	7	300
Queenston	15.7 - 20.7	18.0	1.9	5	320
Clinton-Cataract	17.0 - 17.7	17.3	-	2	60
Guelph-Lockport	4.9 - 7.5	6.6	1.1	4	220
Salina complete	7.6 - 12.5	9.3	1.4	10	500
Salina upper	7.9 - 13.1	9.0	1.4	12	500
Salina evap.	6.2 - 9.2	7.5	1.2	5	130
Bass Island	5.9 - 12.8	8.9	2.2	10	400
Bois Blanc	8.5 - 10.5	9.7	0.9	3	60
Sylvania	8.1 - 9.9	9.3	0.8	3	60
Detroit River	7.5 - 11.5	8.9	1.6	7	300
Dundee	-	-	-	-	-
Traverse	10.2 - 21.0	14.8	3.9	5	290
Antrim	26.9 - 49.2	34.4	8.1	5	290
Coldwater	20.3 - 38.0	25.9	7.2	4	200
Michigan	18.4 - 28.9	24.4	3.2	6	80
Saginaw	18.7 - 25.6	22.3	3.0	6	80
Red beds	18.7 - 21.7	20.2	1.5	2	40

VI-4 DISTURBANCES TO EQUILIBRIUM TEMPERATURES

In the previous section measurements of the geothermal gradient in boreholes have been discussed and anomalous temperatures have been pointed out. These anomalies are caused in a variety of ways and it is the object of this section to show how they arise and what degree of thermal disturbance they are likely to cause. In a general sense, such disturbances are induced either by the drilling of the borehole itself or are caused by local geological or geographical conditions, and may physically be attributed either to thermal conductivity contrasts, heat sources or sinks, or to variable surface temperatures.

VI-4-1 Disturbances Induced by Drilling

The majority of holes are drilled by a rotary method in which heat is generated by mechanical processes near the drill bit and by stresses in, and frictional rubbing of, the drill rods. However, most of this heat is rapidly transmitted to the circulating drilling fluid. This drilling fluid, which usually is water or a water-mud suspension, is pumped down the drill rods to the bottom of the hole and is returned to the reservoir at the surface via the region between the rods and the wall of the hole, bringing the drill cuttings with it. Since the drilling fluid is generally at a different temperature than the wall rock, heat is conducted between them at rates dependent on their relative thermal properties, temperatures, and the nature of the wall-to-fluid contact. In some instances the wall rock is permeable and heat transfer is accomplished partially by flow of drilling fluid into the wall rock. When this leads to a serious loss of drilling fluid, cement is injected into the horizon, thus producing further thermal effects due to the hydration of the cement. Once the drilling is completed, the drilling fluid, or in many cases the water which has been pumped in to displace it, is left to return to thermal equilibrium with the surroundings. The way in which the temperatures in the borehole return to equilibrium is dependent on the interaction of many factors and is probably almost impossible to predict completely. The effects of drilling and the subsequent return to equilibrium are discussed more fully in the context of the U.W.O. borehole in Chapter X.

The hydration of cement is an exothermic process. Both Heiland (1946) and Gretener (1968) have made detailed measurements of temperature rises in boreholes after cementing operations. The latter observed a maximum temperature rise of 5°C recorded 14 hours after the placing of the cement.

In the sedimentary environment many of the open boreholes have been oil or gas producers and as such may have very anomalous temperatures. Garland and Lennox (1962) found a negligible disturbance to the temperature gradient in a 1000m borehole at a time of 2.4 times the length of production. Anomalous temperatures also occur at depths where gas or liquids containing gas enter a borehole and the gas expands and cools. Various authors have made use of this to determine rates of gas flow.

Even when the drill-hole is completed and has been given sufficient time to return to equilibrium, measurements in the fluid column may not represent those in the surrounding rocks. If the local geothermal gradient exceeds a critical value dependent on the hole size and the physical properties of the fluid filling it, the column will become unstable and convective overturn will occur. Krige's (1939) formulation is most often quoted in determining the onset and various authors have attempted to predict this limiting gradient for various conditions of drill-hole: Misener & Beck (1960) and Sammel (1968) for water columns, and Garland & Lennox (1962) for air, gas and petroleum filled boreholes. Few attempts have been made to predict the actual temperature disturbance which might result (Donaldson 1961). Diment (1967) and Gretener (1967) have observed temperature oscillations of several hundredths of a degree with periods ranging from a few minutes to a few hours which they attribute to convection in boreholes with diameters of 25 cm and gradients of $20^{\circ}C/km$ and more. Measurements in the course of this work were made in boreholes of diameters of 3 to 33 cm and gradients observed varying from 5 to $30^{\circ}C/km$. The only temperature fluctuations observed were in Salina sections and are believed due to water movements since they were more noticeable in uncased boreholes.

One of the worst disturbances caused by a borehole is that it provides a pathway between different porous horizons which may, prior to the hole, have been separated by an impermeable layer. If different fluid pressures occur in the horizons up- or down-hole flows will result. This can often be stopped by grouting while the drill-rig is still on the site. Misener & Beck (1960) show some typical temperature logs in boreholes with water-flows. Birch (1947) and Boldizar (1958) have attempted to correct heat flow results for such disturbances. Very many, otherwise uniform, boreholes show small deviations in the heat flow results which are probably due to such flows and are usually attributed to such, although few attempts at quantitative measurement have been attempted.

VI-4-2 Structural and Topographic Effects

Topographic disturbances to equilibrium heat flows due to elevation differences within a few kilometers of the well-site have been investigated in some detail, as is discussed by Jaeger (1965) and Lachenbruch (1968). No large differences in elevation were encountered in the course of this work and no corrections of this type were made. A significant topographic disturbance is caused by the presence of a large body of water within a few kilmeters of a borehole since it usually has a rather different mean annual bottom temperature than the surrounding surface. Suitable methods of correcting borehole temperatures have been given by Lachenbruch (1957). Several of the temperature logs measured in the course of this work were from boreholes in close proximity to the Great Lakes. Both the Colchester South and the CNE boreholes lie at a distance of ½km from the shorelines of Lakes Erie and Ontario respectively. Temperatures measured in the Colchester borehole show a depression of temperatures at shallow depths, when they are compared with those in boreholes further from the shoreline, and a minimum in the temperature log occurs at 200m. Inspection suggests that since local temperature gradients are very low the effect of the lakes has been sufficient to cause a temperature inversion. However temperature gradients at a depth of 520 to 610 m are consistent with Meaford-Dundas gradients measured elsewhere. In order to deduce a model which would give reasonable formation gradients in the other sections of the borehole, the land-to-water differential temperature would need to be 20°C, and the lakes to have persisted for several million years. But the mean annual ground temperatures at nearby Harrow are 10.3°C and mean annual lake temperatures deduced from measurements by such as Rodgers and Anderson (1963) is no less than $3^{o}C$ giving a differential temperature which is too low. The mean annual lakes temperature is quite variable depending on the severity of the winter and the proportion of water surface with ice-cover but the estimate certainly could not be in error by the required amount. Possible the borehole also possesses a shallow temperature inversion due to recent climatic changes which is confusing the interpretation. Although the temperatures in the Gosfield hole show no inversion, the temperature gradient does decrease towards the surface.

A similar correction to the CNE borehole increases the average temperature gradient by approximately 2%.

Distortions to the geothermal field result from thermal conductivity contrasts in other than flat-lying horizontal beds. In the vicinity of features such as salt domes conductivity contrasts as high as 4 to 1 may result if the salt has pushed its way through a mass of shales. If a borehole had penetrated the salt and the salt is assumed to be a flat bed extending to infinity in all directions, a heat flow calculated in that horizon may be too high by 50% if in fact the salt horizon was a dome structure and had been recognized as such and the gradients corrected accordingly. The sediments of southern Ontario contain no salt domes but they do contain both isolated salt bodies remaining after the surrounding salt has been leached away and salt sequences of varying thicknesses. However the sequences surrounding them, both above and below, are usually dolomites, dolomitic limestones, dolomitic or anhydritic shales, all with fairly high mean conductivities. Thus the conductivities differ by about 40%, and the heat flow variation is only 10%.

Reef structures, however, are very common in the sedimentary sequence of southern Ontario and parts of Michigan. These occur particularly in the Guelph and Lower Salina. Many of the boreholes drilled by oil and gas companies are either on or close to such features since they are commonly producers of oil and gas. The rocks immediately below and surrounding these features are usually of the same composition as biohermal reefs (Pounder, 1963). However Sandford (1964) has suggested that a sharp change from limestone to dolomite occurs on approaching pinnacle reefs. Jodry (1969) has pointed out that the dolomitised zone and the mixture of limestones and dolomites is rather larger than the original reef area. The reef may be up to 100 m high and 1½km in extent. The core zone is often very porous but the pores may be filled with a range of materials from gas to salt. Using reasonable estimates for conductivities the heat flow within the reef zone may be 10% greater or less than that in the surrounding rocks; careful measurements on cored holes would detect this condition.

That groundwater movement may be a major disturber of the normal geothermal field has long been suspected. In many early collections of papers dealing with earth temperatures, water has been used to explain anomalous temperatures. Bullard and Niblett (1951) used such ideas to explain the abnormally high heat flows from Eakring. Bullard's (1939) work showed that heat flows increased sharply on entering the less permeable quartzites. Schneider (1964) has studied well discharge temperatures for depths of up to 600m in a highly permeable dolomite and limestone aquifer in Central Israel. The results showed a large lateral flow which was later verified by injection tests with radioactive tracers. Horizontal velocities of 0.6 to 1 m per day were observed. More recently Judge & Beck (1967) have used small horizontal velocities of 6 cm per day to explain the anomalous heat flow layer in the London hole. This is treated in more detail in XI-9. Using the geochemical approach Van Everdingen (1968) has concluded that major systems of circulation of formation water, driven by the fluid potential field, must exist in the Western Sedimentary Basin of Canada.

In addition to lateral movement, convective overturn may also occur in a sufficiently permeable horizon with a high heat flow. The temperature gradients in any of the formations in southern Ontario are insufficient to cause overturn.

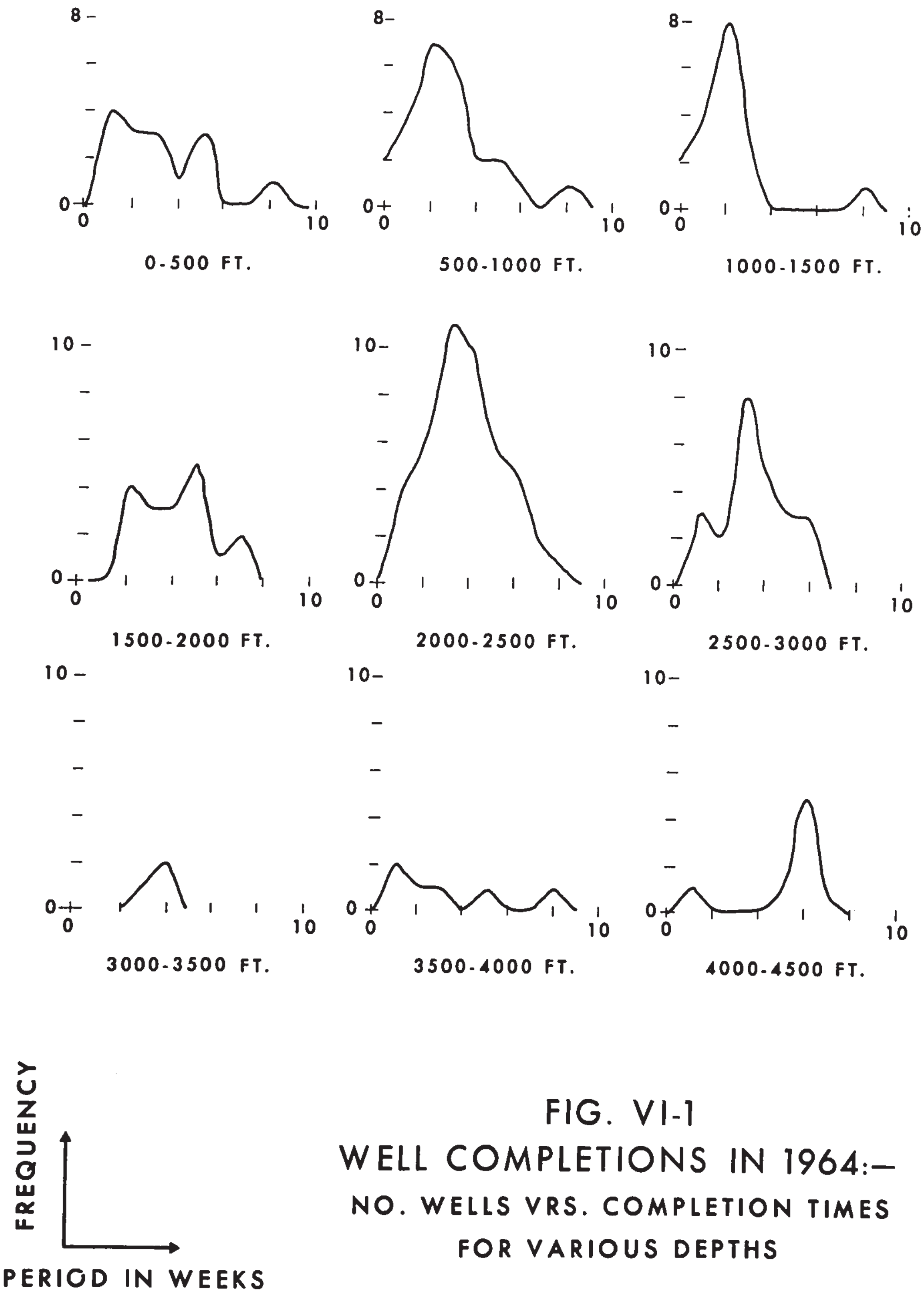
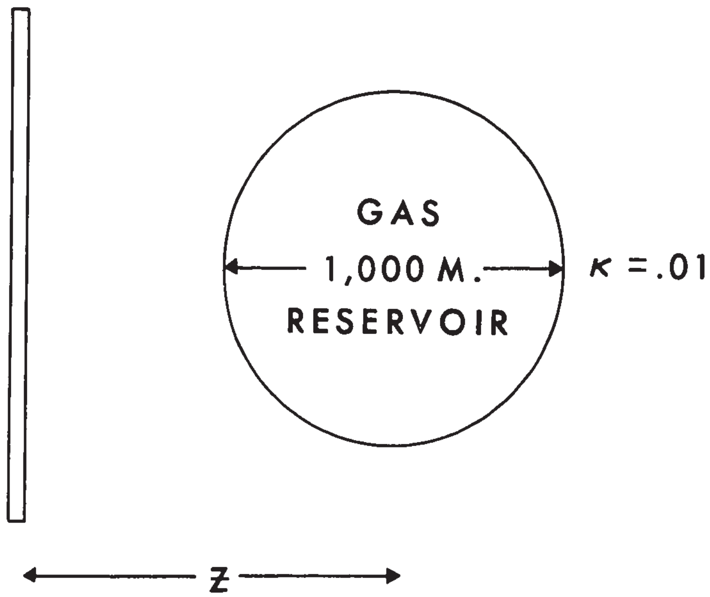


FIG. VI-1
WELL COMPLETIONS IN 1964:—
NO. WELLS VRS. COMPLETION TIMES
FOR VARIOUS DEPTHS



DISTANCE FROM RESERVOIR METERS
750
1000
1500
2000

TIME FOR 1% OF DISTURBANCE TO REACH DRILLHOLE YRS.
200
1000
4000
8000

FIG. VI-2 PROPAGATION OF THERMAL DISTURBANCE FROM A GAS RESERVOIR

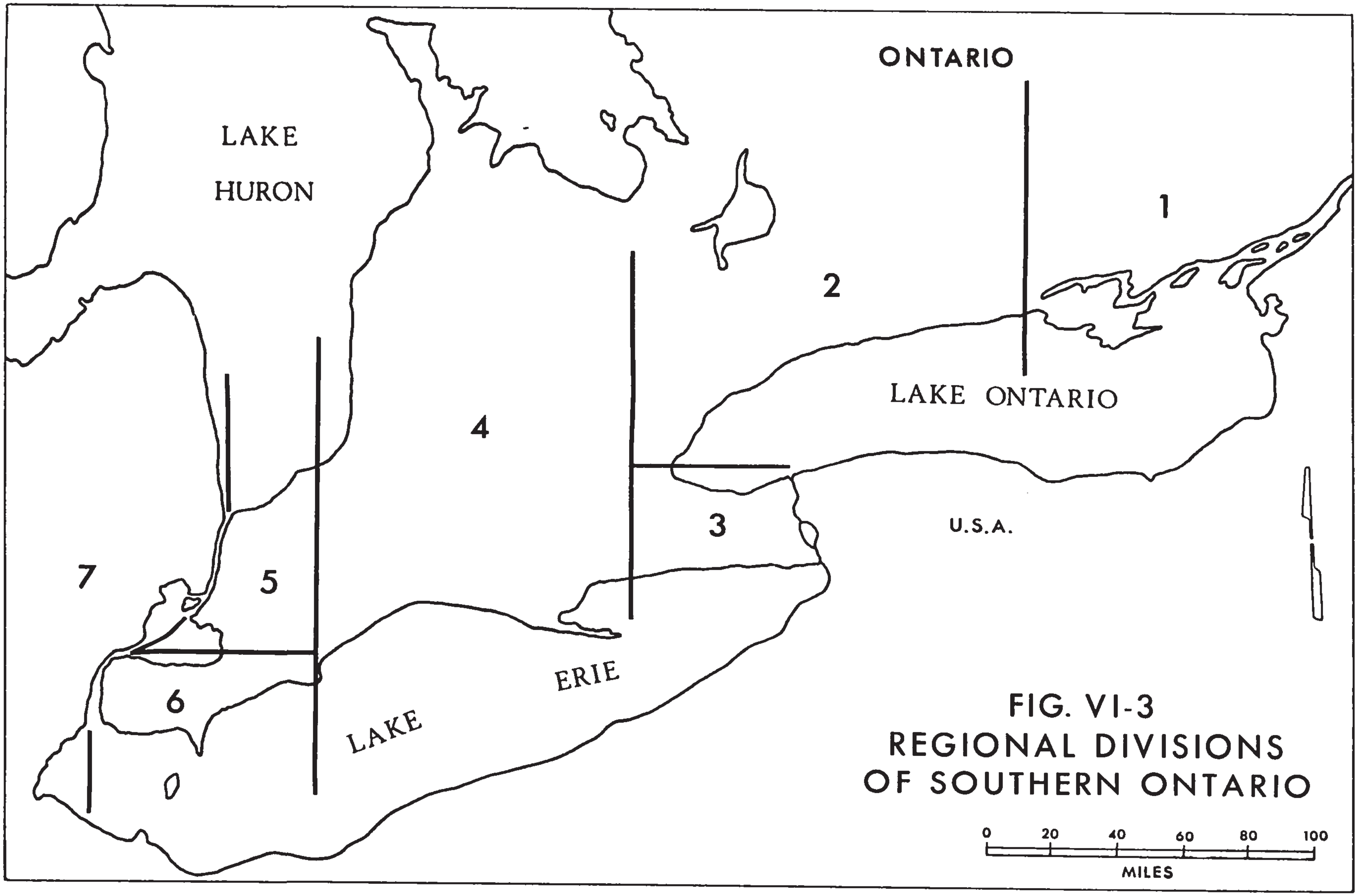


FIG. VI-3
REGIONAL DIVISIONS
OF SOUTHERN ONTARIO

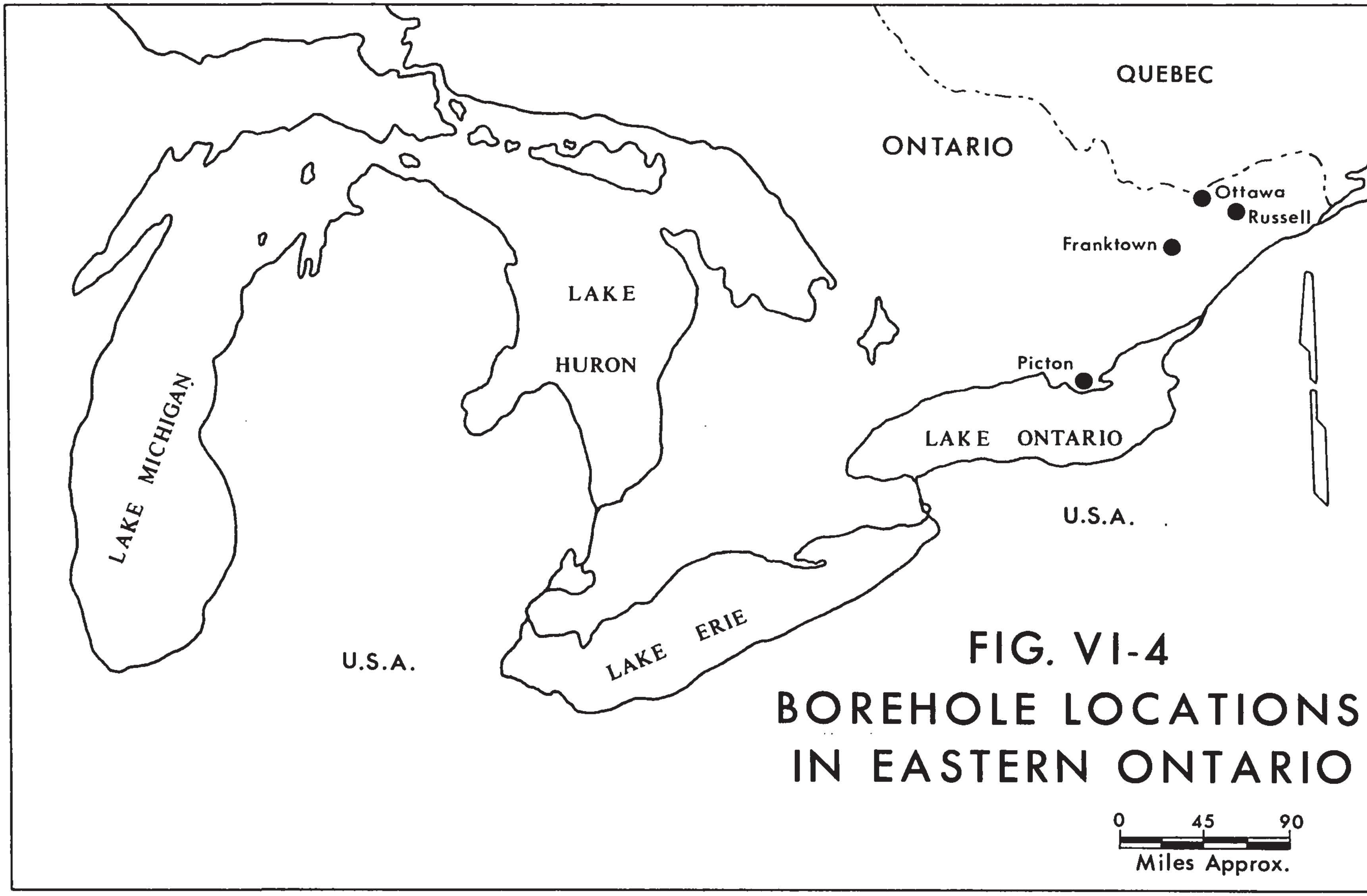


FIG. VI-4
BOREHOLE LOCATIONS IN EASTERN ONTARIO

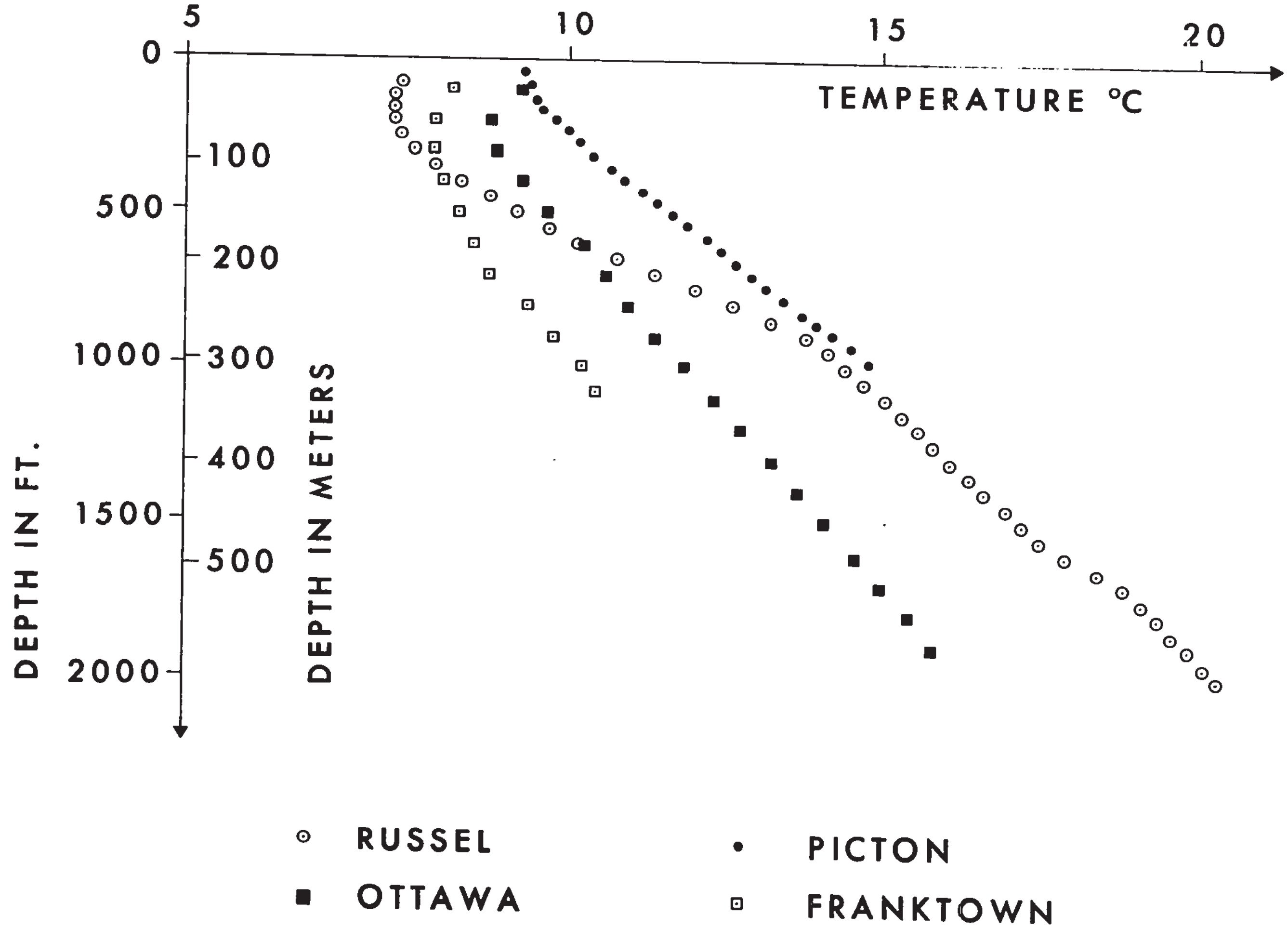


FIG VI-5

BOREHOLE TEMPERATURES IN REGION 1

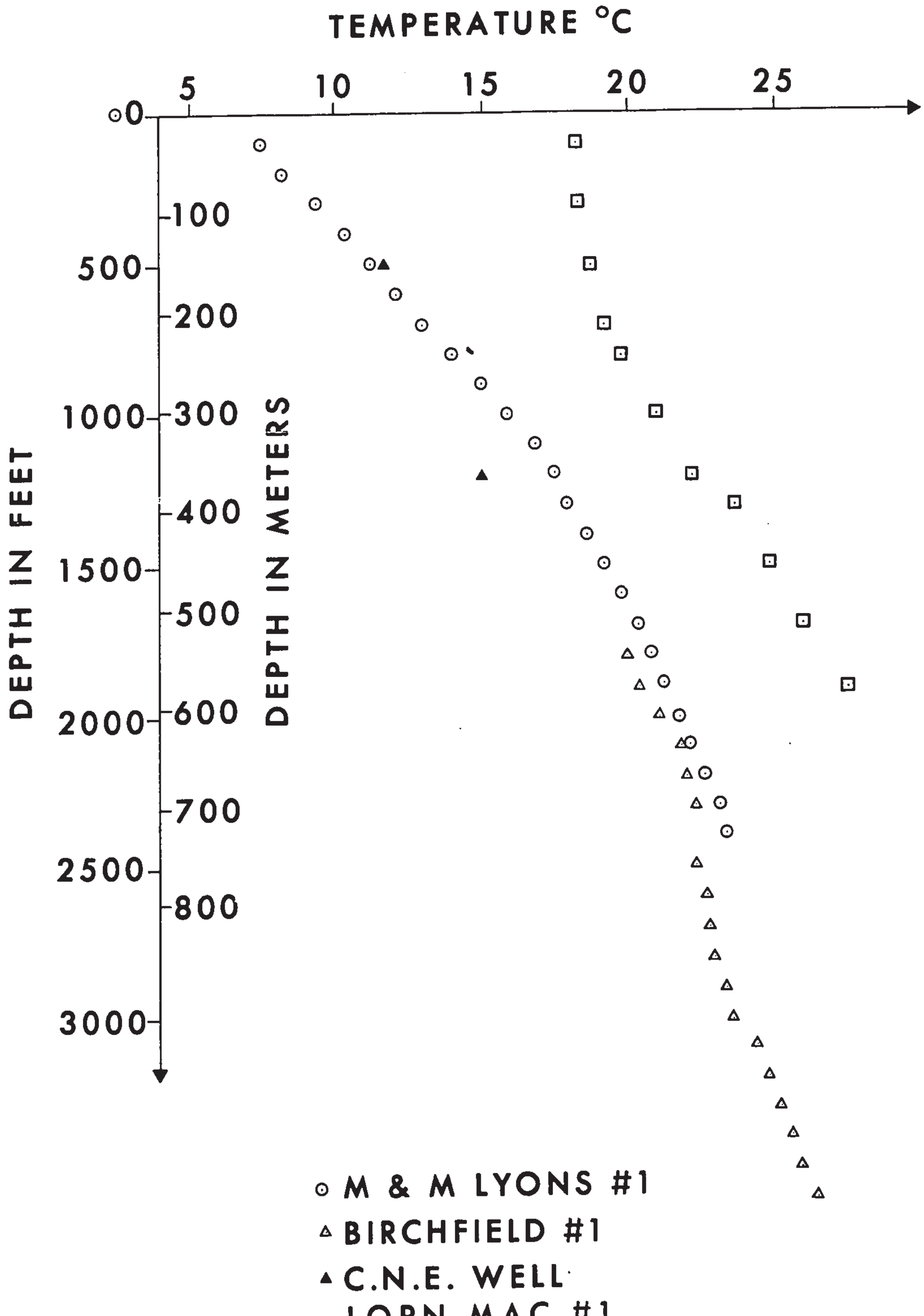


FIG. VI-6

BOREHOLE TEMPERATURES IN REGION 2

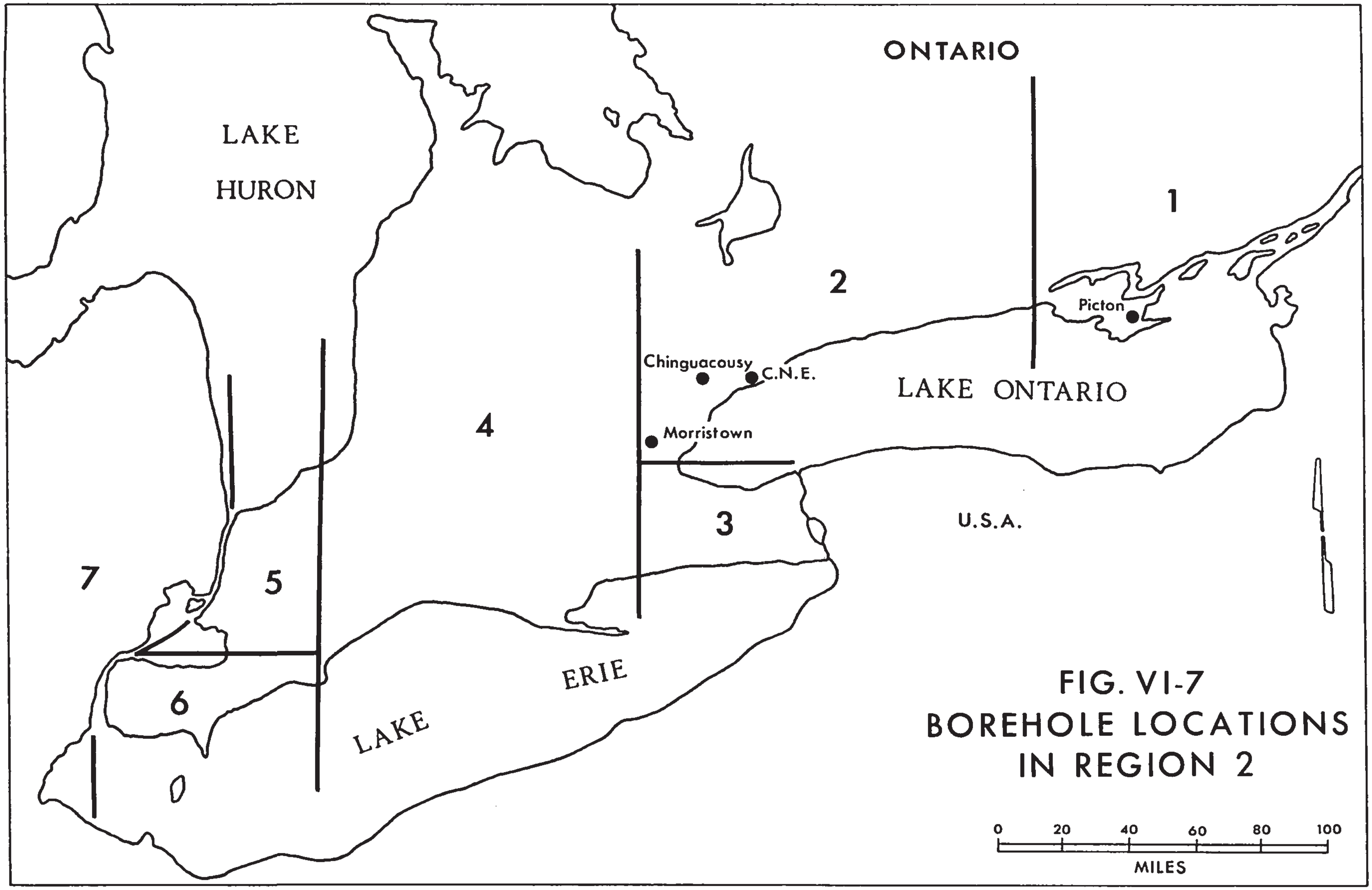


FIG. VI-7
BOREHOLE LOCATIONS
IN REGION 2

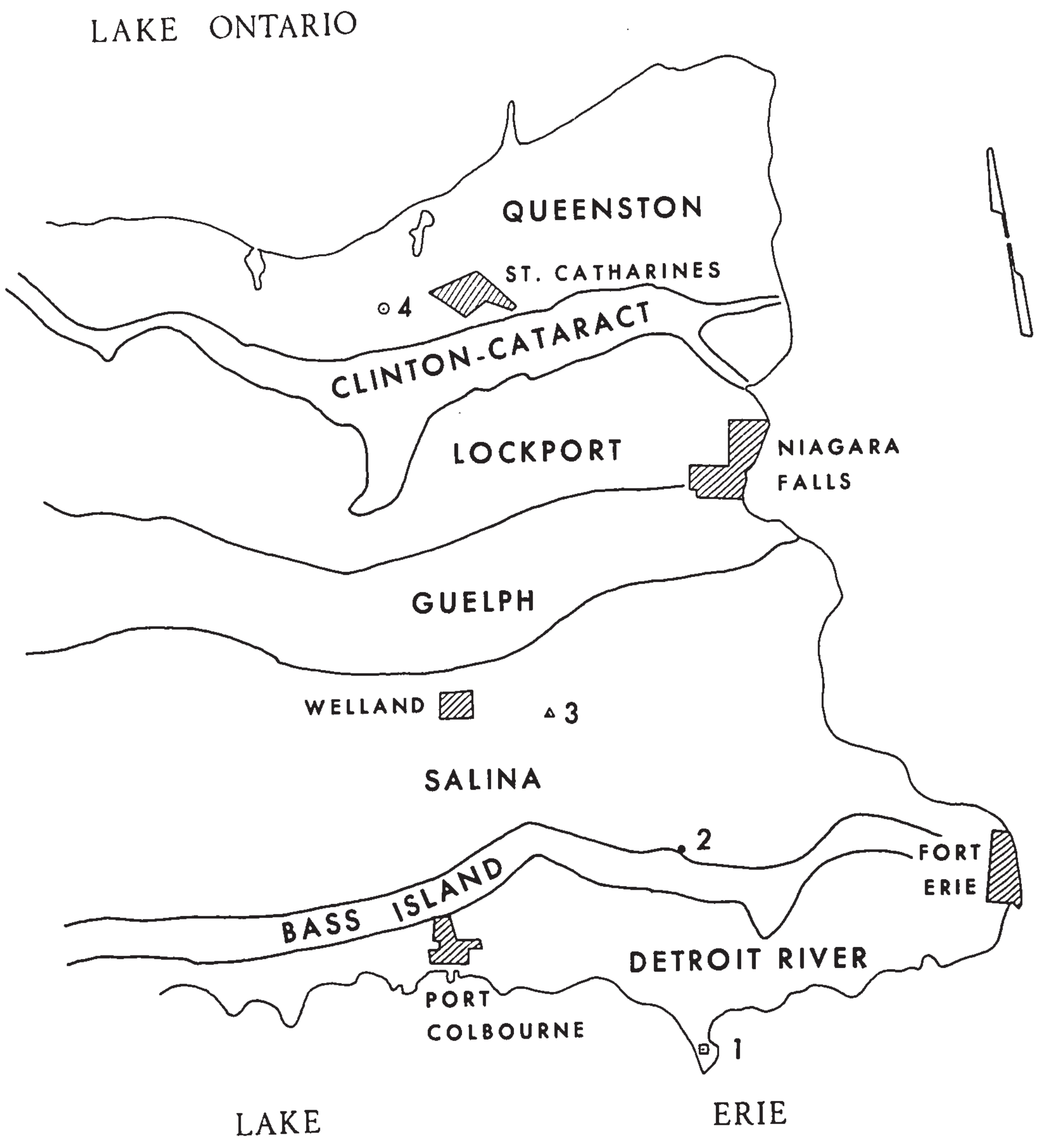
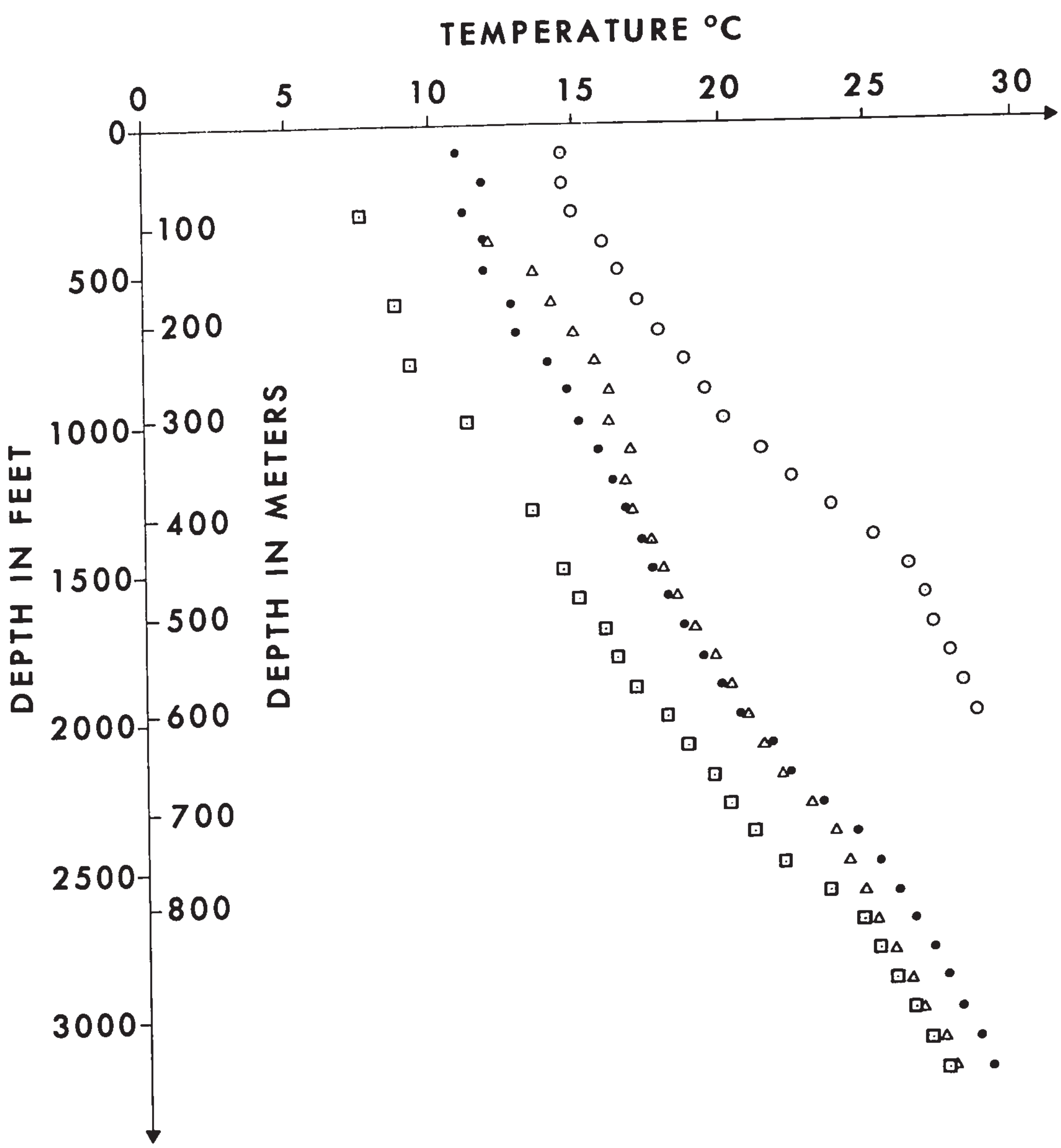


FIG. VI-8
DEEP BOREHOLES IN NIAGARA REGION
1 INCH = 4 MILES APPROX.



1 ▫ BERTIE (32/B.F.)
2 • BERTIE (6/15)
3 ▵ CROWLAND (11/5)
4 ○ LOUTH (4/3)
NO. REFER TO LOCATION
IN FIG. VI-8

FIG. VI-9
BOREHOLE TEMPERATURES IN REGION 3

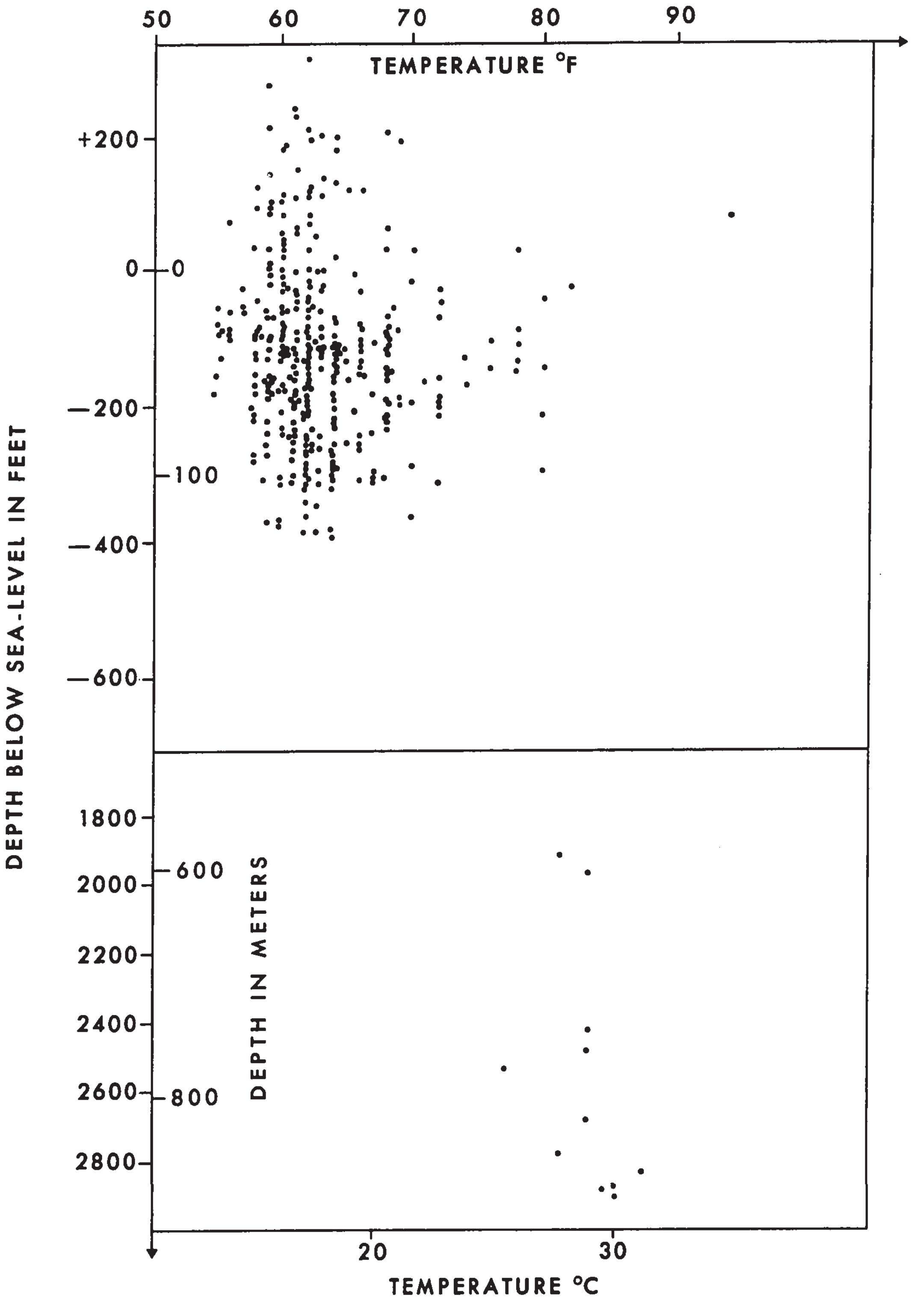


FIG. VI-10
BOTTOM-HOLE TEMPERATURES VRS. DEPTH
IN THE NIAGARA PENINSULA

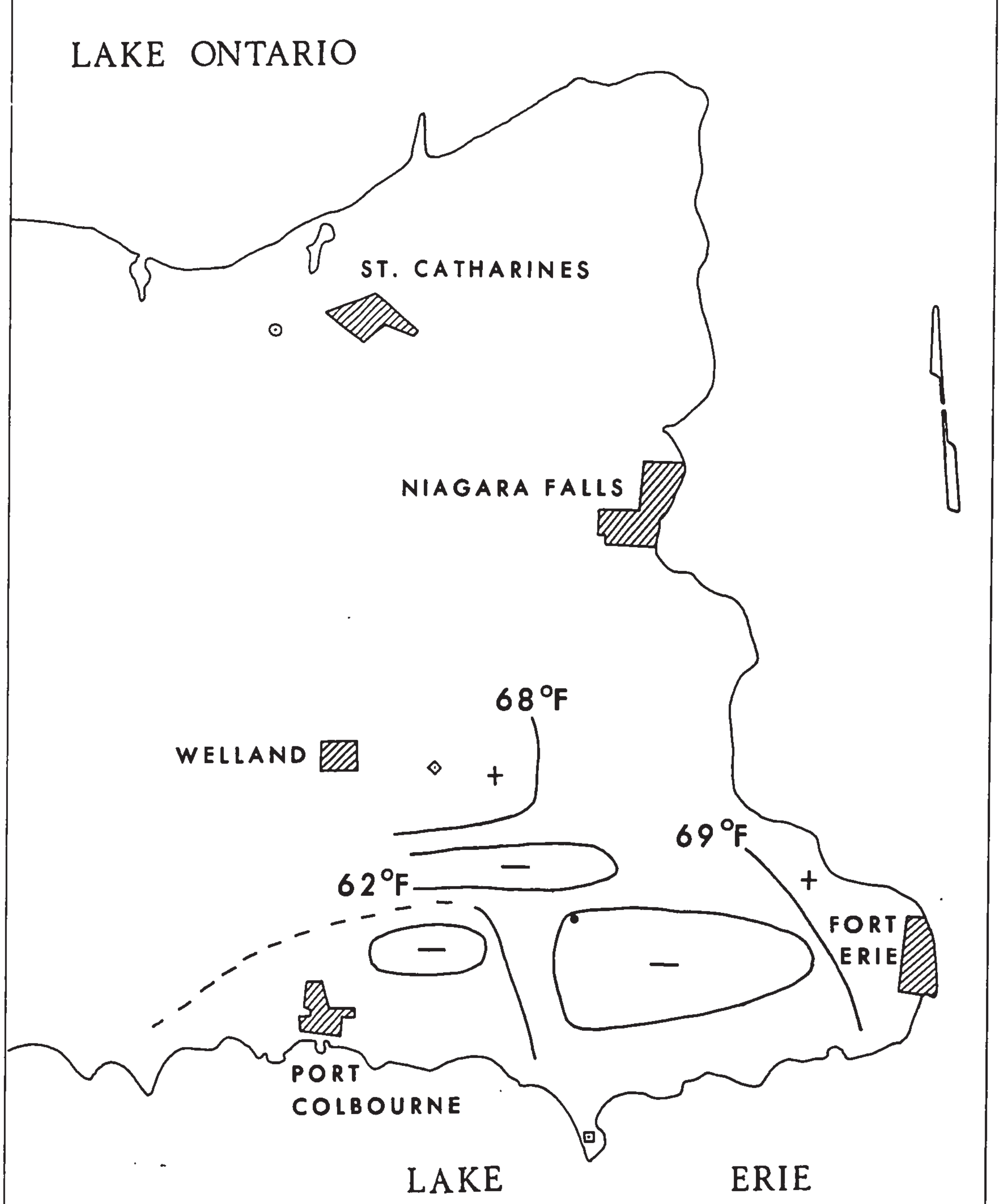
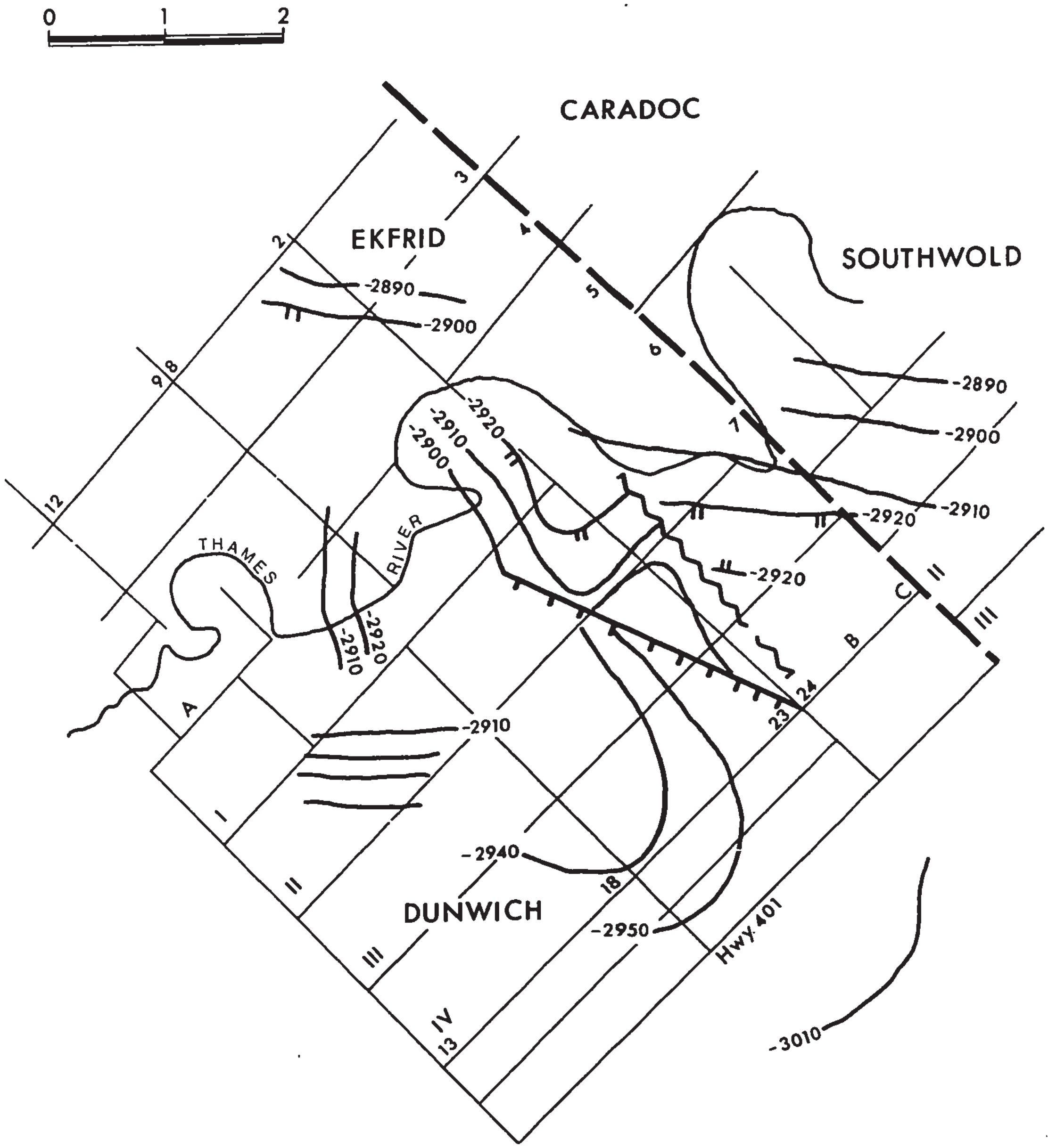


FIG. VI-11
ISOGEOTHERMS 100 METERS
BELOW SEA-LEVEL IN REGION 3.
1 INCH = 4 MILES APPROX.



FAULT

DEPTH CONTOURS ON BASEMENT

FIG. VI-12

BASEMENT STRUCTURE CLOSE TO THE DUNWICH BOREHOLE

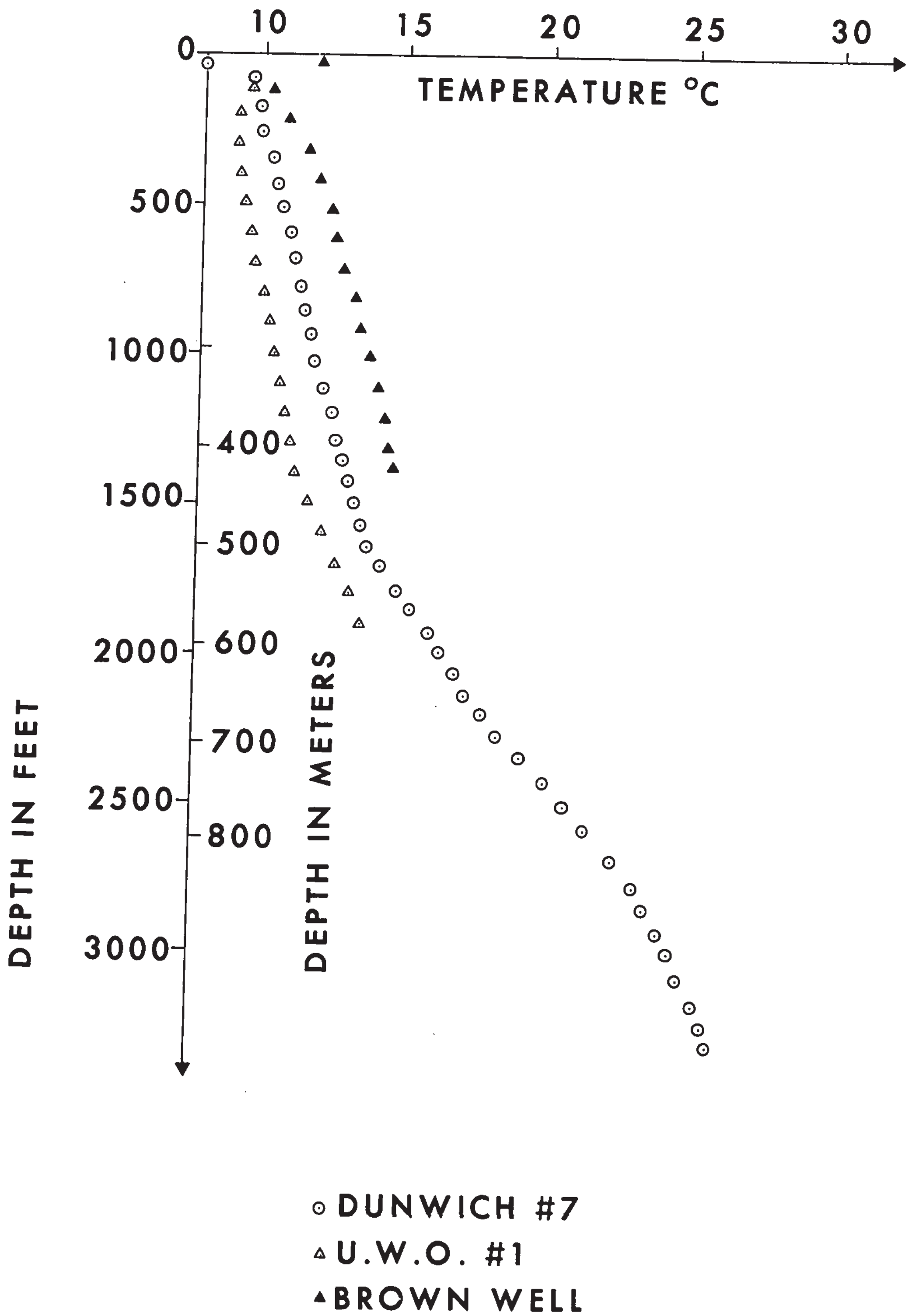


FIG. VI-13
BOREHOLE TEMPERATURES IN REGION 4

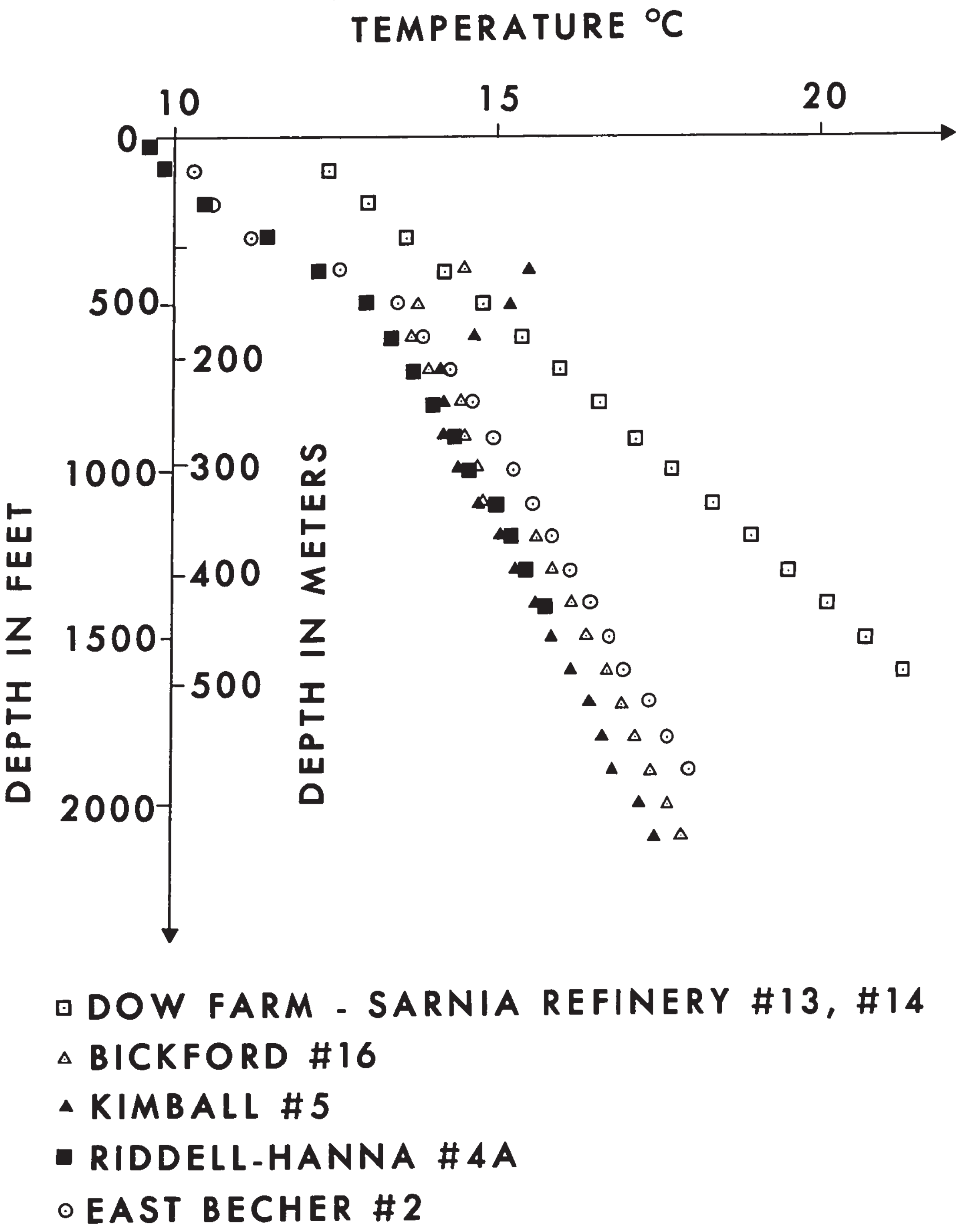


FIG. VI-14
BOREHOLE TEMPERATURES
IN REGION 5

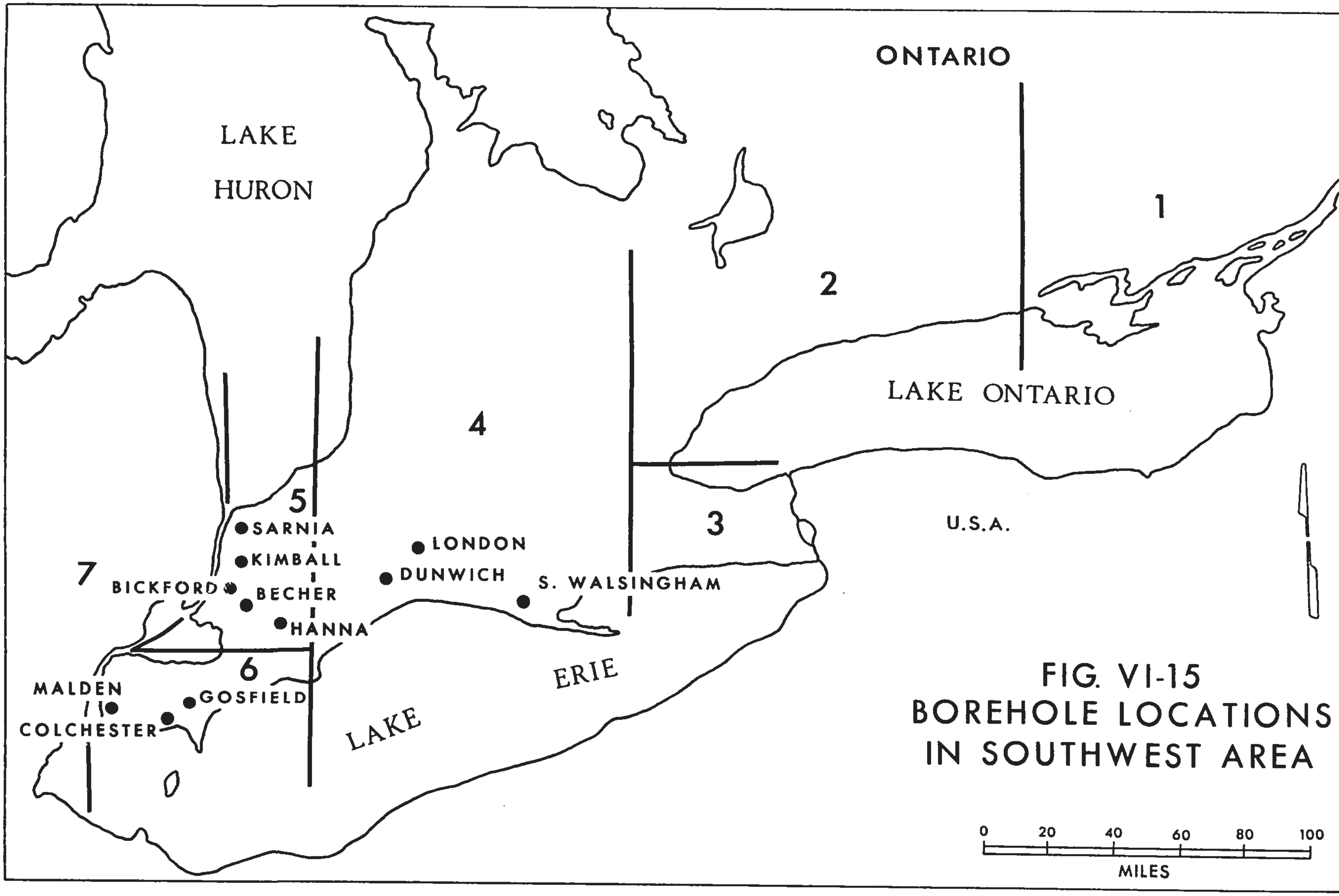
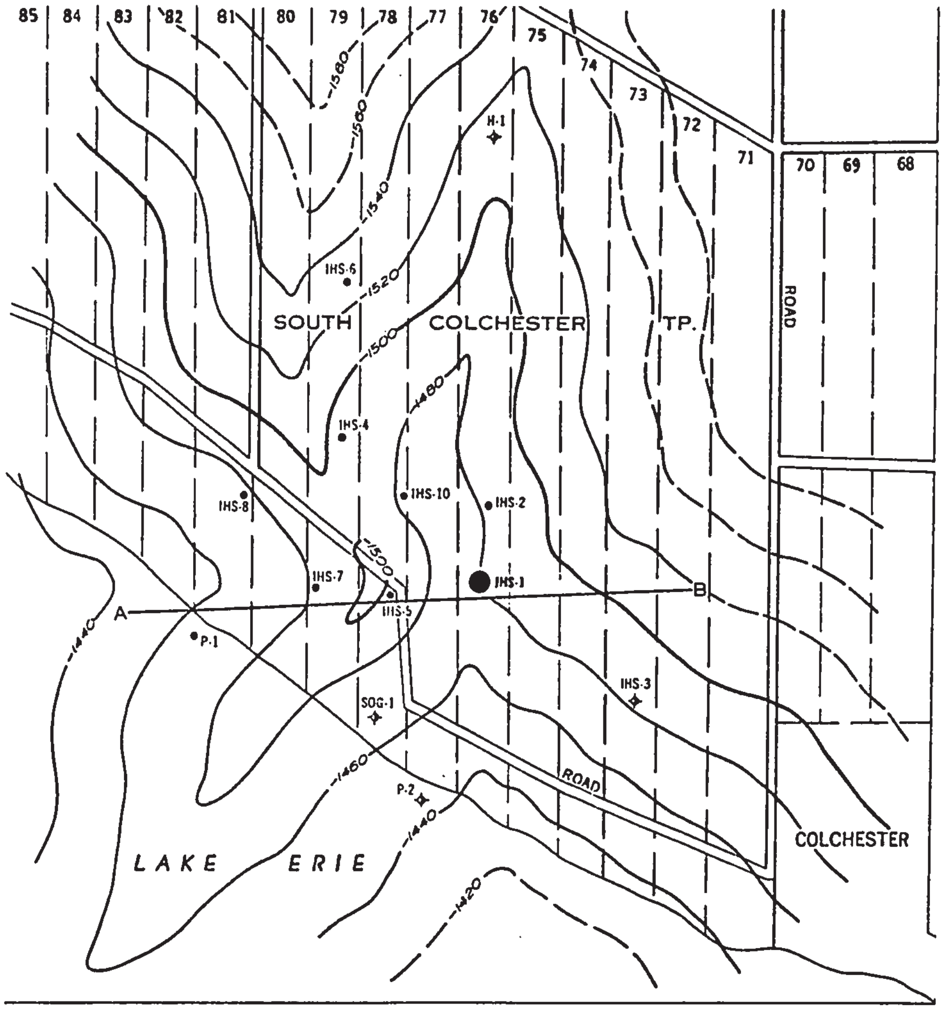


FIG. VI-15
BOREHOLE LOCATIONS
IN SOUTHWEST AREA



LEGEND

Well, with commercial flow of oil ....•
Well, with show of oil ....✦
Structure contours on Cobourg formation (interval 20 feet, datum mean sea-level)..... 1500
Lot numbers, (front concession).............68–85
Imperial-Harvest-Submarine..................IHS-1-10
Harvest Petroleums Ltd.......................H-1
Place Gas and Oil Ltd.........................P-1-2
Southern Ontario Gas Co......................SOG-1

Scale of Miles

0 ½ 1

GSC

FIG. VI-16
LOCATION OF COLCHESTER SOUTH BOREHOLE

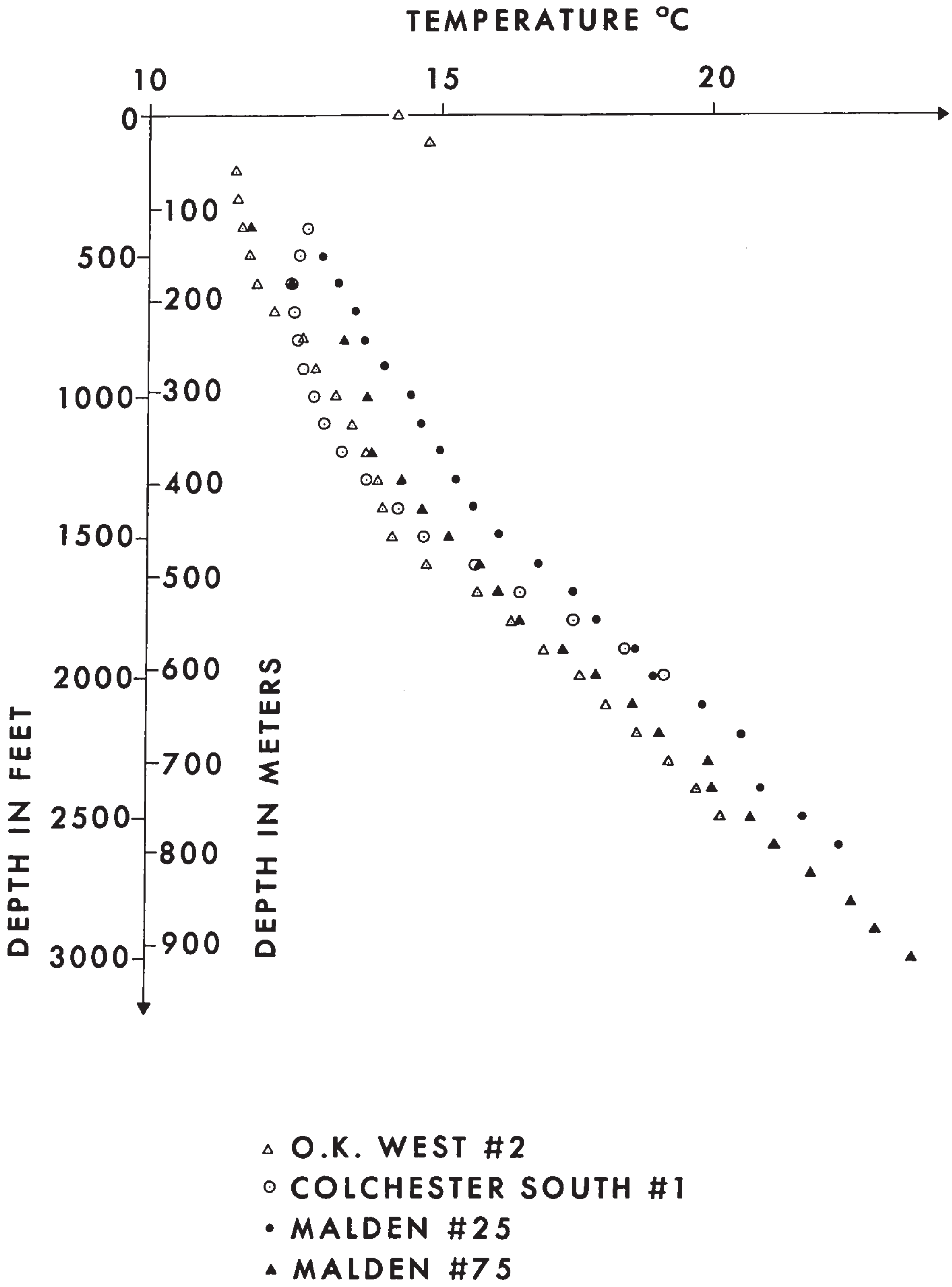
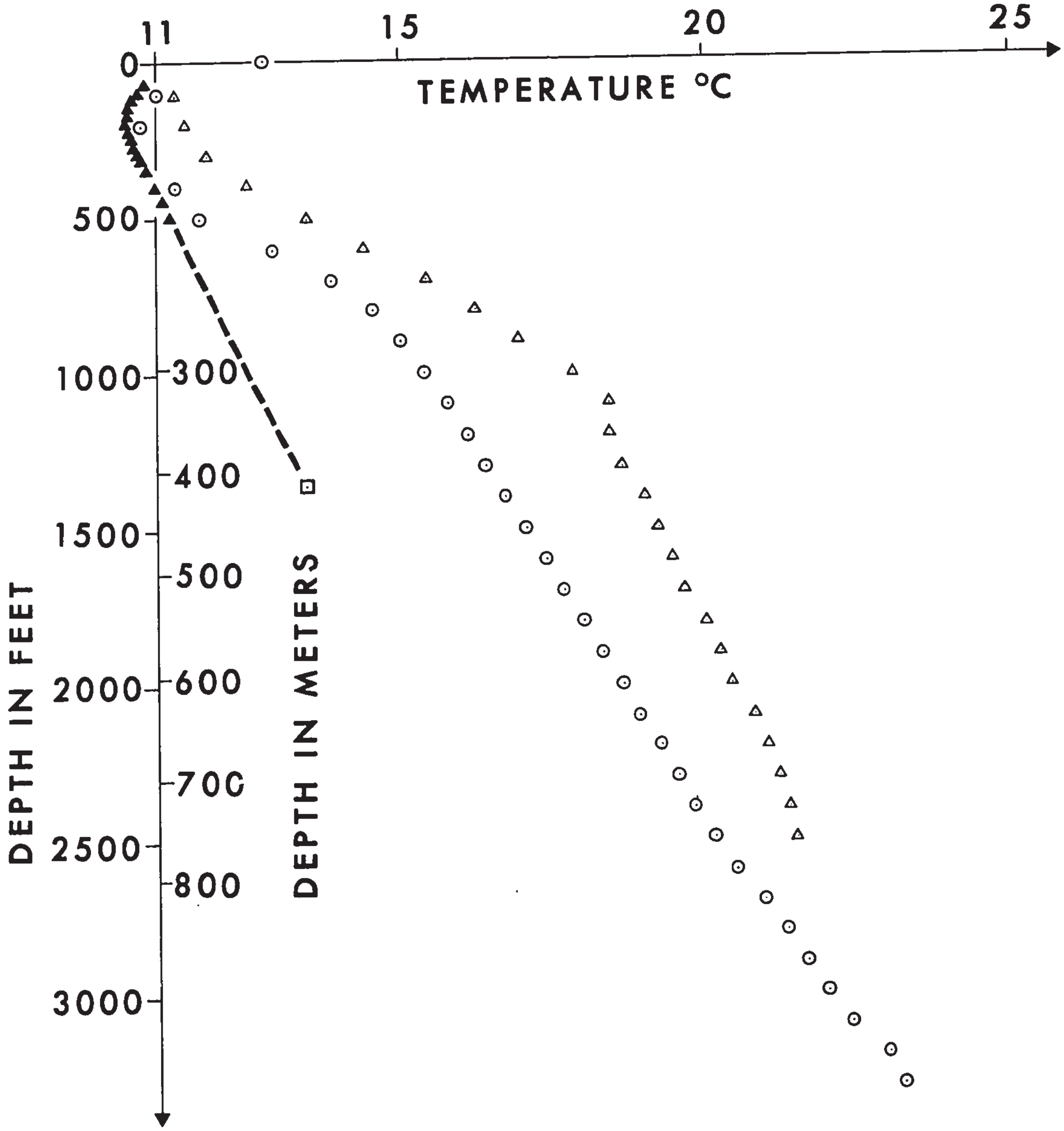


FIG. VI-17
BOREHOLE TEMPERATURES IN REGION 6



○ WELL NORTHVILLE #106
△ WELL MUTTONVILLE #2
▲ WELL WAYNE CO. #2
⊡ WELL MELVINDALE I. S. CO.

FIG. VI-18
BOREHOLE TEMPERATURES IN REGION 7

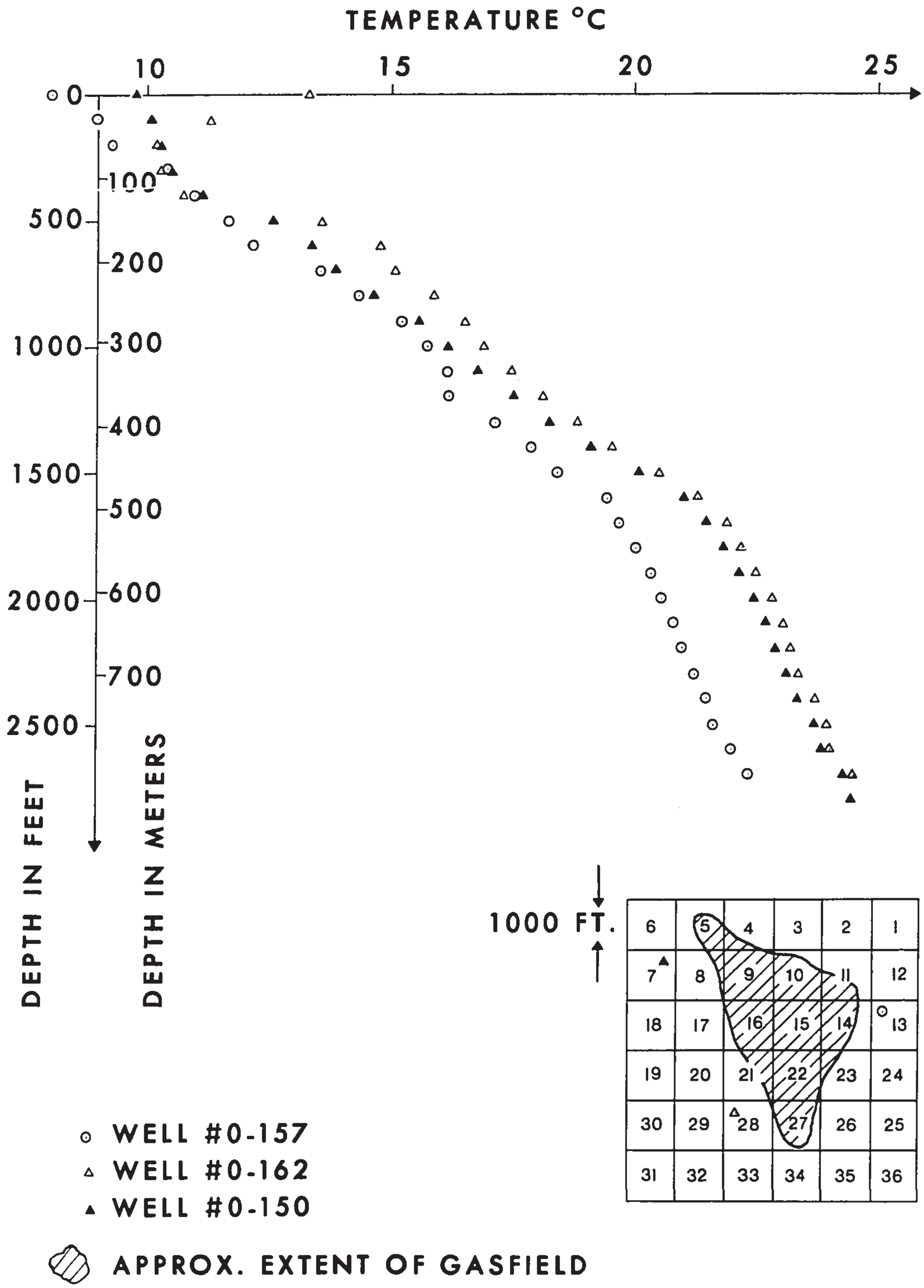


FIG. VI-19
BOREHOLE TEMPERATURES IN REGION 8

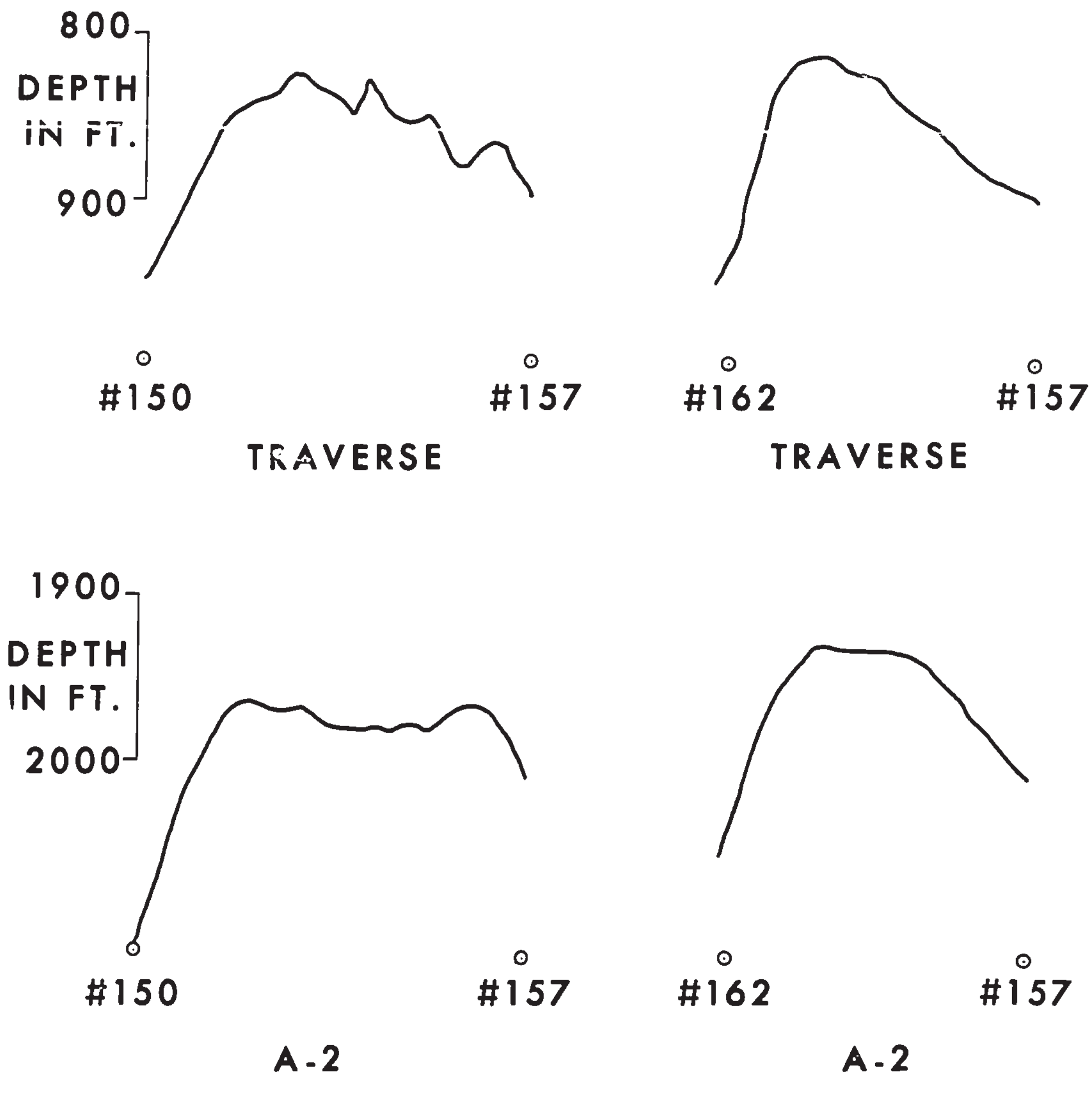


FIG. VI-20
SECTIONS ACROSS OVERISEL FIELD,
ON TWO HORIZONS

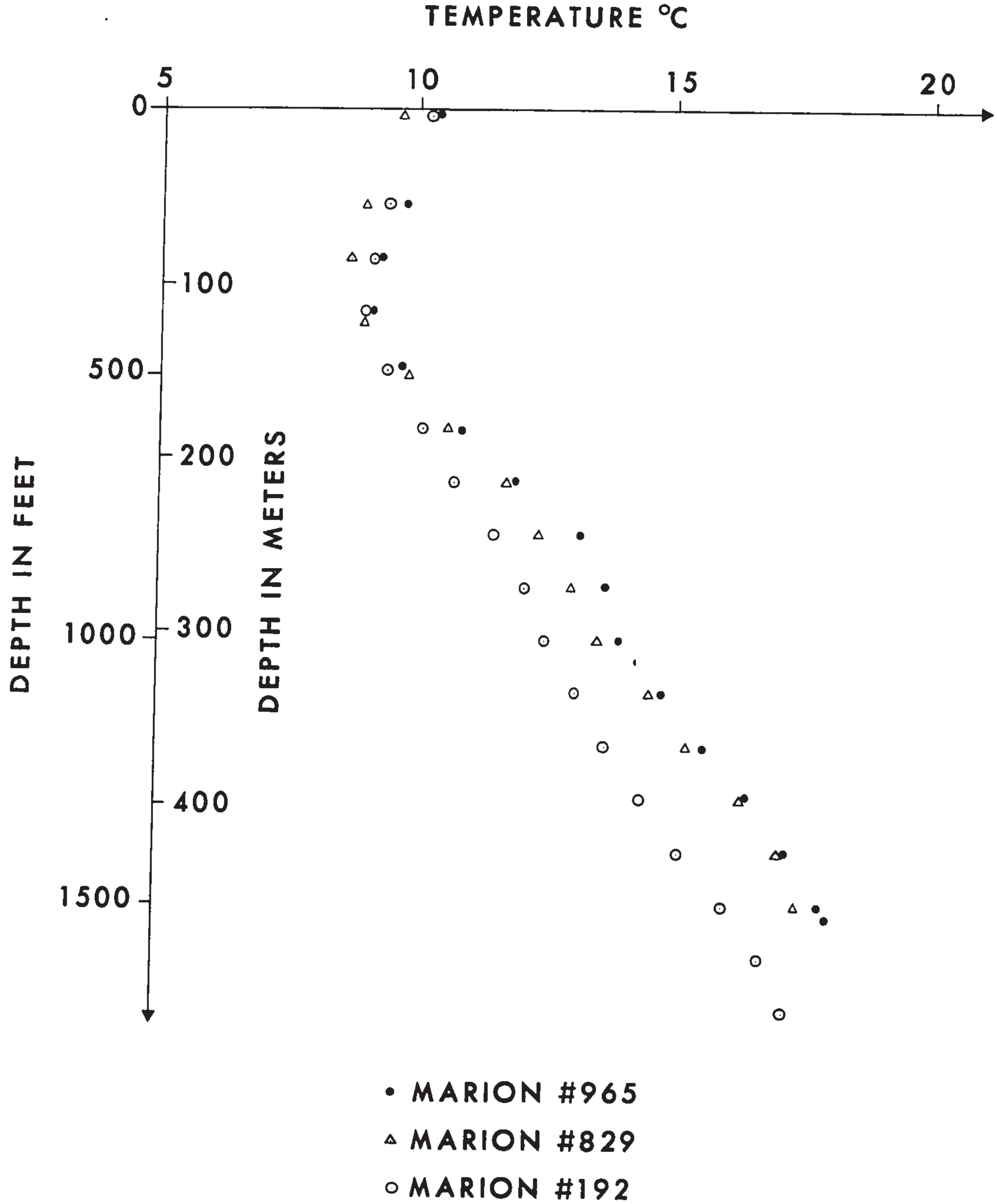


FIG. VI-21
BOREHOLE TEMPERATURE IN REGION 9

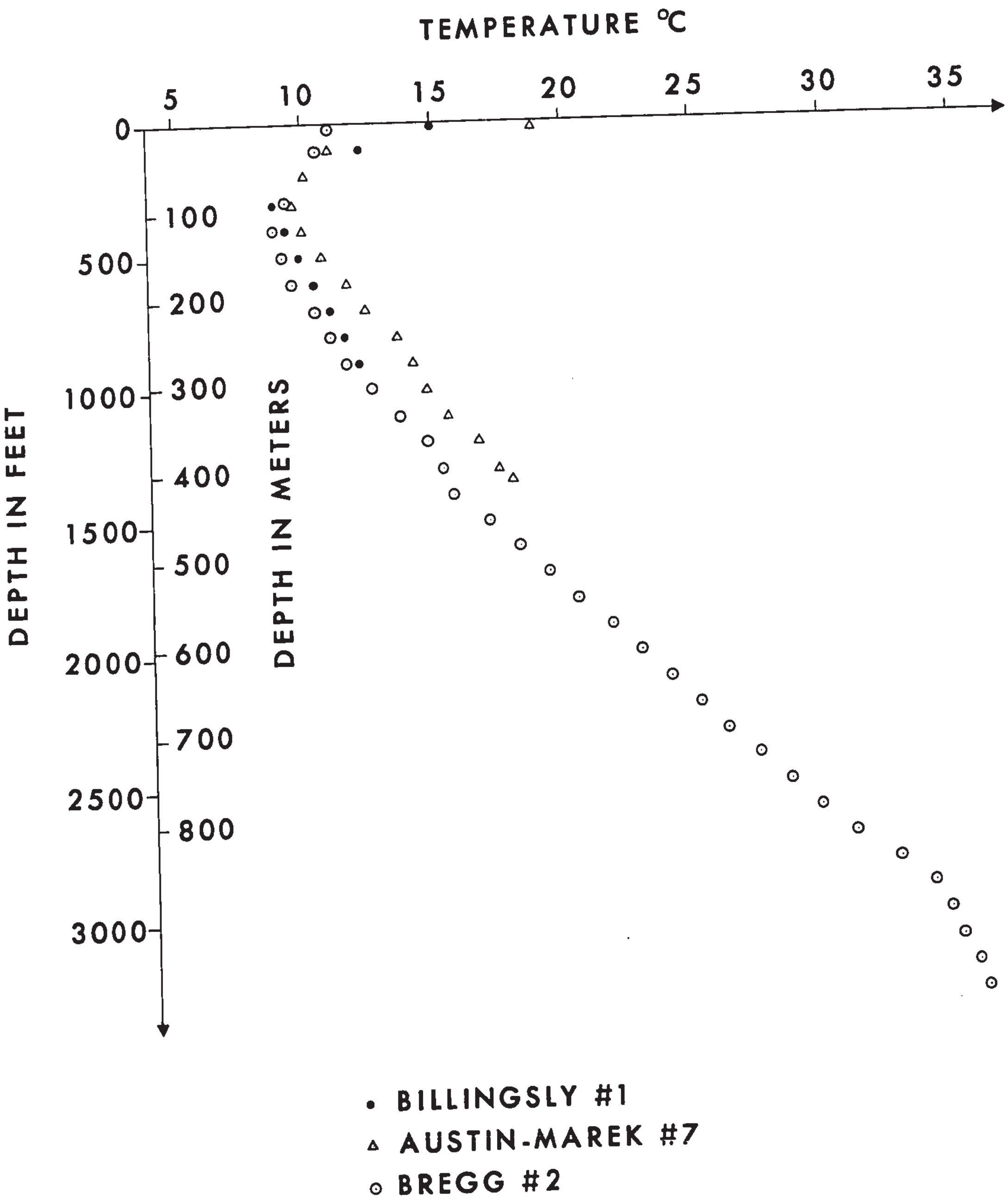


FIG. VI-22
BOREHOLE TEMPERATURES IN REGION 9

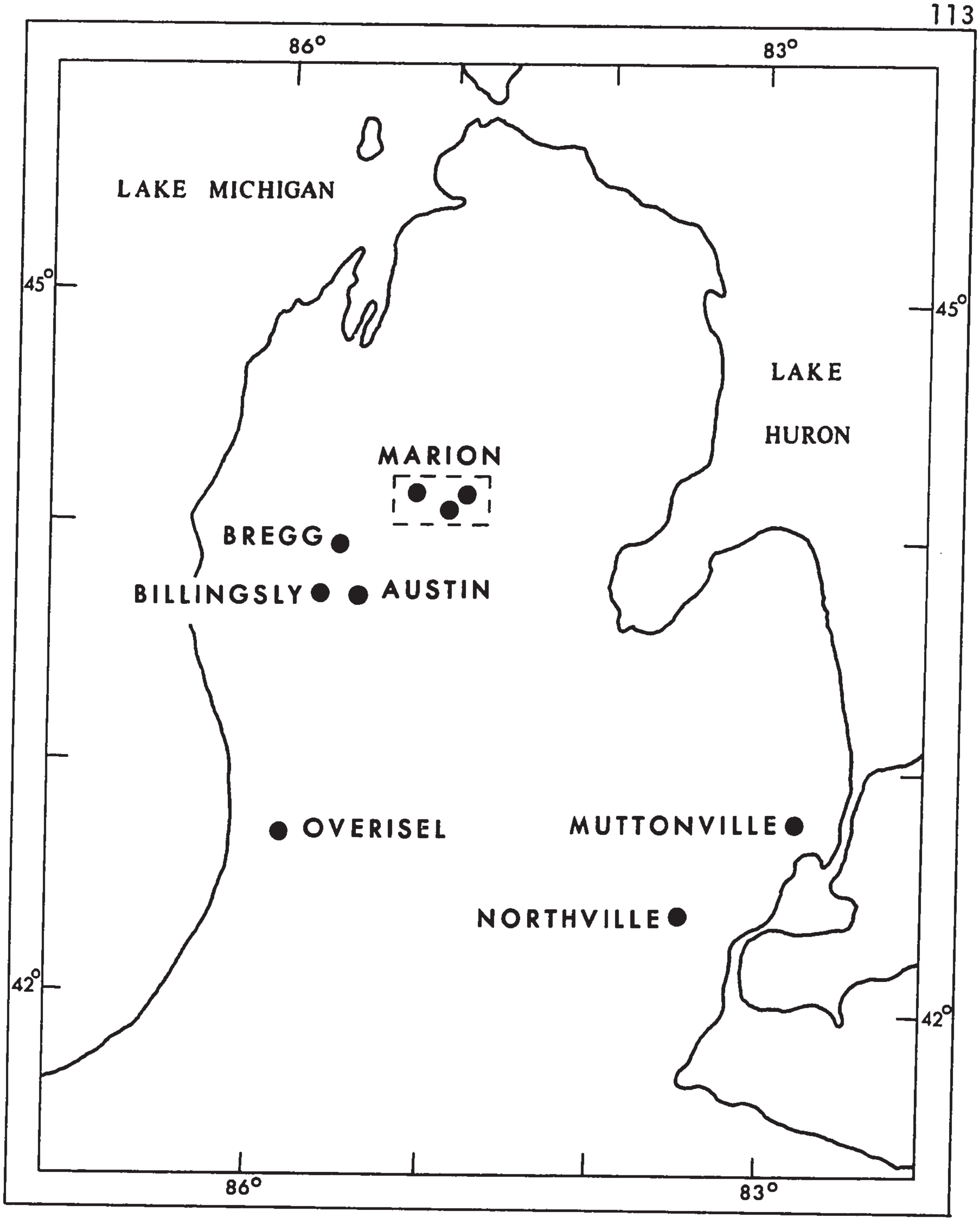


FIG. VI -23
BOREHOLE LOCATIONS
IN THE MICHIGAN BASIN

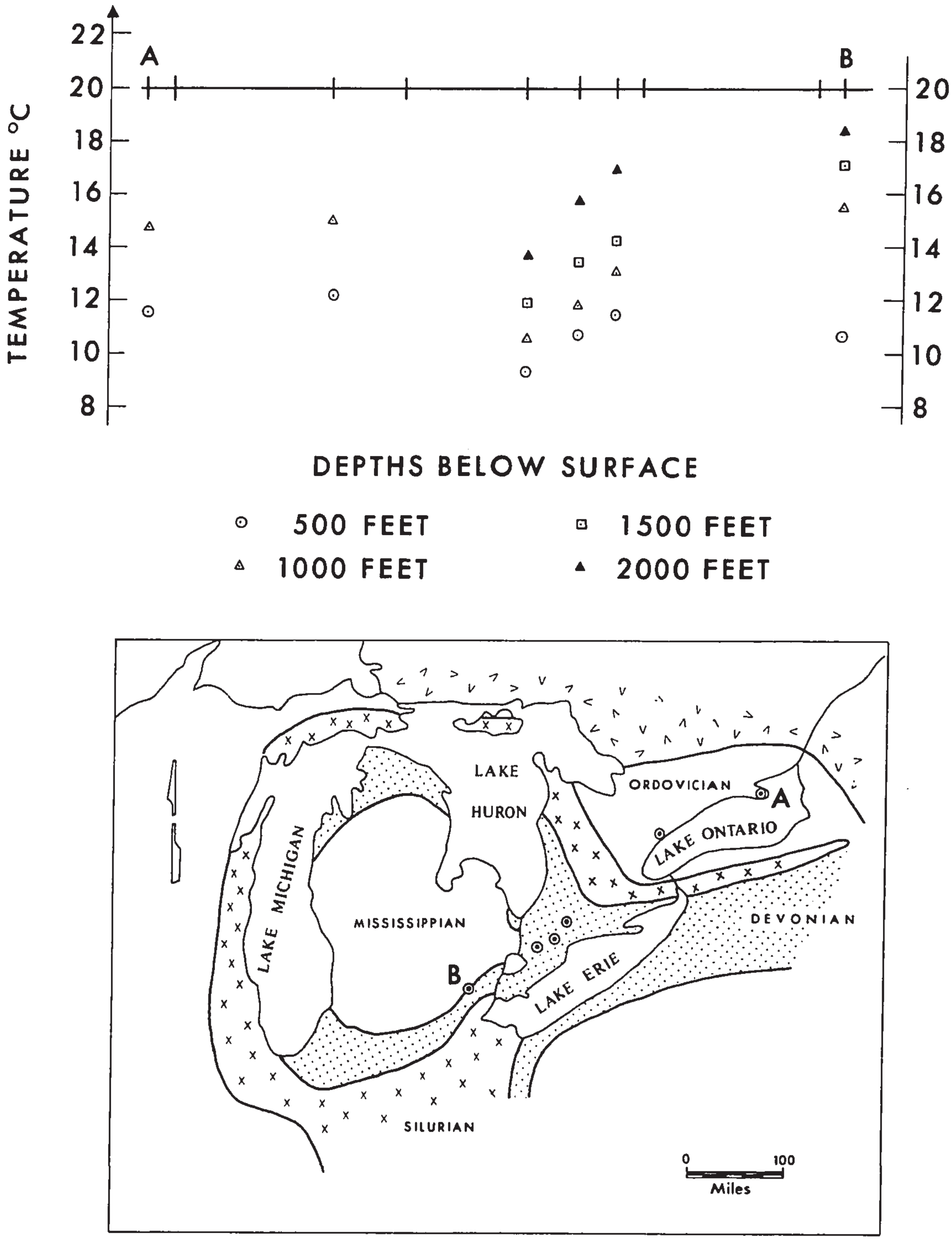


FIG. VI-24 TEMPERATURE DISTRIBUTION ACROSS THE REGION

CHAPTER VII

VARIATION OF THERMAL CONDUCTIVITY

VII-1 PREFACE

In this chapter the horizontal and vertical variation of thermal conductivity is examined. The thermal conductivities used were all measured on the divided bar and have been corrected for contact resistances, and for diameter differences as discussed in Chapter IV. All of the samples were water-soaked and all measurements were made under axial loads of 800-1000 psi at room temperatures between 22 and 26°C

A comparison is made of the average conductivities and of the range of values. Certain horizons are discussed individually giving regard to the composing units and to lithologic changes that may occur spatially through the region.

By an examination of these regional thermal conductivity patterns a technique of using temperatures in uncored holes and estimating the conductivities or alternatively using temperature gradients and conductivities in several regionally uniform formations rather than complete boreholes for the calculation of heat flow might be realised.

The general geology of the region was discussed in Chapter II; now each of these horizons is also described in greater detail.

VII-2 FORMATION THERMAL CONDUCTIVITIES

VII-2-1 Pre-cambrian

The basement rocks underlying the Palaeozoic of Southern Ontario and part of the Michigan Basin consist of granites and metamorphosed sedimentary rocks. These have been dated by Stockwell (1965) at 900 million years and thus form part of the Grenville series of Late Precambrian age. General lithology of this Province and of the others is discussed at greater length in Chapter IX. Depth to basement varies from less than 610 m in Toronto to over 3700 m in the centre of the Michigan Basin.

Very few cores penetrate to basement and where they do, bottom-hole is usually only 3 m into it. During this work a total of 17 discs were measured from 11 holes shown in FIG VII-1. Four of these discs showed severe weathering. Ignoring these, the mean conductivity of the samples is 7.1 ± 0.2 mcal/cm.sec.^{o}C , suggesting a uniform conductivity of unweathered material from place to place within the region. Many holes show a weathered layer at the top of the Precambrian which reaches a thickness of over 61 m at Ottawa. Since few of these holes penetrated to any great depth in the Precambrian it was felt that some measurements of the vertical as well as the spatial distribution were necessary. A borehole drilled by the Dominion Observatory at Ottawa penetrated 400 metres of Precambrian rocks descriptively similar to those of Southern Ontario. However the average conductivity in the Ottawa hole is very much lower than the average over southern Ontario, being only 5.7 ± 0.8, with only 11 out of the 50 measurements in the range of the unweathered southern results. The Ottawa discs were measured dry but soaking them increased the values by a maximum of 3%; the porosity is very low and must be centred in microfractures. Information on the Ottawa densities was not available so this parameter could not be compared. The variation does suggest that several more holes cored in the Precambrian are required across the region so that a detailed mineralogical study might be made particularly of quartz content.

The conductivities obtained in Southern Ontario are consistant with Misener's results in Northern Ontario (1951). The granites are of similar conductivities to the horneblende gneisses.

VII-2-2 Cambrian

There is an unconformity between the Cambrian sediments and the Precambrian marked by thick layers of weathered rocks at the surface of the latter. The first transgression of the sea over the Algonquin Arch in Palaeozoic times was during Upper Cambrian times. These sediments are subdivided by Sanford et al (1959) into three units consisting of a basal sandstone (Mount-Simon and Potsdam), interbedded sandstones and carbonates (Eau-Claire and Theresa) on top of which are carbonates (Trempeleau-Little Falls). The Cambrian formations pinch out against the Arch over most of central Southern Ontario and reach a maximum thickness of 61 m in Essex County and in the southern part of the Niagara Peninsula. They appear to thicken into the centre of the Michigan Basin to greater than 730 m in Ogemaw County. In the past few years some coring has been done in this section but this has been reduced as the Cambrian programme of most companies has been curtailed or stopped. No attempt was made for this report to make measurements on the Cambrian section since its occurence is highly localised. However for the sake of a complete profile, three composite discs in the U.S. Steel hole, each representing a one-foot section, were measured.

U.S. Steel	
3511/6	12.6
3610/0	8.8
3756/0	7.0

Sanford's log of the hole shows the Theresa unit to be the only one present.

The top disc is a darkish dolomite, the lower two are coarse sandstones. Several results were also available for the Potsdam sandstone from the Russell borehole. A total of 13 measurements give a mean conductivity of 12.1±2.1. Dry measurements on a similar sequence in the St. Lawrence Lowlands (Doig, 1961) gave results of 11.0±1.5. Both lithologies are composed of sandstones, arkoses and quartzites.

VII-2-3 Ordovician

A Trenton-Black River Groups

This group of limestones, shaley limestones and dolostones lying unconformably on the Cambrian are of Middle Ordovician age. Their total thickness varies from 61 m at the top of the Bruce Peninsula to 301 m under the eastern part of Lake Erie and the western part of Lambton County and thickens to 300 m in the central basin (Cohee, 1948). Sanford (1960) subdivides the Black River horizons into a basal unit of shaley or sandy dolomite, dolomitic shale and arkoses (Shadow Lake) succeeded by a brown to cream coloured lithographic limestone (Gull River) and a finely crystalline to granular buff limesone (Coboconk). On top of this are the formations of the Trenton group consisting of a medium grey, shaley carbonaceous partly fragmented limestone (Kirkfield), a fragmental limestone with interbedded shale (Sherman Falls), succeeded by a dark brown dense argillaceous limestone(Cobourg).

A total of 173 measurements of thermal conductivity have been made of rocks in this group and further measurements were available from Observatory measurements on cores from Eastern Ontario boreholes and from

Misener's results at Toronto (1951). Of the author's results 85 of the samples came from the Colchester township area of Essex County.

The explanation for this high concentration of samples from one area is that oil and gas fields in these formations, ie., the potential sites for temperature measurement, occur in the porous dolomites or dolomitised fracture zones. Thus a survey of such an area may give information on the occurence of dolomites and limestones and their inter-relationships and assist in making conductivity estimates. The measurements also give an idea of the local variation of conductivity which is important in putting limits on the heat flows. FIG VII-2 shows the regional distribution of thermal conductivity measurements.

The regional spatial variation is first considered. Using the results for the U.S. Steel, CNE, Russell, Picton and the Colchester-2 holes, the last of which showed less dolomitisation than the other holes in the Colchester area, the average conductivities of the Trenton-Black River formations are:-

Colchester-2	6.7 ± 0.7	South-west
U.S. Steel	6.4 ± 0.6	
C.N.E.	6.1 ± 0.8	
Russell	6.6 ± 0.6	↓
Ottawa	5.8 ± 0.6	
Picton	5.0 ± 0.5	North-east

Table VII-1 MEAN TRENTON-BLACK RIVER CONDUCTIVITIES

From east to west across the region the formation conductivities are generally increasing. This is due to an increasing shaliness of the cores in the eastern region, particularly Picton. Conductivity variations between Essex County and Toronto would seem to be insignificant except areas of heavy dolomitisation. This problem will be considered again in the light of results from the Colchester region.

Not all of the formations of the Trenton-Black River group are always present so it is useful to divide the results by formation and see if any systematic differences occur between them. The table below shows the means in the different formations as well as the standard deviations. Where the formation was not present there is an 'n' in the column. Since there are so few samples in some of the units the results beyond a Trenton and Black River division may not be reliable.

Borehole Name	Thermal Conductivities							
	TRENTON				BLACK RIVER			
	TOTAL	Cobourg	Sherman	Kirkfield	TOTAL	Coboconk	Gull R.	Shadow L.
Colchester-2	6.3	n	6.3±0.6	n	6.9±0.4	n	6.9±0.4	n
Colchester-6	7.9	n	7.9±1.6	n	9.4±1.7	n	9.4±1.7	n
Colchester-7A	6.8	n	6.8±1.1	n	6.4	n	6.4	n
Sarnia	n				6.2	n	6.2	n
Romney Gore	6.1	n	6.1	n	7.4	n	7.4	n
Bidwell	7.9				n			
Saltfleet 11-1	n				5.3			
Manitoulin Road Cut	6.6				n			
U.S. Steel	6.3±.3	6.2±0.2	6.3±0.3	6.8±0.3	6.8±0.7	n	6.8±0.7	n
C.N.E.	5.3				6.5			
Ottawa	n				5.9±0.6	n	5.9±0.6	n
Picton	4.9±.6	n	4.9±0.6	n	5.1±0.5	n	5.1±0.5	n
Russell	n				6.6±0.6	n	6.6±0.6	n

Table VII-2 THERMAL CONDUCTIVITIES IN THE TRENTON AND BLACK RIVER

Conclusions to be drawn are as follows:-

1) The Trenton conductivities are more uniform than those of the

Black River.

2) Trenton conductivities are generally lower than Black River values.

3) Only in the U.S. Steel hole are units recognised which subdivide the Trenton.

4) The U.S. Steel record shows increasing conductivities through the Cobourg, constant values in Sherman Falls with a large drop at base corresponding to increased shaliness, higher conductivities in the Kirkfield, even higher at the top of the Black River, dropping to very low values and then average values in bottom section, as shown in FIG VII-3.

5) In general, density follows conductivity with the highest densities occurring in highest conductivity zones.

6) Porosity is constant with the exception of zones of high conductivity where it increases. These zones are generally dolomitic. The relation between thermal conductivity, porosity and density is examined in VII-3.

As mentioned before, the conductivities in the Colchester region are of course of very great significance for the following reasons:-

1) Holes drilled by oil companies and particularly suspended holes awaiting development are usually in anomalous areas.

2) Very often only one borehole in a field is available for temperature measurement.

3) Core is rarely available for that particular hole but is usually available for a limited number of holes and generally only the reservoir section is cored.

In the Colchester area the production reservoir is in the Trenton and production occurs particularly in dolomitised zones in the limestones. Cores were available for three holes shown in FIG VII-4 as C2, C6 and C7A. The bulk of the samples came from between the top of the Sherman Falls and the base of the Gull River. Mean conductivities in the three holes are 6.7 ± 0.7 in C2, 8.5 ± 1.7 in C6 and 7.1 ± 1.0 in C7A. A histogram of the conductivities in the three holes is shown in FIG VII-5. The highest formation conductivities and standard deviations are found in C6 which lies in the synclinal structure on the west side. In this well the average conductivity in the Black River is much higher (9.4) than that in the Trenton (7.9). Examination of the samples indicates that those with higher conductivities are either dolomitic limestones or dolostones, whereas most of the lower ones are limestones or shaley limestones. This relation is true in each of the holes and thus since dolomite is denser than calcite and since average formation porosities are low, 0.6 to 1.0%, in each of the holes, the densities should vary proportionally with the conductivities. This in fact does happen, the average densities being 2.77, 2.68 and 2.65 in C6, C7A and C2 respectively.

Although conductivities in these sections vary from 5.1 to 11.8, the average section conductivities vary by only 30% in the three holes with 20 to 30 samples per hole. Thus this is the order of magnitude of the maximum error to be expected in a heat flow determined from a temperature gradient in this region. This could be further improved by having a density log profile carried out in the hole in which gradient measurements were made but for which no core was available. It would seem reasonable to conclude that there are no problems in calculating heat flow even under difficult circumstances such as these, although of course the isoflux lines may be distorted by the structure and require a correction. Deeply penetrating sonic logs together with surface seismic reflection profiles would assist the structural interpretation.

The measured variation of thermal conductivity in any Black River-Trenton section in the region is a maximum of 30%, this variation being composed of an increasing shaliness in the eastern area and zones of local dolomitisation in the west. With the addition of density logs and a good core description many of the variations would be predictable. Thus, because of these facts, plus its thickness, widespread occurrence in the subsurface and its interest to the oil companies as a reservoir horizon, this group is judged to be a useful one in which to make heat flow determinations.

B <u>Collingwood and Blue Mountain</u>

No cores were available for the Collingwood and Blue Mountain shales of the Upper Ordovician in Southern Ontario. They usually occur as very fissile dark grey to grey shales with occasional thin laminations of grey argillaceous and silty limestone and cores rapidly become very broken up. One measurement was available from the Collingwood formation in the Russell hole east of the Frontenac axis. This was a limestone with a conductivity of 6.2. East of the Frontenac axis the Billings formation of brown to black shale is found beneath the Blue Mountain and on top of the Trenton. Five conductivities were obtained in the Russell borehole for this section giving a mean conductivity of 4.4 ±1.5.

C <u>Meaford-Dundas Formation</u>

The Meaford-Dundas is described by Sanford (1961) as a grey, calcareous siltstone and grey silty limestone, with minor interbedded greenish grey shale at top, grading into a greenish-grey to medium grey shale with interbeds of crystalline limestone and siltstone. The formation is fairly constant in thickness averaging between 61 m and 91 m. Since a section of this unit occurred in the U.S. Steel borehole, conductivity measurements were possible. The mean value was 6.3 ±0.4 on 6 samples with generally lower conductivities at the top of the formation. One further measurement was made on the lower section of the U.W.O. core which penetrated a few feet into the Meaford. This value was 5.7. Mean density of all the samples is 2.54.

The eastern equivalent of the formation is the Carlsbad which commonly consists of shales with limestone interlayers. The conductivities of two samples, described as a quartzite and a sandstone, in the Russell hole were 8.8 and 7.3 respectively.

In spite of the fact that in some parts of Southern Ontario the combined thickness of the Collingwood-Blue Mountain and Meaford-Dundas is 183 m, the condition of the cores and the small number of them do not make them a good prospect for heat flow measurements. Locations of sample sites are shown in FIG VII-6 for both the Meaford-Dundas and the next formation, the Queenston.

D Queenston

Sanford (1961) describes the formation as fairly variable in lithology being commonly a brick red shale with green shale and grey crystalline limestone interlayers throughout the Niagara region and a distance to the north, and in Bruce and Essex Counties grading into a grey medium crystalline dolomite at the base. The formation, varying in thickness from 45 to 370 m between the Bruce and the eastern part of

Lake Erie, represents the youngest Ordovician strata present in Southern Ontario. In southeast Michigan the Queenston is recognised as a red shale but further west the Upper Ordovician becomes a laterally variable series of grey shales and interbedded limestones and dolomites (Cohee, 1948).

As with the Meaford-Dundas this formation is only cored infrequently in stratigraphic tests, so few cores are available. The complete section was cored in the U.W.O. hole and a few sections of core were available from an offshore well in Haldimand County which permits some comparisons. A few samples were also available from a Toronto brickyard.

The mean conductivity of 32 samples in the U.W.O. hole spaced at 3 m intervals is 5.3 ± 1.0. In the lower section of the borehole the conductivity is higher than in the upper because of the increased occurrence of limestone interlayers. Two samples from the Place Walpole hole give a conductivity of 5.2 and the brickyard sample 4.7. Four measurements were made on the sandy limestone beds from the Niagara area which are used to face the University buildings. The mean conductivity was 8.3 ± 0.2.

In Chapter XI it is apparent that the lithological section giving the most uniform heat flow, undisturbed by water-flows in the U.W.O. borehole is the Queenston. Since shales are usually impermeable one such section should be used in heat flow calculations. For this reason plus that of the reasonable competance of the rocks for cutting and its great thickness, this formation should be a very useful one for heat flow purposes. Ideally more coring should be done to improve information on its spatial variation.

VII-2-4 Silurian

A Clinton-Cataract

The Clinton-Cataract groups are the oldest of the Silurian and lie unconformably on the Ordovician strata. They consist mainly of clastic rocks, shales and sandstones, with some carbonates. The beds are everywhere thin shallow water deposits reaching a greatest group thickness of 73 m beneath the eastern end of Lake Erie. To the south the group grades into a red-bed sequence.

Cataract strata, according to Sanford (1962) consist of the Whirlpool sandstone and its equivalent the Manitoulin dolomite with a gradational contact. Overlying it is the Cabot Head which is typically a shale. In the Niagara area the Manitoulin and Cabot Head are not separable but grade into an arenaceous shale and sandstone. Above them is the reddish coloured sandstone facies of the Grimsby which grades northward into the Dyer Bay dolomite, the green and red shales of the Wingfield and the St. Edmund dolomite.

At the base of the Clinton group is the Thorold sandstone resting unconformably on the Grimsby. It grades to the north and west into dolomites of the Reynales and locally to the shales of the Neagha. The Reynales is present throughout the region west of Niagara. Overlying this in Niagara the Irondequoit consists of a grey crystalline dolomite succeeded by the dark grey dolomitic shales of the Rochester formation. Core samples for this section were available in the U.S. Steel, Place Walpole and U.W.O. holes shown in FIG VII-7 and the values are summarised in the table following.

Each formation in the group forms a thin section which is lithologically very different from those on either side. Even within a section the conductivity variation is very large as is evidenced by the standard deviations on the Cabot Head. The formations also change facies across the region making the group a very unsuitable one in which to determine heat flow.

Group	Unit	Borehole Conductivities		
		U.W.O.	Place Walpole	U.S. Steel
Clinton	Rochester	5.4 ± 0.8	4.2	6.5
	Irondequoit	n	9.2	12.0
	Reynales	8.5	n	n
	Thorold	n	8.6	n
Cataract	Grimsby	n	11.6 ± 0.5	11.4
	Cabot Head	4.6 ± 1.6	5.9 ± 2.1	n
	Manitoulin Whirlpool	5.6	9.1	7.1
Total No. Samples		16	22	6
Total Length Section (metres)		49	57	61

Table VII-3 THERMAL CONDUCTIVITIES IN THE CLINTON-CATARACT FORMATIONS

B <u>Guelph-Lockport</u>

The Guelph and Lockport formations of middle Silurian age overlie the Clinton formation conformably. They are described by Sanford (1962) as consisting of buff, white and crean coloured dolomites ranging from fine to coarsely crystalline in texture and varying in thickness from 24 to 168 m. A thickening of the beds occurs to the north-east continuing into Manitoulin Island and northern Michigan, which seems to correspond to the existence of a reef bank around the Michigan Basin during the middle Silurian. Pounder (1961) divides the stratigraphy of the formations into lithological units which can be correlated throughout the study area. The lowest unit immediately overlying the Rochester is correlative with the Lower Albemarle in the Bruce and the Gasport in Niagara. Thickness of the unit varies from 0 to 118 m and the lithology grades from a brown crinoidal limestone or dolomite to a grey reefal dolomite. Most of the

Guelph stratigraphic section is composed of the Middle unit which is characteristically a brown-grey argillaceous finely crystalline to granular dolomite varying from 9 to 76 m in thickness. The change from Middle to Upper units appears as a break from a grey-to-buff crystalline fragmental dolomite to a brown argillaceous granular to finely crystalline dolomite. This unit varies in thickness from 3 to 30 m and is primarily composed of erosional products. In the centre of the Michigan Basin the Guelph formation thins and is present as a light coloured dolomite. On the limbs of the Basin it thickens into the reef complexes of southern Ontario and western Michigan.

Reefing is very common in each of the units usually in the form of bioherms but with the exception of a halo 30-40 km wide ringing the south and east shores of Lake Huron in which pinnacle reefs occur. Whereas the bioherms are usually dolomitic the pinnacles may be dolomites or carbonates depending on the nature of the formation into which they penetrate. For example Kidoo (1962) discusses some 39 reefs which had been discovered to that date, of which 20 are dolomite and 19 are limestone. Continuously cored sections were available in two boreholes; U.W.O. and MacGillivray 5-19; and top sections of the Guelph were available in a substantial number of holes. Comparing the continuously cored sections, the thermal conductivities are as follows:

U.W.O.	11.8 ± 0.4
MacGillivray	10.6 ± 1.6

Samples penetrating some metres into the Guelph give the following average conductivities:

Argor #1	10.9

Kimball #30	8.1
U.S. Steel	10.9
Union #25	11.1
Colinville #11	11.1
Central Mich.	11.0

Table VII-4 THERMAL CONDUCTIVITIES IN THE GUELPH FORMATION

Not as many small sections of this formation could be measured as would have been ideal, however the sites that were sampled are shown in FIG VII-7. The reason for this was partially the amount of core in good condition which could be positively identified as Guelph and partially the small amounts remaining in storage. Not as much Guelph is available as is given in the coring reports and storage reports, probably because much has been removed for laboratory studies. However there are sufficient results here to show the regional variation and its dependence on facies. In the sequence in the U.W.O. hole the conductivities are all greater than 10.5 and rocks are dolomites, of density 2.65 to 2.87 with porosities ranging from 1.2 to 6.9% The sequence of the MacGillivray hole has conductivities ranging from 7.0 to 13.0 with densities between 2.61 and 2.90 Several of the discs cut in this section were limestones. This picture is consistant with that of Pounder (1961), with zones of mixed limestones and dolomites in the northwest section (Region 5) whereas in the remainder of southern Ontario the formation is dolostones. Sanford has subdivided the Guelph-Lockport in the U.W.O. hole into several units which are given below together with the average conductivities:-

Unit	Conductivity	Density	Porosity
Guelph	11.8 ± 0.7	2.70 ± 0.08	3.2 ± 1.6
Goat Island	11.4 ± 0.3	2.75 ± 0.05	2.2 ± 0.8
Gasport	12.4	2.73	2.3

Table VII- 5 PHYSICAL PARAMETERS MEASURED ON THE GUELPH-LOCKPORT FORMATION IN THE LONDON BOREHOLE

Although the Guelph in the MacGillivray hole is not subdivided on the logs, the thermal conductivities suggest that a subdivision might be made into the following:-

Depth (m)	Conductivity	Density	Porosity
432 - 460	8.9	2.73 ± 0.13	5.9 ± 3.7
463 - 518	11.9 ± 0.6	2.75 ± 0.08	4.8 ± 2.3
524 - 562	9.7 ± 1.7	2.72 ± 0.10	5.1 ± 4.5

Table VII-6 PHYSICAL PARAMETERS MEASURED ON THE GUELPH-LOCKPORT FORMATION IN THE MACGILLIVRAY BOREHOLE

The variation in conductivity over the two complete sections is little more than 10% suggesting that this formation is likely to be fairly uniform across the region. Samples from the top of the Guelph vindicate this argument. The variations in the separate subunits of the Guelph-Lockport are insignificant in the U.W.O. hole but are quite large in the MacGillivray hole. Although the formation appears to be a good one for heat-flow measurements it would be useful to have further complete cores to see if the variations correlate with Pounder's subdivisions.

C <u>Salina</u>

The Salina is the name given to the evaporite deposits of the Upper Silurian. Landes (1945) has divided the Michigan sequence into seven subdivisions from A at base to G at the top. This is equally applicable in southern Ontario although the thick salt sequences of Michigan eventually pinch-out in this region. Landes description of the units is as follows:-

Series	Unit	Description
Upper Series	G	Fine buff dolomite, shaley dolomite, some anhydrite
	F	Grey shaley dolomite and dolomitic shale, locally red shale
	E	Fine buff dolomite, some shaley beds, anhydrite
	D	Grey shaley dolomite, dolomitic shale, anhydrite, locally red shale
	C	Fine buff and brown dolomite, buff dolomite, buff anhydrite, green shales
	B	Salt; where salt absent base marked by anhydrite
Lower Series	A2	Fine to medium brown to brownish grey dolomite Fine grey to dark grey dolomite, some dark shale Salt; where salt absent base marked by anhydrite
	A1	Buff to brown dolomite - fine to medium Fine to dense brown-grey and dark grey dolomite with dark grey shale Anhydrite at base

Table VII-7 LITHOLOGY OF THE SALINA FORMATION

Lower Series

Evans (1950) describes variations from the above description as follows: Both the A1 and A2 consist of lighter coloured upper parts and dark lower parts whether they are limestones or dolomites, and describes the A1 as only occurring in the northwest two-thirds of the region, whereas the A2 is general. Roliff (1954) shows a facies map in which the units are composed of dolomites to the east and south and are partially limestones and partially mixed limestones and dolomites in the north and northeast following the shaores of Lake Huron. The A-unit is overlain in the west part by the thick B-salt unit and in the east by argillo-calcareous beds. Total thickness of the A-unit varies from 15 to 137 m.

Upper Series

As well as occurring in the B and A2, commonly known as the upper and lower salts, salt can also occur as the D unit in Lambton County grading into dolomites, and in the F unit which grades into dolomitic shales and dolomites. The total salts reach a maximum thickness of 213 m in the Sarnia region and thicken still more to the westward and into the Michigan Basin. The occurrence of salt and its gradation into shales to the east, until only shales are found in the Niagara area, are discussed by Sanford (1965), Hewitt (1962) and Bolton (1957). Grieve (1955) has suggested that the absence of salt or its thinning in many areas is due to leaching by solvent waters moving along fractures and faults. This is to some extent borne out by the thickening of the Bass Island and Devonian sequences in some areas.

Numerous cored sections were available through the Salina, which for the purposes of the discussion of conductivities, is subdivided into the Upper and Lower Series. These sections include U.W.O., U.S. Steel, MacGillivray #5-19, Argor #1, Enniskillen #20-5, Plympton #1-3 and Midrim #'s 3A and 4. In the lower section numerous measurements were also made on short cores taken by oil companies as reservoir tests. All locations are shown in FIG VII-8.

Lower-Section Conductivities

Borehole	A_2 Carb.	A_1 Carb.
MacGillivray 5-19	8.5	
Plympton 1-3	8.0	
Enniskillen 20-5	9.0 ± 0.6	
U.W.O.	12.0 ±	11.4 ±
Argor 1	8.7 ± 2.5	10.8 ± 0.8
Becher 77		8.7 ± 2.2
Seckerton 6	7.4	8.7 ± 2.6
Corunna 17		7.3 1.4
Midrim #3A} Midrim #4	6.8 ± 2.2	
Sections with 3 or less samples:-		
Grand Bend 1		8.5
Warwick 1		8.5
Union 25	10.6	

Table VII-8 THERMAL CONDUCTIVITIES OF THE LOWER SALINA

The thermal conductivities of the A-units of the Salina tend to cluster into two groups of values which correlate with either dolomites or limestones forming the units. Mean conductivities for the limestone sections appear high for such a composition but this is caused by the sampling of anhydritic zones in order to provide a reasonable weighted average for the section. The lower mean in Corunna #17 is caused by a lack of anhydrite in the measured samples. Again the same problem arises for determination of heat flow - it is necessary to know if the section is composed of dolomites or limestones. However from section to section in the north-west region the mean conductivities appear to be fairly uniform and if the types of carbonates can be differentiated the conductivity estimates are not likely to be in error by more than 15%. Consistent with the facies distribution of Roliff (1949) the A-unit in the U.W.O. hole is entirely dolomites. However the Midrim holes in Adelaide township appear to be still in a limestone or mixed limestone and dolomite zone. In the U.S. Steel hole the Salina units have not been differentiated on the log and so trying to split it up would be mostly guesswork.

Several samples of the A_2 salt were measured to complete the conductivity picture. These gave a mean value of 14.0.

Upper-Section Conductivities

This section, as is apparent from the lithological description, shows a very wide range of conductivities varying from those of very fissile shales to those of salt and anhydrite. Several holes - U.W.O., U.S. Steel, MacGillivray #5-19, Argor #1, Enniskillen #20-5, Plympton #1-3 and Midrim #'s 3A and 4 - completely penetrated this section. The only samples taken from other short sections were of the salt horizons since this, in many instances, forms a large part of the Upper Salina section. Measured conductivities are given in the table below:

Borehole	Unit					
	G	F	E	D	C	B
U.W.O.	7.2	7.0	13.0		10.4	
Enniskillen #20-5	11.2	10.1	n		6.8	
Plympton #1-3	-	8.5	n		6.5	
MacGillivray #5-10	11.1	8.2	n			13.8
Midrim #3A	11.4	7.2	11.2		6.9	14.5
Argor #1	-	9.6	12.3		7.4	
Wayne Co.					6.6	13.3

Table VII-9 THERMAL CONDUCTIVITIES IN THE UPPER SALINA

The horizon as a whole is not suitable for heat flow determinations since it is so variable lithologically. However individual units such as the B-salt which show uniform thermal conductivity and great thicknesses in the north-west region are very suitable.

Salina conductivities for boreholes penetrating the complete formation are summarised in the table below; all individual units are considered to be suitably weighted except the B-salt which is weighted and non-weighted in the sets of results.

Borehole	Salina Conductivities		No. Samples	Thickness Section (m)
	non-weighted	weighted		
U.W.O.	10.8 ± 2.7	10.8	34	140
U.S. Steel	8.8 ± 2.9	8.8	13	107
Argor #1	10.0 ± 2.7	10.9	22	305
MacGillivray #5-19	8.8 ± 2.6	9.8	18	197

Table VII-10 AVERAGE CONDUCTIVITIES IN THE SALINA FORMATION

D <u>Bass Island</u>

The youngest Silurian beds are those of the Bass Island described by Sanford (1961) as a grey to brown cream coloured, finely granular dolomite overlying the Salina G unit gradationally. Caley (1943) has called these beds the Bertie-Akron series and describes them as a lithologically uniform series of fine crystalline to dense dolomites and dolomitic limestones. The southerly wells have traces of gypsum or anhydrite in the lower sections and those to the east, in Niagara, tend to have rather more shale partings present. In thickness the formation varies considerable from about 18 to 120 m. Locations of the cores are shown in FIG VII-8.

Continuous cores were available through the section in the same boreholes as those listed previously. The table below gives the mean conductivities.

Borehole	Conductivity
U.W.O.	10.0 ± 2.5
Enniskillen #20-5	9.9 ± 0.5
Plympton #1-3	11.5
MacGillivray #5-19	9.7
Midrim #3A	10.0 ± 1.0
Midrim #4	13.3 ± 1.4
Argor #1	12.4 ± 1.2
U.S. Steel	8.7 ± 1.6
Wayne Co.	11.1

Table VII-11 THERMAL CONDUCTIVITIES IN THE BASS ISLAND FORMATION

In general all of the conductivity measurements are high which is to be expected from the lithology. However the range of mean values is quite low, about 20%, if the most easterly result from the U.S. Steel hole is neglected. The lowest results are in the section furthest to the east and the highest in the section in the north-west region(region 5) excluding the Midrim #4 hole which contains much anhydrite. Once again the main problem in the section in an uncored borehole is in distinguishing between the limestones and dolomites.

VII-2-5 Devonian

A Bois Blanc

The lower Devonian strata are represented by a light grey glauconitic sandstone which is present only in the Niagara Peninsula. Above this there occurs a thick continuous carbonate sequence, the lowest of which is a grey limestone, sandy limestone or dolomite containing a lot of chert, called the Bois Blanc. Best (1953) divides the Bois Blanc into a lower Springvale member composed of coarse-grained friable fossiliferous calcareous quartz sandstone interbedded with glauconitic limestones and dolomites and an upper member of limestones, chery limestones and interbedded cherts. The lithology of this member varies considerably through the region, although no notable facies changes. Corals are more common in the Niagara area, whereas limestones are more shaley in the south-east. Since in many wells this series is not differentiated from the Detroit River, an analysis of the conductivity variations is left until after a discussion of the Detroit River Group which is part of the same carbonate sequence.

B <u>Detroit River</u>

The Bois Blanc grades upwards into the limestones and dolomites of the Detroit River. In the Michigan Basin section of Ontario the group consists of granular dolomites with interbeds of anhydrite and salt whereas in the Appalachian section it consists of finely crystalline dolomitic limestones and dolomites, becoming pure limestones at top. Best (1953) describes the upper part of the southern facies as consisting of a pure, evenly bedded, very finely grained, light brown and grey high-calcium limestone with abundant corals. Lower in the section the limestones become more granular with abundant bituminous laminae and styolites. This he considers to be typical of the Lucas. In the underlying Amherstberg formation the rocks are more dolomitic and silty with irregular bedding and many bituminous laminae. In the northern section the high calcium facies changes to a dolomite. These dolomites are soft, evenly bedded, uniform, light to medium-brown and fine grained. Occasionally the dolomite is finer grained, uniform and cream coloured containing lath-like crystals of calcite. Interbedded with the dolomites and rare limestones are thick biostromes of medium-grained medium-brown crystalline dolomite.

The broad north-west to south-east facies pattern is a change from the high calcium limestones of Ingersoll to the interbedded pure limestones and dolomites of St. Mary's to pure dolomites in the north accompanied by a thickening of the group from 9 m in South Walsingham to 107 m at Clinton and 76 m in Essex County. Landes (1951) reports a thickening into the Michigan Basin and a facies which is primarily interbedded dolomites, anhydrites and salt in the Lucas.

Since the formation lies at a shallow depth and is oil and gas producing, it has very frequently been cored. A large amount of this core was available for conductivity measurements and thus is the most thoroughly investigated for spatial variations. A table of the mean value is given below together with the thickness of the section and information as to whether or not the complete section was penetrated. The latter is important should there be any systematic variations with depth. As well as the borehole name, a number is given which identifies the site on FIG VII-9.

No. on FIG VII-	Borehole	Total	Conductivity Detroit R.	Bois Blanc	Thickness (m)
1	U.W.O.	7.7 ± 1.3	7.2 ± 1.2	8.5 ± 0.82	136 c
2	U.S. Steel	7.8 ± 2.0	-	7.8 ± 2.0	59 c
8	Argor #1	9.4 ± 1.6	9.4 ± 1.6	9.1 ± 0.5	126 c
9	MacGillivray #5-19	8.7 ± 0.9	8.6	11.3	160 c
10	Sarnia Disposal #3	9.2 ± 1.7			>97 in
6	Enniskillen #20-5	8.1 ± 0.9	undivided		114 c
11	Plympton #1-3	8.2	undivided		160 in
12	Corunna #18	7.9	undivided		143 in
13	Moore #10-9	7.6	undivided		>20 in
7	Enniskillen #18-2	7.0	undivided		>28 in
5	Dunwich	7.1	undivided		123 in
14	Sarnia 8R2	8.6 ± 1.5	undivided		>61 in
3	Midrim #3A	7.9 ± 1.4	undivided		131 c
4	Midrim #4	8.5 ± 1.1	undivided		>130 c
15	Wayne Co.	-	7.7	13.2	120 in

Table VII-12 THERMAL CONDUCTIVITIES IN THE DETROIT RIVER GROUP

The combined conductivities are lower in the sections where only the upper part of the Detroit River is penetrated and also to the east and south where the formation tends to be less dolomitic. The mean unweighted conductivity for all samples is 8.3 ± 1.2, which probably would be higher if all the measurements were on complete sections. It would appear that the conductivity of the Bois Blanc is more variable between boreholes than that of the Detroit River. However the Bois Blanc conductivities in given boreholes, such as Argor#1 and U.W.O., are less variable.

C Delaware

Above a thin sandstone horizon, the Columbus, or above the Detroit River but lying unconformably on it are the limestones of the Delaware. It overlaps both the Columbus and Detroit River, and in the east, e.g., the U.S. Steel borehole, rests directly on the Bois Blanc. The Delaware commonly consists of a buff clastic limestone grading upward into a finely crystalline limestone, with interbedded black shale at its eastern margin (Sanford, 1961). Best (1953) divides the formation into two members; a lower massive limestone occurring only to the area north of St. Marys and a gradational contact to an upper member composed of a partly argillaceous fine-grained sublithographic limestone. Thicknesses range up to 60 metres. In the Michigan section the equivalence between the Delaware and the Dundee seems somewhat confused. However the conductivities measured in the Dundee of Wayne Co., Michigan and the Delaware of southern Ontario are similar.

Borehole	Conductivity	No. of Samples
U.W.O.	7.3	2
MacGillivray #5-19	6.4	2
Enniskillen #20-5	7.0	5
Plympton #1-3	7.7	3
Argor #1	7.4	4
Midrim #3A	7.1	3
Midrim #4	7.2	3
Sarnia #8R2	7.3	1
Wayne Co.	7.1	1

Table VII-13 THERMAL CONDUCTIVITIES IN THE DELAWARE FORMATION

D Hamilton

Sanford (1961) describes this formation as composed of grey calcareous shales and shaly limestones and correlates it with the Traverse group in Michigan. Caley (1943) has described it in outcrop in some detail. It occurrence is limited to the north-west section of the Peninsula where it ranges up to 60 m in thickness.

Samples were available from only two boreholes for which the conductivities are given below:-

Borehole	Conductivity	No. Samples
Argor #1	6.2	3
Midrim #3A	5.1	1

Table VII-14 THERMAL CONDUCTIVITIES IN THE HAMILTON FORMATION

VII-3 THERMAL CONDUCTIVITY, DENSITY AND POROSITY OF SEDIMENTARY ROCKS

Several formations of different composition are examined for relations between the three parameters measured; k is the thermal conductivity, ρ the density, σ the porosity and z the depth.

VII-3-1 Trenton-Black River Group

Borehole	Correlation k with z	Correlation k with ρ	Correlation k with σ	Correlation ρ with σ	Geological Group
Cols. 2	-51.2	-42.6	- 1.6	23.6	Trenton
Cols. 6	13.4	78.2	35.0	26.5	Trenton
Cols. 6	5.3	78.3	-32.1	-45.5	Black River
Cols. 7	-51.9	88.3	14.0	9.1	Trenton
All	- 4.2	63.0	- 0.4	3.3	Trenton
All	67.9	81.2	-32.7	-58.4	Black River

Table VII-15 CORRELATION OF PHYSICAL PARAMETERS MEASURED IN THE TRENTON- BLACK RIVER GROUP

The preceeding table shows correlation coefficients (x100) calculated for the respective linear regressions. While there must be a relation between all of the physical parameters of the rock including grain and pore sizes and sorting and cementation factors as well as the parameters of density, porosity and thermal conductivity, it is apparent that there is little or no linear correlation between thermal conductivity and porosity whereas the correlation between conductivity and density indicates some association. This latter suggests that the conductivity variation in the section correlates with rock type which is further borne out by the Cols. 2 correlation which is the poorest. It is this hole which has the least change in rock type, and thus the smallest density contrasts. The lithology of the groups and the individual boreholes is discussed in greater detail in VII-2-3A. In the Colchester boreholes the cored material from the Trenton is all Sherman Falls member and all of the Black River is of the Gull River member. Since a relationship seems to exist between the thermal conductivity and the density, it may be possible to relate this to a simple model expressing the variation in terms of a mixture of two components. Assuming that the two components are a limestone and a dolomite mixed randomly in the matrix, the total conductivity of the mixture k_T will be given, using Woodside & Mesmer (1961), as:-

$$k_T = k_L^{\alpha} \cdot k_D^{1-\alpha} \qquad \ldots\ldots \text{v)}$$

where k_L and k_D are the thermal conductivities of the limestone and dolomite respectively and α is the proportion of limestone by volume and k_D is less than 10 x k_L.

The density ρ_T is given by:-

$$\rho_T = \alpha\rho_L + (1-\alpha)\rho_D \quad . \quad . \quad . \quad . \quad . \; . . \; \text{vi)}$$

where ρ_L and ρ_D are the densities of the limestone and dolomite respectively.

The thermal conductivity of the limestone is taken as 6.3 and the density as 2.66, that of the Sherman Falls in the U. S. Steel hole which is away from any dolomitisation effects. The conductivity of the dolomite is taken as 11.8, the highest value encountered with a density of 2.83. Similarly the conductivity of the Gull River is taken as 6.7, with a density of 2.68 and again the same values used for the dolomite. FIG VII-10 shows all the Sherman Falls and Gull River conductivities plotted on a graph of thermal conductivity versus density. Also plotted is the predicted variation. Comparison of the results suggest that the pure dolomite probably has a slightly higher conductivity and density, but that the assumption of a two component model is not a bad one. Scatter will be caused by porosity and shaliness of the samples and also perhaps by the size and orientation of crystals in the samples. This crude approximation however does verify further the statement made in VII-2-3A that a knowledge of the proportions of calcite and dolomite present in the vicinity of a borehole enables a reasonable prediction of thermal conductivity for a Trenton-Black River section.

VII-3-2 Queenston

The formation as it appears in the U.W.O. borehole consists of a fairly uniform series of reddish-brown slightly calcareous cryptocrystalline shales. Interbedded with these and in some instances forming up to 35% of the core are greyish-green shales which are usually rather more calcareous than the former and in places become almost a shaley grey limestone. Over very short intervals anhydrite occasionally makes up 3 to 5% of the core. FIG VII-11 shows a plot of density versus conductivity indicating a density increase with conductivity, and a porosity decrease. There is however virtually no dependence of porosity on the the density. The respective correlation coefficients are 40, -52 and -18. As a further attempt to explain a possible pattern in the conductivity variations a simple chemical analysis was performed on the samples. The samples were dissolved in dilute hydrochloric acid and the residues weighed. Residues of the red shales cluster between 60 and 75% with a conductivity range of 4.2 to 5.4 mcal/cmsec^{o}C. The few samples of reddish appearance which had higher conductivities proved to have high residues and also higher densities and lower porosities. Samples of green shale tended to have lower residues and higher conductivities than the red shales. Thus it would seem that in a very variable shale formation where sampling becomes a problem, some useful information regarding weighting of the results to give a mean formation conductivity might be obtained by simple chemical analysis. It is however doubtful, in the light of these tests, whether this type of chemical analysis could be used to predict thermal conductivity directly, although it might be used to distinguish types of shales.

VII-3-3 Guelph-Lockport

This formation in the U.W.O. borehole is composed of very pure dolomites varying from buff to tan in colour and from fine to medium crystalline in structure. A section at the top contains scattered corals but most of the porosity throughout is contained in pin-point vugs. The conductivities are very uniform ranging from 10.5 to 13.0, with the highest conductivities occurring at the top of the formation and at the base of the Lockport. It is very strange to note that in the section at the top of the Guelph both the conductivities and porosities are high while the densities are the lowest in the section, whereas in the Gasport unit of the Lockport the conductivities are high, the densities high and the porosities low.

Correlations of the various physical parameters yield coefficients of -58, 42 and -80 which would suggest that much of the density variation is due tc porosity changes. However the variation of conductivity is only partially explained by density changes. It has often been questioned in the past as to how much of the dolomite in a stratigraphic sequence might be syngenetic and how much is replacement of limestone (eg, Ingerson, 1962). This is of special significance in southern Ontario where most of the reservoirs are in carbonates. Usually in these reservoirs dolomite is the larger producer of petroleum because of its higher porosity. Hohlt (1948) showed that whereas calcite crystals in a limestone showed a pronounced tendency to orient their c-axis in the bedding plane, dolomite crystals are always oriented at random. Now if the replaced calcite retains the original orientation of the limestone and if the conductivity of the dolomite is greater along the c-axis then the replaced dolomites should have a slightly lower conductivity than the primary ones. To prove this would require a good deal of careful thin section work along with conductivity measurements. It is however a possible explanation of the noted discrepancies.

VII-3-4 Detroit River

In the U.W.O. borehole this formation consists of a tan to light grey-brown limestone mostly very calcitic and containing abundant corals. Scattered beds of dolomite occur in some sections. This section varies quite considerably in porosity and therefore the data is analysed to attempt to show that the conductivity is dependent on porosity and to find the nature of the dependence. One simple approach to this is to take one of the simple models mentioned later in the section and to see if correcting the conductivities and densities to the matrix values, ie., removing the porosity, causes the parameters to cluster better. FIG VII-12 shows histograms of the parameters before and after correction using equation i) and the geometric model (equation iv)).

The dolomitic zones in the formation are accompanied by increased conductivity, density and porosity, whereas the shalier sections are accompanied by a decrease in all of these parameters.

Comparison of the correlation coefficients yields 60 between conductivity and density, -29 between conductivity and porosity and -73 between density and porosity. The most significant correlation is the decrease of density with increasing porosity, whereas the thermal conductivity is less sensitive to variations in density and porosity.

VII-3-5 Intercomparisons

The last part of this section is to intercompare the variations of the measured properties in all three formations as follows.

Fitting a linear relationship through each pair of parameters shows the dependence of each. As expected, density decreases with porosity increase. However if the samples are homogeneous and the only difference between samples is the pore space capable of holding water, then the relationship between density and porosity should be given by

$$\rho = \sigma\rho_w + (1 - \sigma)\rho_m \qquad \text{i)}$$

where σ is the volume porosity and ρ_m and ρ_w are the densities of the rock matrix and of water respectively. The values calculated on this basis are shown in FIG VII-13 in comparison with the best least squares fits through the data. It is apparent that the Queenston and Detroit River data brackets the line fairly well, but the Guelph-Lockport results are far from the expected line. Probably the reason for this is the small range of porosities encountered as most of the data lies between 1.2 and 4%. However as previously mentioned the correlation between density and porosity is good which leaves the question open. The Trenton-Black River carbonates have a poor correlation of density with porosity reflecting the large compositional changes which swamp the porosity effect. A further difficulty in understanding the Guelph results is in the relation between the thermal conductivity and the density, in which the Guelph conductivities would appear to increase with decreasing density whereas the Queenston, Detroit River and Trenton-Black River conductivities increase with a density increase. It is very common practise to attempt to model thermal conductivity with porosity since these parameters are of interest to the oil industry. The simplest models and the most used ones are the 'series' and 'parallel' arrangements of Birch et al (1940) and the 'geometric' arrangement of Woodside and Mesmer (1961):-

Series: $$\frac{1}{k_r} = \frac{\sigma}{k_w} + \frac{1 - \sigma}{k_m} \qquad \text{v)}$$

Parallel: $$k_r = \sigma k_w + (1 - \sigma)k_m \qquad \text{vi)}$$

Geometric $$k_r = k_w^{\sigma} \cdot k_m^{1-\sigma} \qquad \text{vii)}$$

where k_r represents the measured conductivity, k_w and k_m that of water and the rock matrix respectively and σ the porosity of the rock. In each of these cases the value of k_m is found by extrapolating the least squares regression of conductivity on porosity to zero porosity. The geometric model approximates Maxwell's (1892) two- and three-phase models, provided that the thermal conductivities of the phases do not differ by an order of magnitude or more.

The expected variation of conductivity with porosity is shown in FIG VII-13 for each of these models and compared with the actual relationship. It is apparent that each formation fits a different model best; the Guelph and Trenton a parallel arrangement of pores, the Detroit River, Queenston and Black River a series arrangement. The significant correlations were between density and thermal conductivity for the Trenton & Black River, in which presumably changes in rock type are predominant. In the Detroit River there is a good correlation between density and porosity and between conductivity and density which suggests that both factors are important. The Guelph also shows a good correlation between density and porosity and a lesser one between thermal conductivity and density. Correlation between any set of parameters in the Queenston is very poor. The only coefficient greater than 50 is a negative correlation between thermal conductivity and porosity.

These correlation comparisons suggest that while it is a very uncertain procedure to attempt to predict thermal conductivities on the basis of lithology, density and porosity logs, such as one might have from conventional oil-wells, the use of such logs together with thermal conductivity, density and porosity measurements on a few pieces of core may provide an adequate solution to the problem of obtaining conductivities in sedimentary basins. The possible value of well logs is considered again briefly in VII-5.

VII-4 THERMAL CONDUCTIVITY AND ROCK TYPES

In the following tables, the mean borehole conductivities for various formations are compared across the region:-

Formation	Rock Type	Conductivity mcals/cm.sec.°C			Std Dev	No of Holes	Extent (km)
		Max	Min	Mean			
Hamilton	Shaley Limestones	6.2	5.1	5.7	-	2	80
Delaware	Limestone	7.4	6.4	7.2	0.3	9	200
Detroit River	Limestone	9.4	7.2	9.0	-	3	100
Bois Blanc	Cherty limestone	9.1	7.8	8.5	-	3	100
Total undivided	Limestone, dolomite, chert	9.4	7.0	8.1	0.7	14	200
Bass Island	Dolomite	13.3	8.7	10.7	1.4	9	200
Salina total[1]	Mixed	10.8	8.8	9.6	0.8	4	200
Guelph	Dolomites	11.8	8.1	10.7	1.0	8	200
Clinton-Cataract	Mixed	9.3	4.9	7.3	-	3	80
Queenston	Shales	-	-	5.3	-	1	-
Meaford	Siltstone and shaley limestone	-	-	6.3	-	1	-
Collingwood	Shale	-	-	4.4	-	1	-
Trenton-Black River	Limestone, dolomite in west, shalier in east	6.7	5.0	6.1	0.6	6	600
Cambrian	Sandstones	12.1	11.0	11.3	-	3	300
Precambrian	Gneisses	7.5	4.4	7.1 (unweathered)	0.1	11	200

Table VII-16 THERMAL CONDUCTIVITIES IN SOUTHWESTERN ONTARIO - INTERBOREHOLE COMPARISONS

[1]Subdivisions in following table

Salina Subdivision	Thermal Conductivities mcals/cm.sec.^{o}C		No of Holes
	Mean	Std Dev	
G	10.2	1.8	4
F	8.4	1.1	6
E	12.2	-	3
C	7.4	1.4	6
B	13.9	-	3
A2	8.9	1.6	8
A1	9.1	1.4	7

Table VII-17 THERMAL CONDUCTIVITIES IN SUBDIVISIONS OF THE SALINA FORMATION

The formations with the smallest regional variation of thermal conductivity, and hence the most suitable for use in heat flow determinations are:- the Precambrian gneisses, the mixed limestones, shales and dolomites of the Trenton-Black River, the dolomites of the Guelph, the mixed rocks of the Salina(if weighting factors for each horizon can be deduced), the limestones and cherty limestones of the Detroit River and Bois Blanc, and the limestones of the Deleware. In each of these cases, although the downhole variability of conductivity in a single hole may be considerably more, the mean formation conductivity has a standard deviation of 10% or less. In boreholes with no available core but in which temperature measurements could be made, the formations mentioned above should give the most reliable heat flow determinations. Any major anomalies in the heat flow field such as the Guelph values at Gosfield and London are easily revealed. As examples of the greater overall variability of conductivity, several formations have been treated by taking all of the individual conductivities and calculating a mean.

Formation	Thermal conductivity mcals/cm.sec.°C				No. of Samples
	Max	Min	Mean	Std.Dev	
Trenton	11.8	5.29	7.1	1.3	53
Black River	11.0	5.78	8.5	1.7	17

Table VII-18 VARIATION OF THERMAL CONDUCTIVITY IN THE TRENTON-BLACK RIVER GROUP IN THE COLCHESTER OIL FIELD

All of the samples in the table above came from holes in the Colchester oil-field and thus are from within a few miles of each other.

Attempting to give specific conductivities for rock-types rather than formations is difficult because most rocks or formations are mixtures of lithologies. However a few values can be given which should be treated with suspicion. For example the Delaware and some holes through the Trenton contained fairly good limestones of conductivity between 6.4 and 7.2. Similarly in the Guelph pure dolomites were encountered with conductivities between 11.0 and 12.4. In the Salina, salt and anhydrite were encountered with respective conductivities in the ranges 13.3 to 14.5 and 13.0 to 14.8. Sandstones and shales are more variable in conductivity.

VII-5 THERMAL CONDUCTIVITY AND WELL LOGS

In order to be complete the potential use of improved well logs indicating porosity changes or proportional changes in lithological components should be mentioned. Such logs can be used to predict changes in thermal conductivity of a formation in a given hole, within limitations such as those discussed in VII-3-5. When Joyner (1960) attempted to use logs for such purposes they were rather primitive. Today, for example, a limestone, dolomite mixture can be effectively solved using a Sidewall Neutron Log and a Density Log. If a Velocity Log is added than a third component such as evaporites can be sorted out. No suitably complete logs were available in southern Ontario where the standard procedure is to run only a Gamma and Neutron Log. However the Gamma Log proved to be a useful qualitative tool for shaliness particularly in the Trenton-Black River.

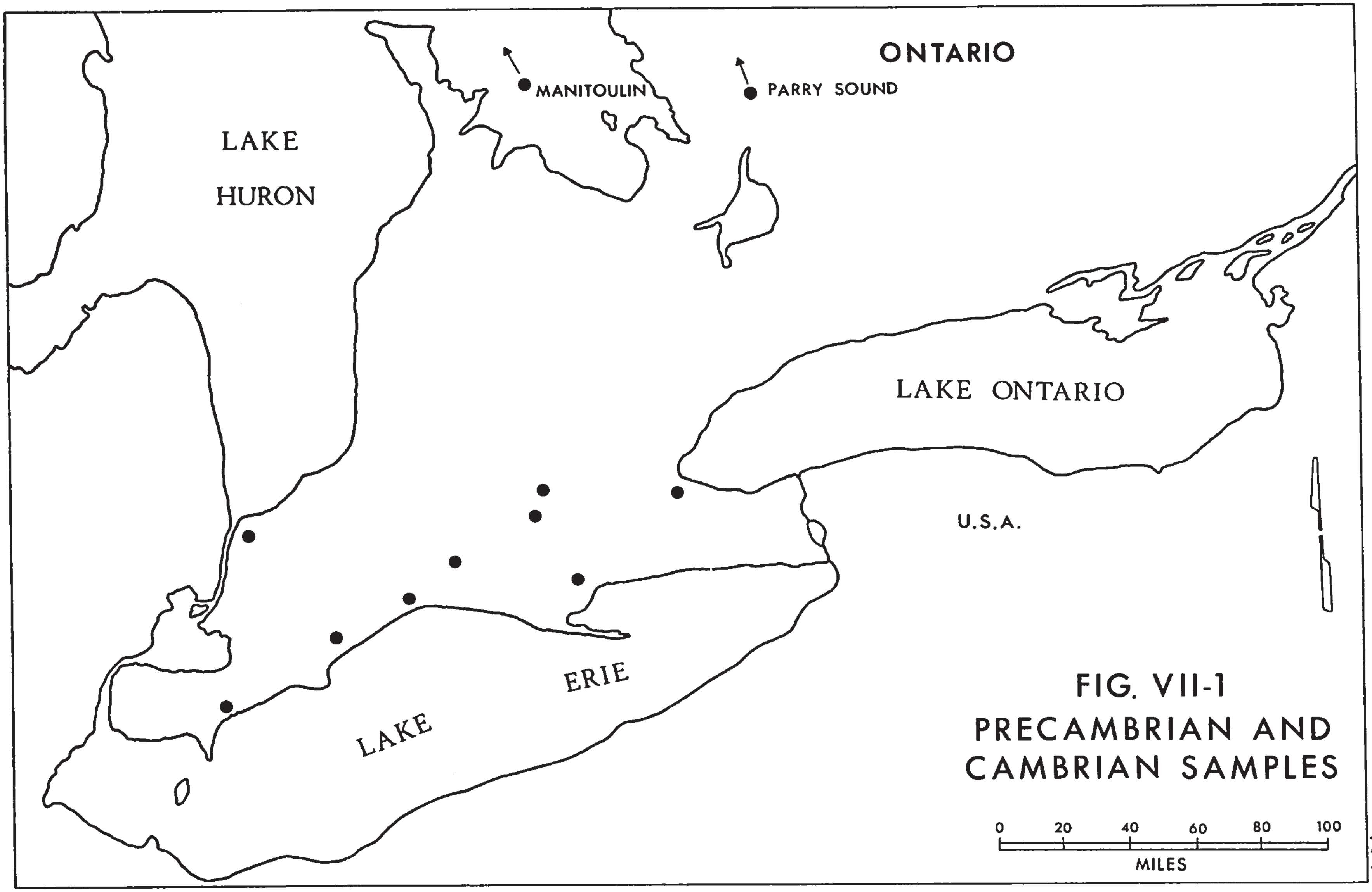


FIG. VII-1
PRECAMBRIAN AND
CAMBRIAN SAMPLES

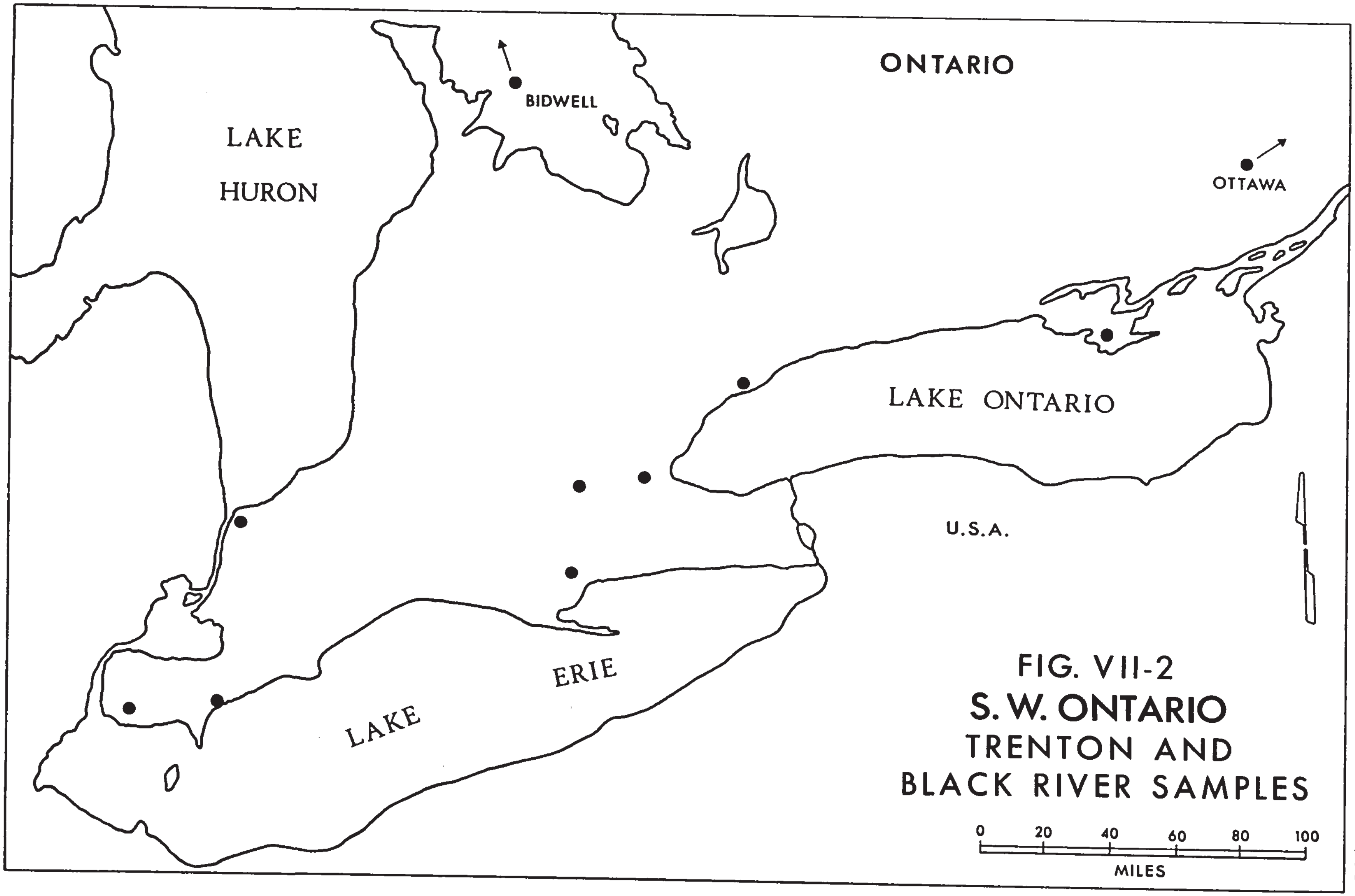


FIG. VII-2
S. W. ONTARIO
TRENTON AND
BLACK RIVER SAMPLES

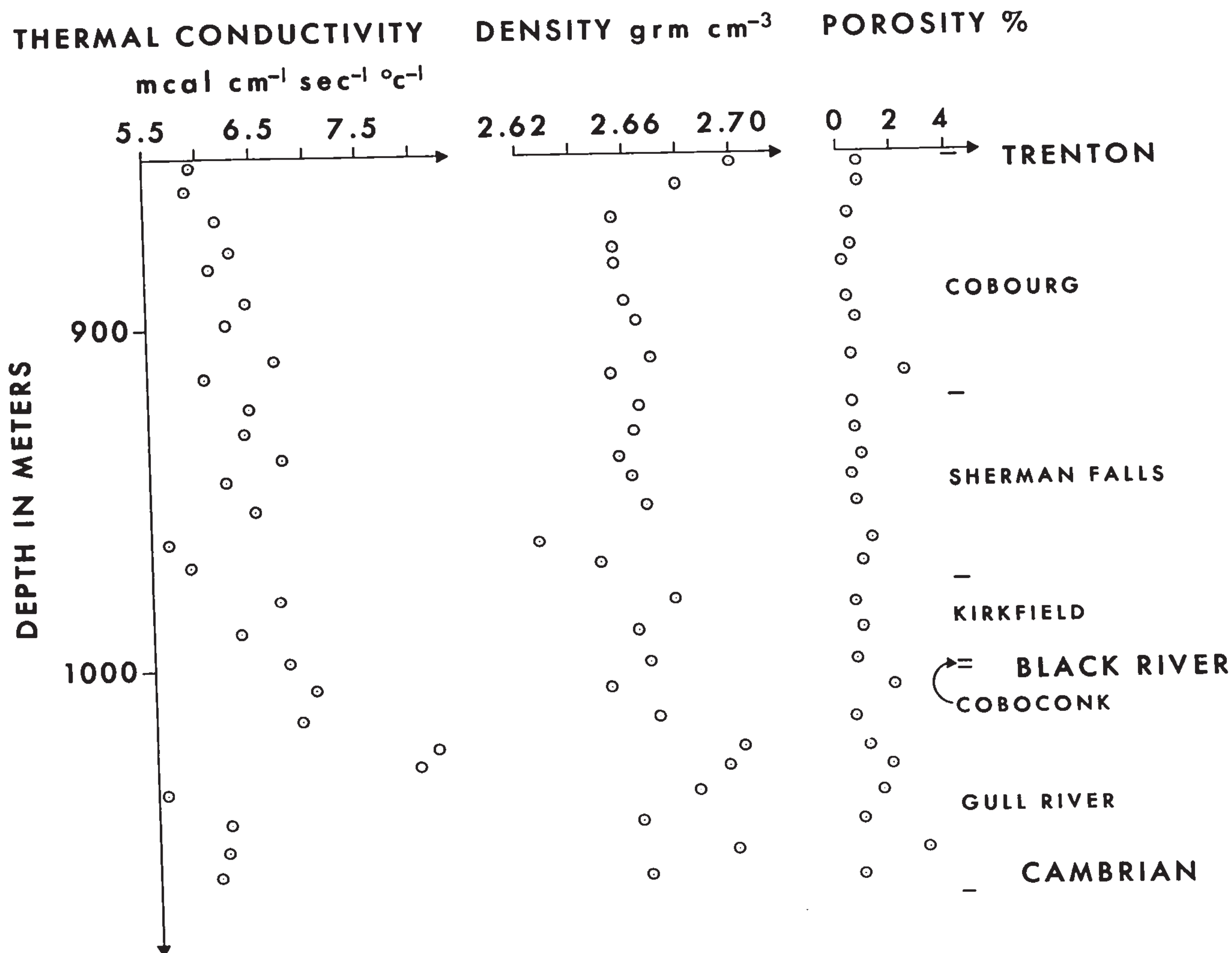


FIG. VII-3
TRENTON-BLACK RIVER CONDUCTIVITIES IN U.S. STEEL HOLE

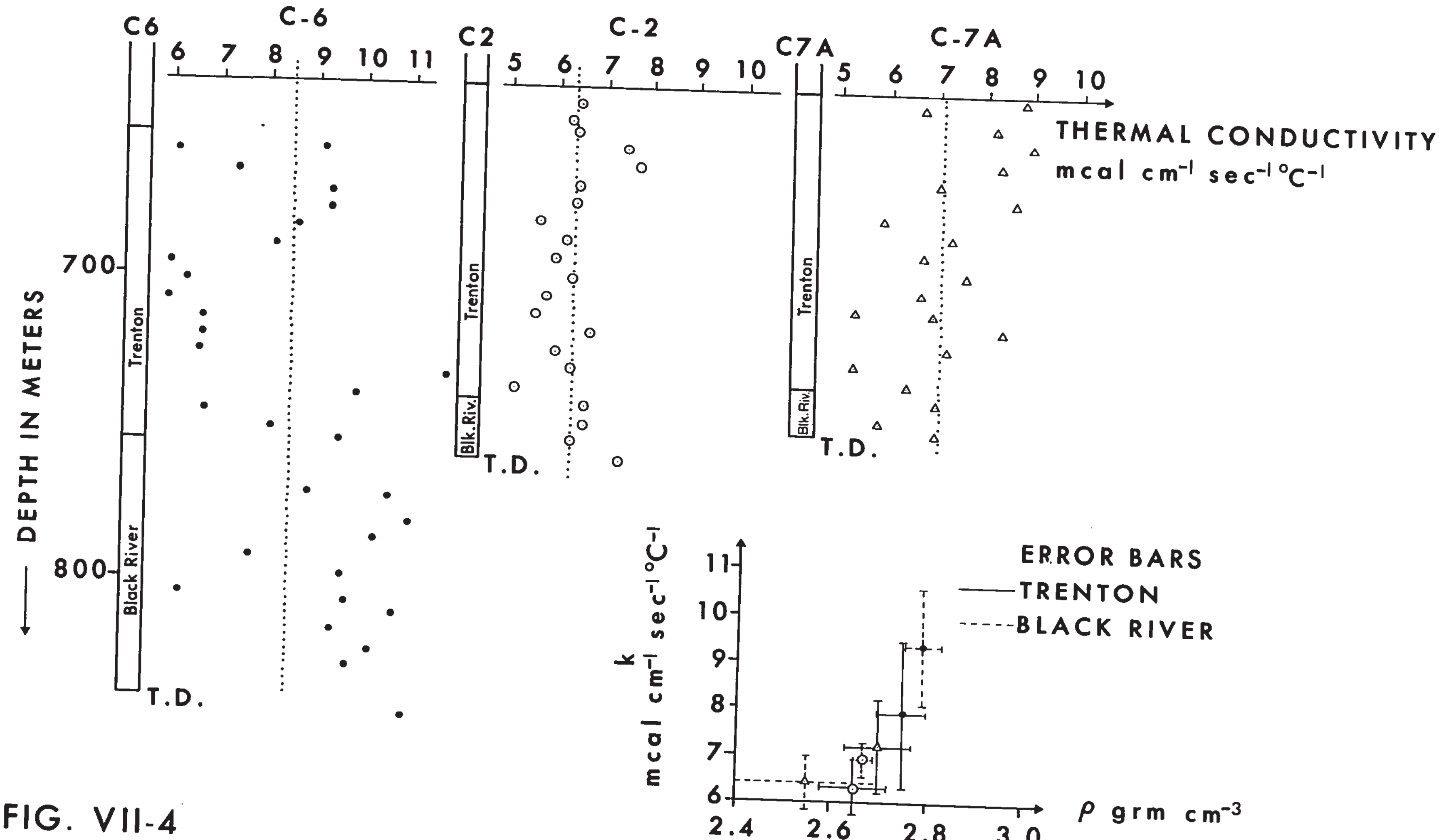
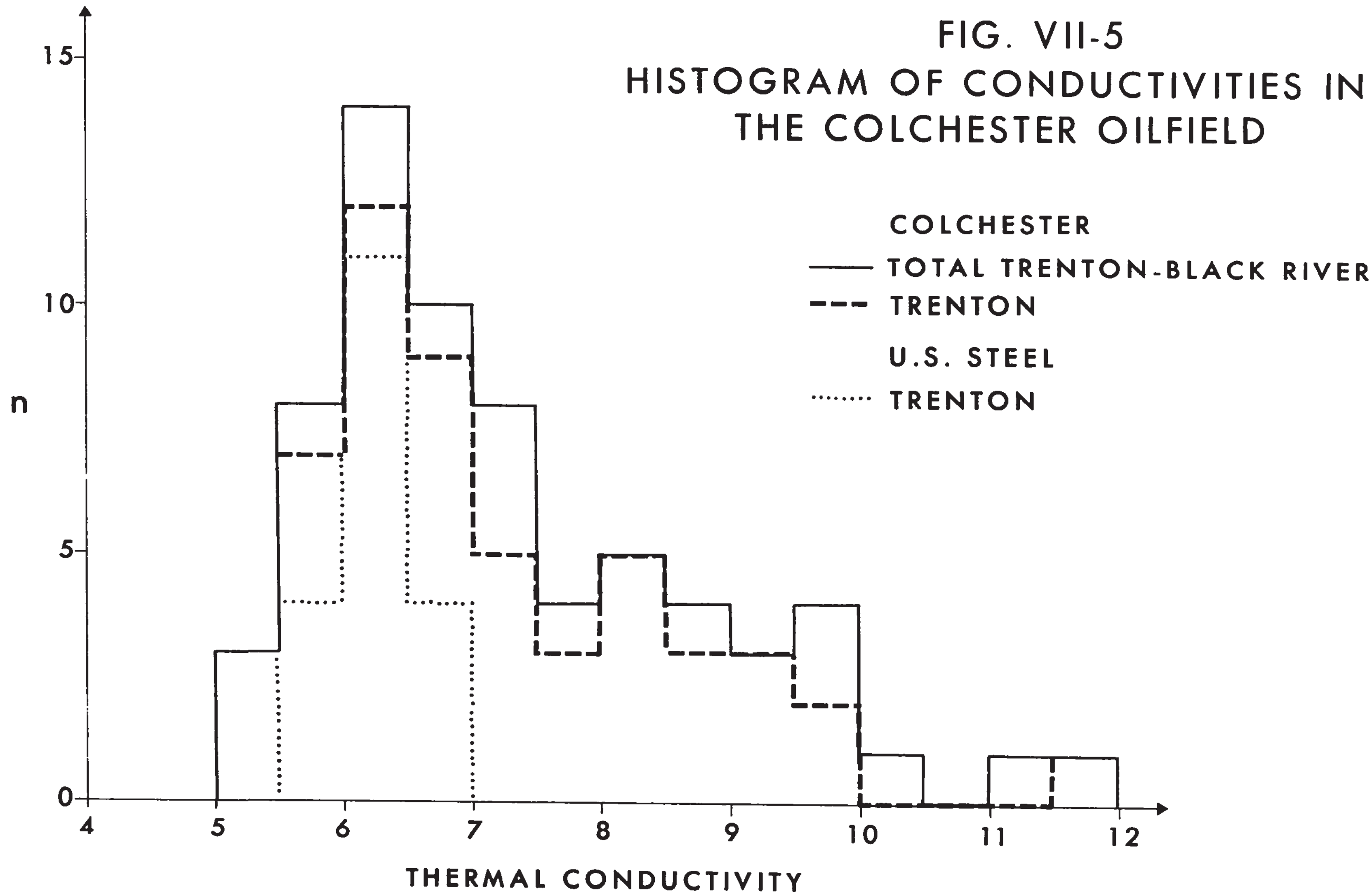


FIG. VII-4
TRENTON - BLACK RIVER CONDUCTIVITIES
IN THE COLCHESTER OILFIELD

FIG. VII-5
HISTOGRAM OF CONDUCTIVITIES IN THE COLCHESTER OILFIELD



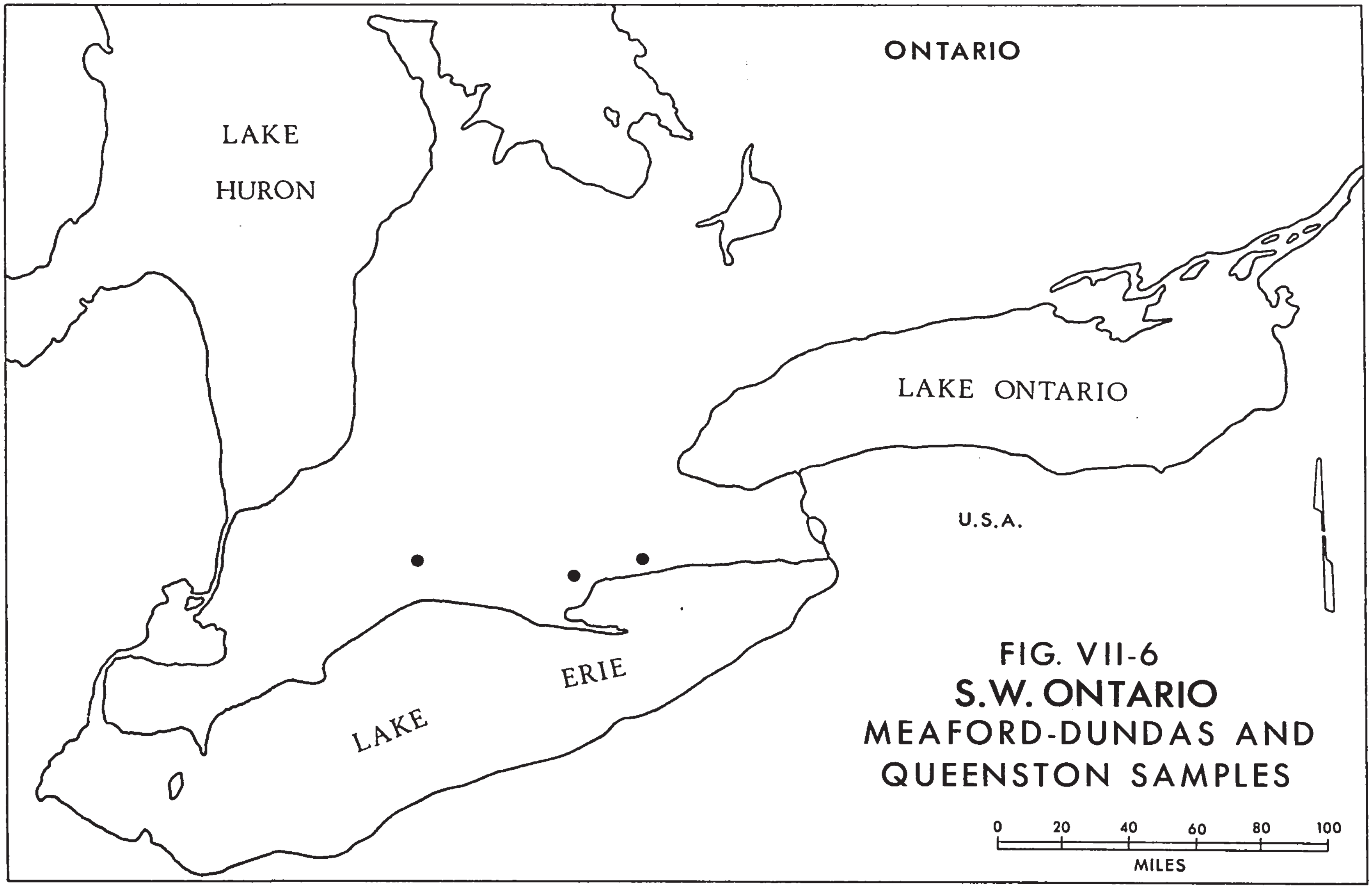


FIG. VII-6
S.W. ONTARIO
MEAFORD-DUNDAS AND
QUEENSTON SAMPLES

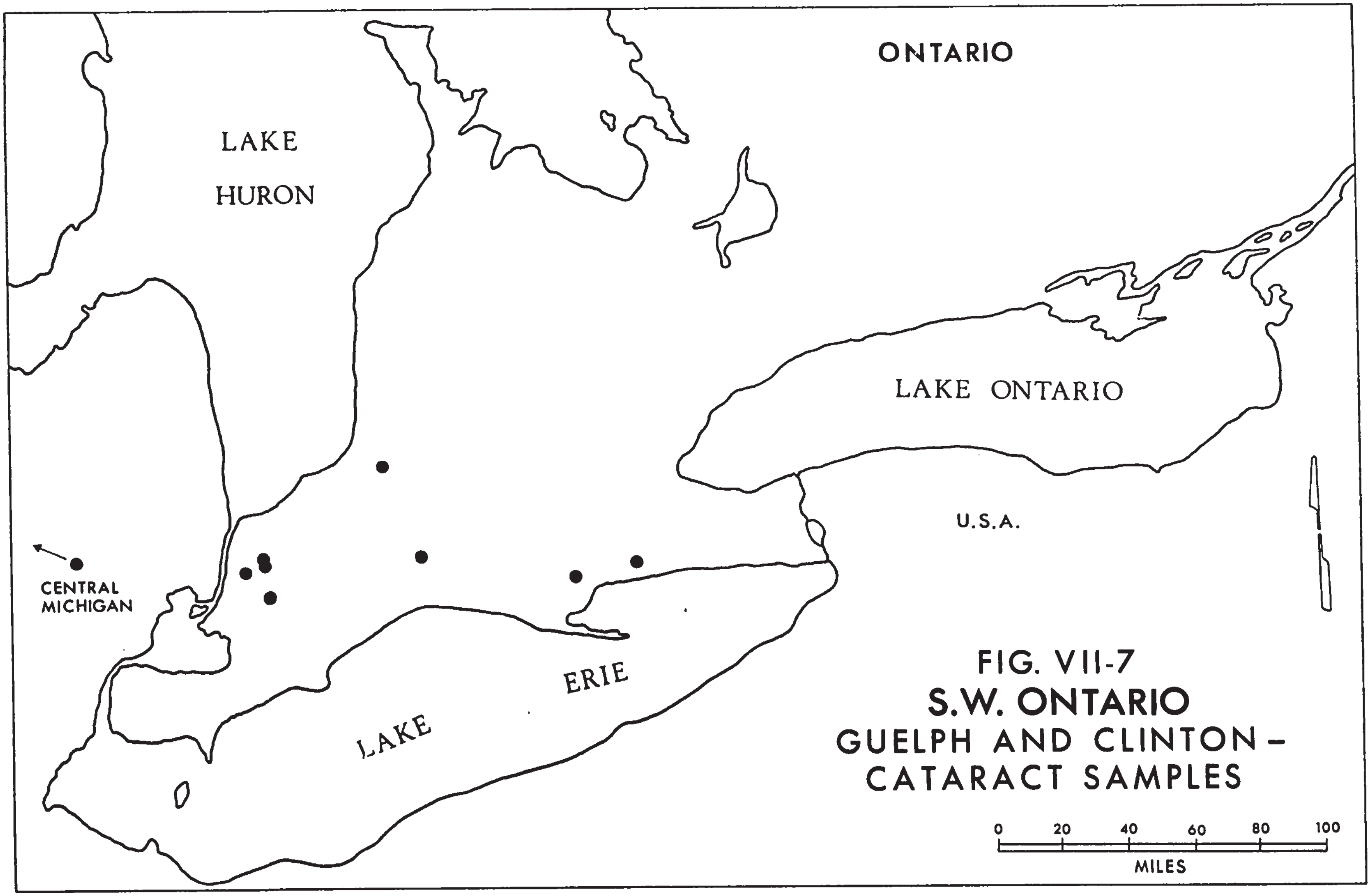


FIG. VII-7
S.W. ONTARIO
GUELPH AND CLINTON – CATARACT SAMPLES

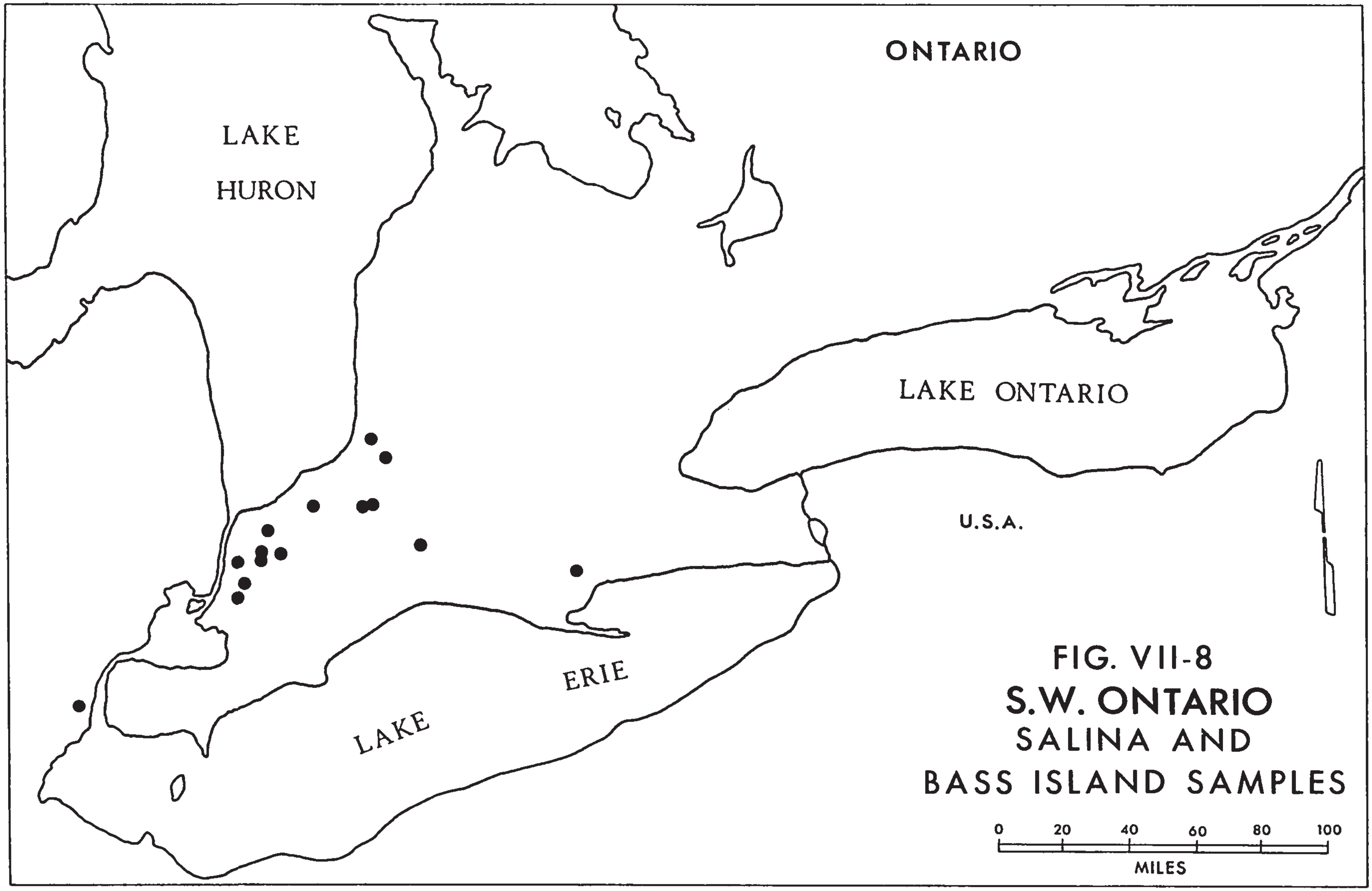


FIG. VII-8
S.W. ONTARIO
SALINA AND
BASS ISLAND SAMPLES

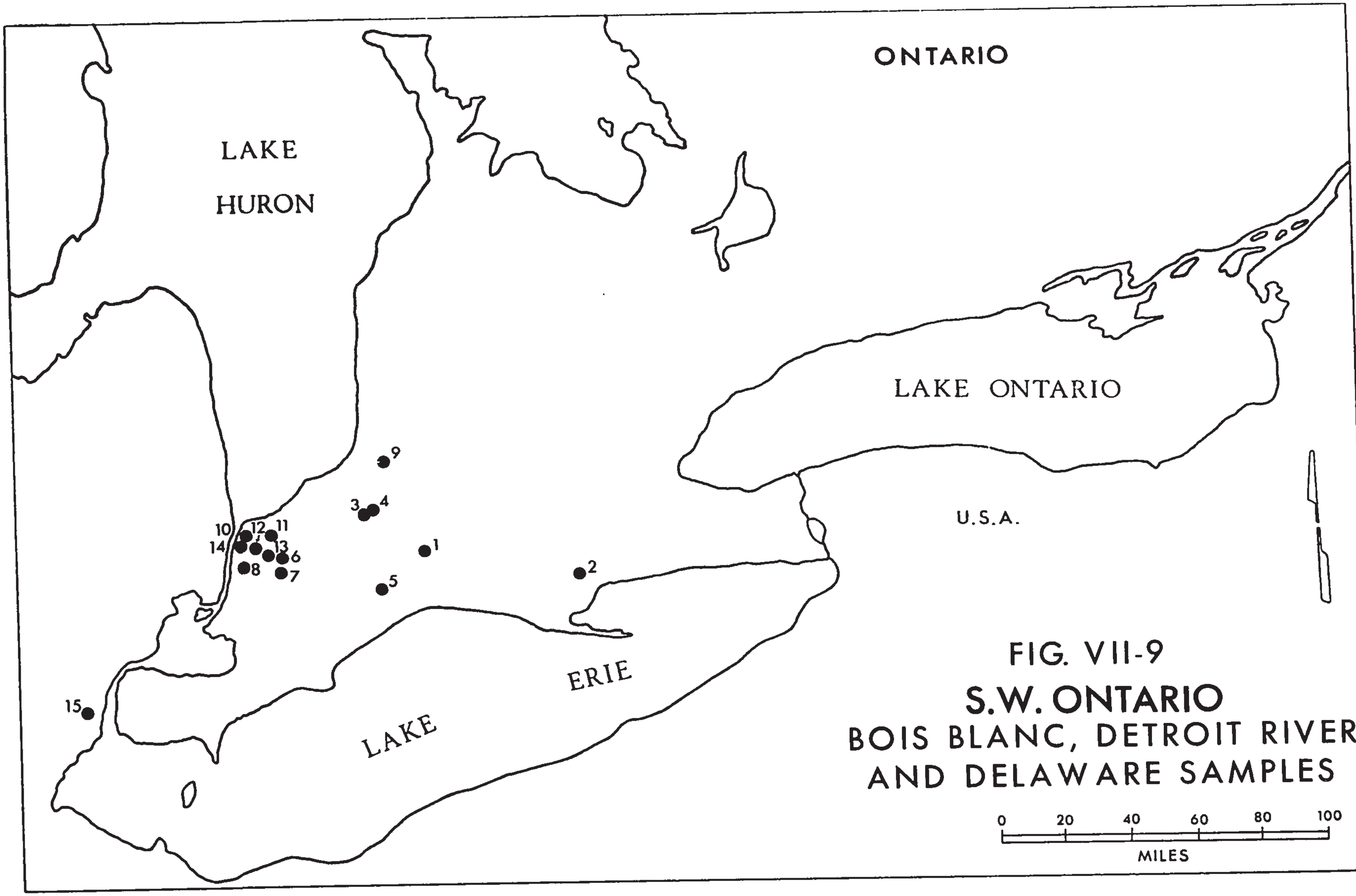


FIG. VII-9
S.W. ONTARIO
BOIS BLANC, DETROIT RIVER AND DELAWARE SAMPLES

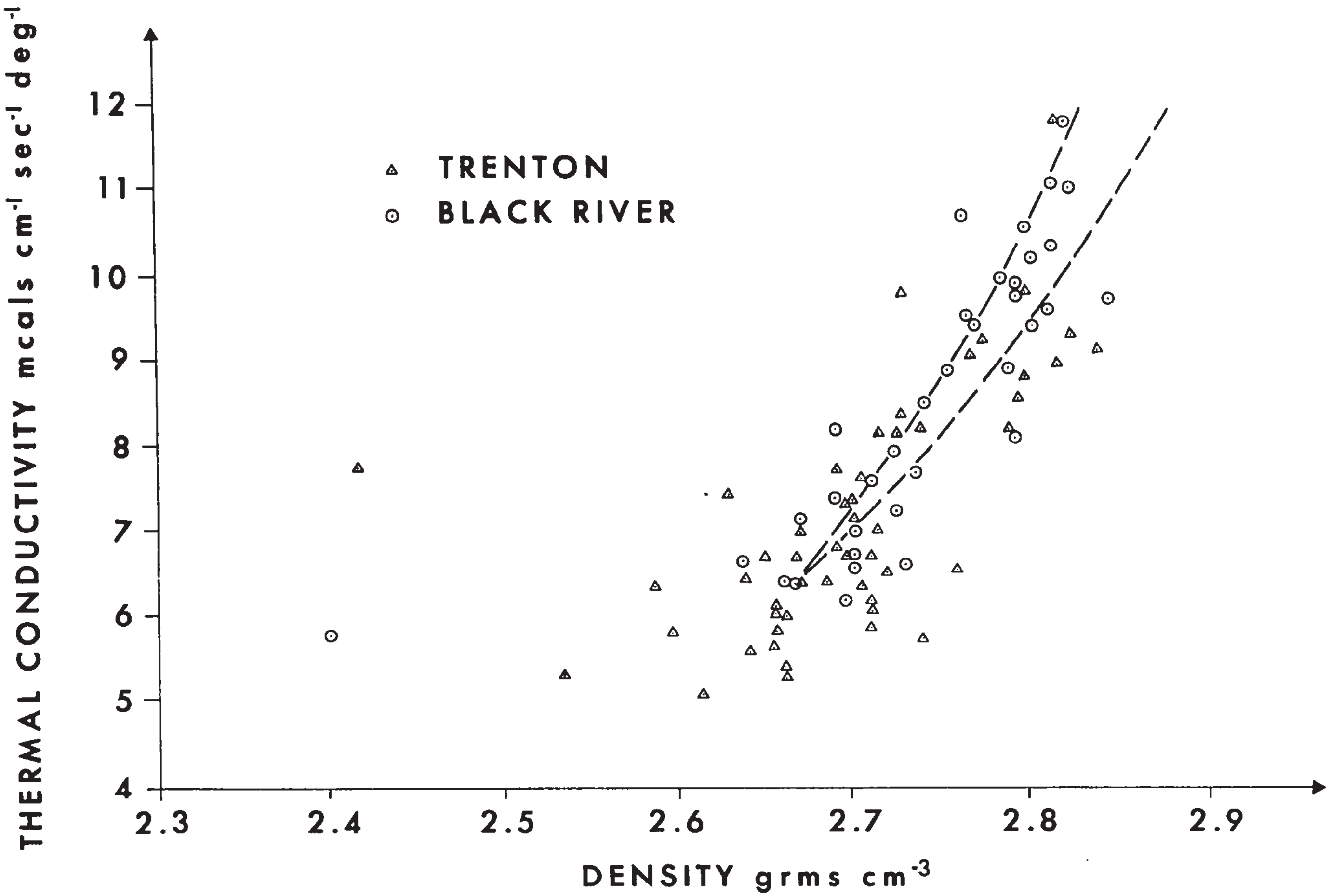


FIG.VII-10 VARIATION OF THERMAL CONDUCTIVITY WITH DENSITY : -TRENTON-BLACK RIVER

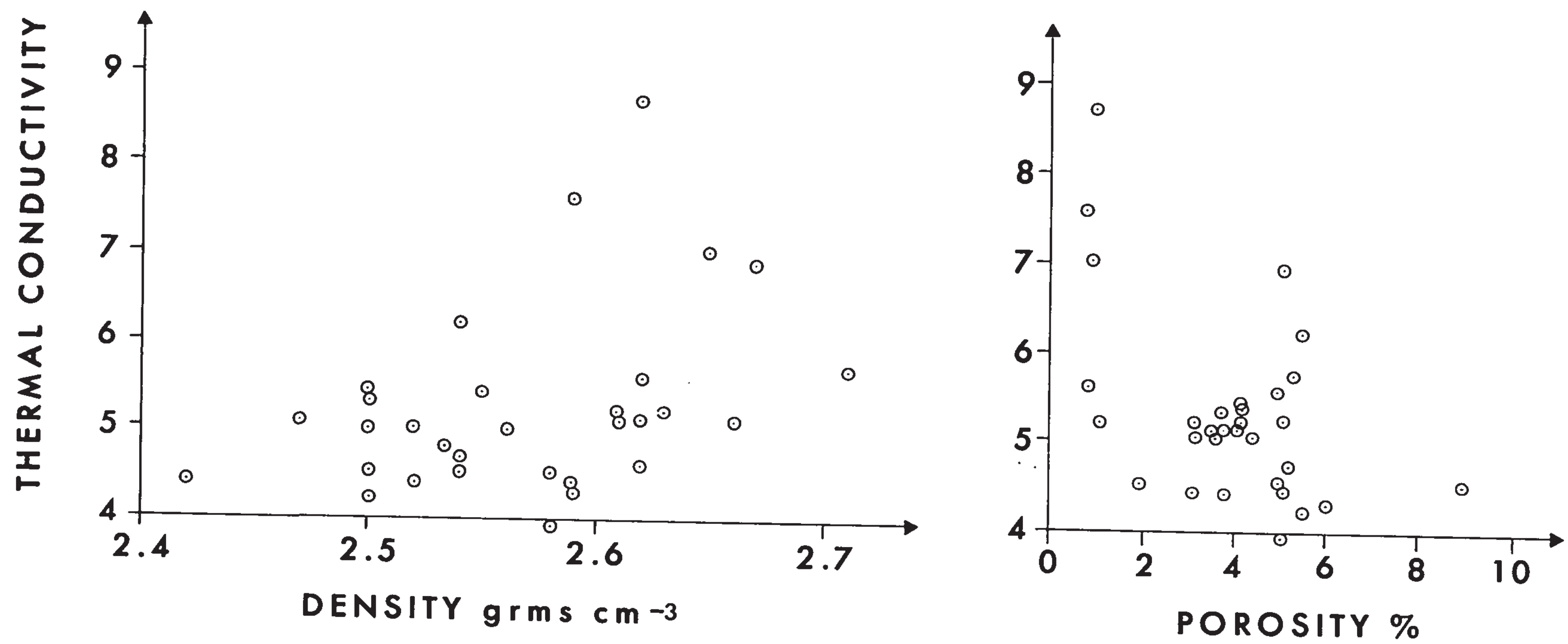
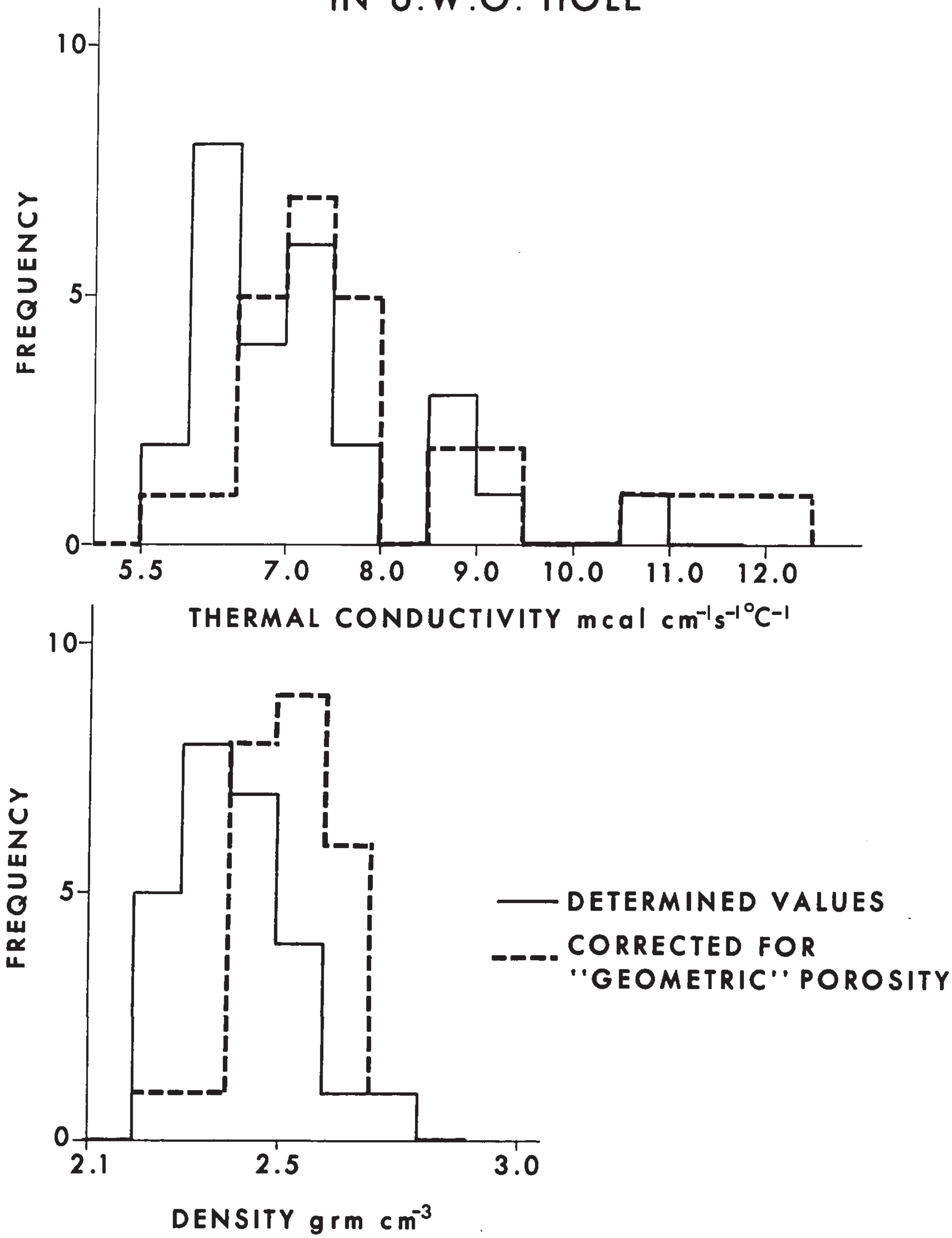


FIG. VII-11 VARIATION OF THERMAL CONDUCTIVITY WITH DENSITY & POROSITY :- QUEENSTON

FIG. VII-12
HISTOGRAMS OF THE DETROIT RIVER SAMPLES
IN U.W.O. HOLE



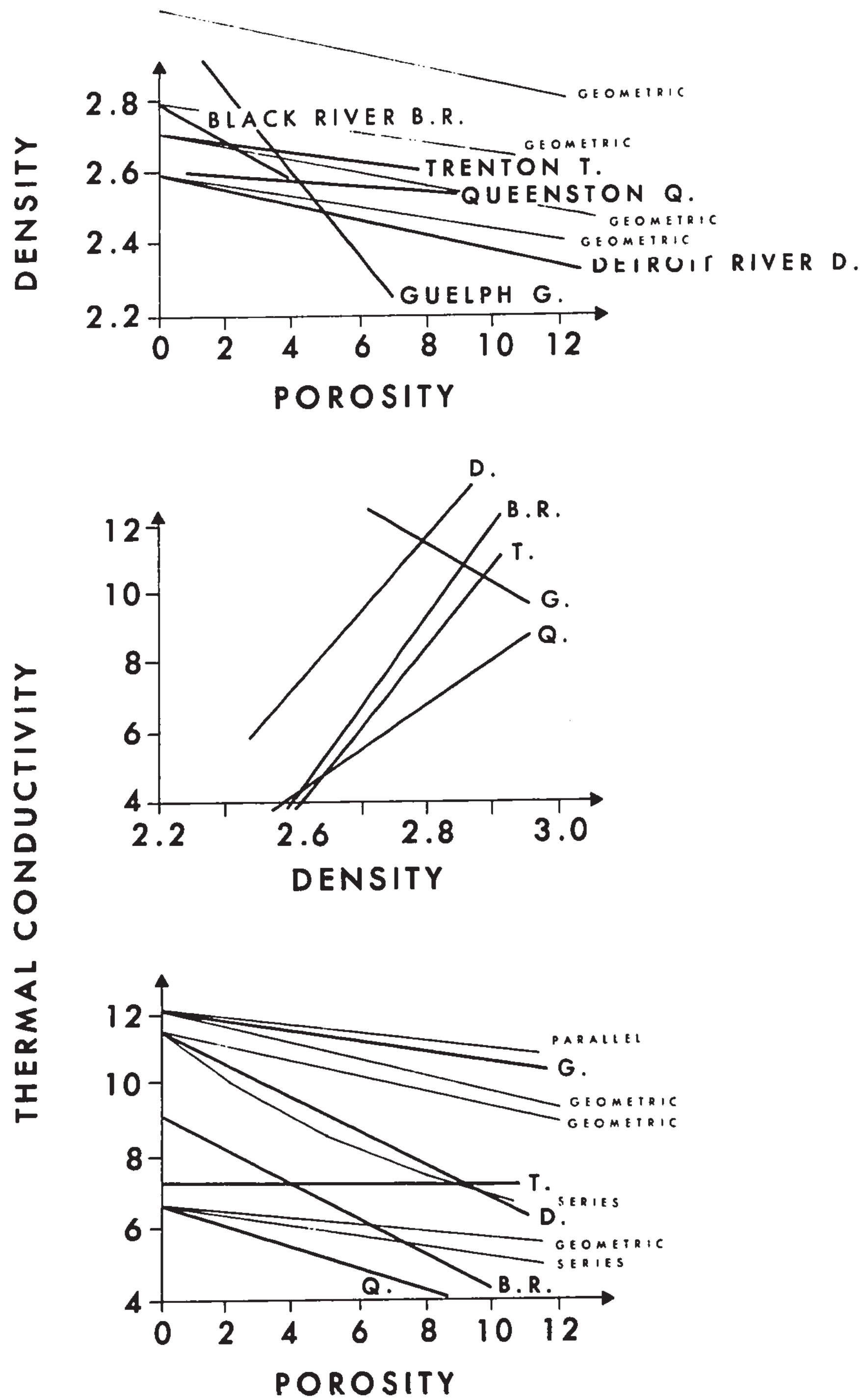


FIG. VII-13 LINEAR RELATIONSHIPS CONNECTING THERMAL CONDUCTIVITY, DENSITY AND POROSITY OF A SHALE, A LIMESTONE AND A DOLOMITE FORMATION

CHAPTER VIII

REGIONAL HEAT FLOWS

VIII-1 SEDIMENTARY HEAT FLOWS - AN INTRODUCTION

Where boreholes have been especially drilled as part of a geophysical or structural programme, a coring bit is often used and a complete or almost complete core obtained. This permits measurements of temperature gradient and thermal conductivity to be made in the same borehole and the combination will give a very high order of accuracy in the heat flux. The U.W.O. drillhole discussed in Chapter XI is such a hole. Certain problems still arise if the hole is drilled by oil or gas companies since they tend to test anomalous regions as indeed do mining companies in hard rock areas. Sampling of the core to get a good estimate of the thermal conductivity still has to be undertaken with a great deal of care since conductivities of individual units may be several hundred per cent different from those of a different unit. Gough (1963) in his results for four boreholes in the Southern Karroo of South Africa found that the strata he encountered could be divided into some seven types ranging from dark shales to coarse-grained sandstones. To obtain thermal conductivity estimates he took a 200 ft section each time and made measurements on the two major components in each section and then produced a weighted average for the section. A better approach is to use geological horizons for gradient and conductivity combination. Diment & Robertson (1963) tried several methods of combining gradients and conductivity. They selected four intervals each of 100 metres length through which the temperature depth line was relatively linear and combined the slope of the temperature depth line and the harmonic mean conductivities of the rocks. A second method was to divide the hole into 30 metre intervals and perform the same analysis. Lastly they used the method suggested by Bullard (1939) which uses the observed temperature and a conductivity depth summation for each interval. The authors suggest that in this case the second method probably gave the best results for heat flow at Oak Ridge but leave the question open. Somewhat more will be said on this topic in connection with the U.W.O. hole discussed in Chapter XI. No concise survey using purely this type of technique has ever been possible in a sedimentary basin; usually results are quoted for one or two holes in a limited area.

By far the most promising techniques are the ones using various degrees of estimation of the thermal conductivity. Very great dangers exist in attempting to estimate thermal conductivities from descriptions as drillers logs can be incorrect in their description of rock type, particularly in old wells, and as discussed in the previous chapter conductivities of a given rock type such as dolomite can vary by 100%. Birch & Clark (1945) attempted such estimations of conductivities in a borehole in West Texas, and arrived at results twice as large as the more recent estimates by Herrin & Clark (1956). The latter authors themselves estimated the conductivity of evaporites and used gradients in these horizons. The fact that gradients were relatively constant in these horizons over a wide area led them to conclude that the conductivity of evaporites does not vary by more than 10%. As part of the work for this project evaporites have been collected in Southwestern Ontario and conductivity measurements made, since measurement of gradients in several hundred feet of salt might present a rapid method of obtaining good heat flow coverage.

The results of these conductivity measurements given in Chapter VII-4 confirm the earlier measurements on evaporites and thus present a useful method of making heat flow determination in as yet untested basins such as those in Nova Scotia, the Williston in Saskatchewan and the Sverdrup Basin in the Arctic, all of which are believed to contain great thicknesses of evaporites. Measurements of gradient and conductivity in the evaporite sequence alone may not however be sufficient since salt in particular is often highly fractured and thus the question of water movement arises. A similar technique has been used more recently by Sass et al (1968) and Combs & Simmons (1969) in the United States. Garland & Lennox (1962) made several determinations of heat flow in Alberta, Canada. For one field they were able to collect core from other holes within 23 km of the one in which gradient measurements were made. They then suggested that the lithological description of the samples from a further field some 80 km away was quite similar. Therefore the same conductivities were used for both fields, resulting in heat flows that were some 10% different. Whether or not their efforts to explain the difference on the basis of radioactive element distribution is of any value is rather doubtful without more conductivity measurements from around the latter field. Coster (1947) encountered a similar problem in his heat flow determinations for a South West Persian oil-field where only small numbers of representative specimens from the penetrated formations were available and no information as to where exactly they were collected. In all the thermal conductivities showed standard deviations of some 20% making structural interpretations of heat flows varying by 10% seem a little futile. Of a rather different nature is the work of Joyner (1960) in the Appalachian Basin. His object was to use much of the temperature data collected by Van Orstrand and others and to determine conductivities using the detailed lithological logs available for many of the wells. By comparison with the average values for the conductivities of sedimentary rocks measured by other workers in other areas, he hoped to compute conductivities from the well descriptions. Such determinations were believed by the author to give errors of only 25% at the maximum and he further suggests that the additional use of electrical resistivity, neutron or velocity logs might improve this figure to 10%.

The only basin mapped thermally in any great detail is in Hungary This has been achieved by an ad hoc variety of methods, of low accuracy, by Boldizer (1958, 1964), none of which are suitable for examining variations in a low heat flow.

In the present survey temperature gradients have been collected where possible as discussed in Chapter VI-2 and core sections collected for thermal conductivity measurements from as wide a region as possible, as discussed in Chapter VII-2. These are now combined into heat flow measurements.

VIII-2 REGIONAL HEAT FLOWS

A measurement of the heat flow can be obtained in several ways as discussed in Chapter XI. However, the object here, since core sections come from a variety of holes different to those in which temperature gradients were measured, is to obtain a reconnaissance heat flow distribution. This is probably best done by comparing calculated heat flows across several different geological horizons in each borehole and examining them for consistency within a given borehole. Sections which show a near surface curvature due to recent climatic effects are ignored. Using the average heat flow in a borehole it is then possible to calculate thermal conductivities for unsampled horizons, using the reults of VII-4.

Region 1

Several heat flows have already been published in this region. The borehole at Franktown by Jessop (1968) have 0.99 μcals/cm^{2}sec ± 12% and that at Ottawa by Jessop & Judge (1971) have 0.80 ± 4%. Both of these results are based on results in Precambrian formations. At neither Picton nor Russell do the boreholes penetrate the Precambrian and so heat flows are determined in the Middle Ordovician rocks. The results from these sites are significantly higher than the earlier results, 1.04 and 1.28 respectively. Since the Russell and Ottawa holes are only 29 km apart, the differences are rather striking, and require some explanation. A possible one lies in the conductivity measurements. In Chapter VII the difference in Precambrian conductivities was discussed in some detail. The average Ottawa values were 5.7 compared with 7.1 for other South-western Ontario samples, and the latter were water-soaked where the former were not. Walsh & Decker (1966) have discussed the possible effect of microfractures on the thermal conductivity of granites. Thus the heat flow at Ottawa could be as high as 0.90. However this would still leave a 0.38 difference between the sites. A further explanation and a very possible one is that the samples chosen for the Russell borehole were the more competant and hence higher conductivity ones. At the time of the conductivity measurements workers at the Observatory had less experience in handling friable sediments than U.W.O. workers. At the Picton site the heat flow is 1.04 with lower conductivities at that site than at Russell. Further measurements were available in the St. Lawrence Lowlands (Saull et al, 1962). Heat flows vary from 0.74 to 1.05 with an average of 0.80. Conductivity measurements were made on dry rocks. Mean conductivities corrected for the saturated condition will be increased by approximately 10% leading to an average heat flow of 0.90 if the saturated values for the Montreal borehole are similar to the saturated Potsdam sandstone in the Russell borehole.

Region 2

Misener et al (1951) have published a heat flow value for the C.N.E. borehole of 1.03. If a small correction is made for the proximity of Lake Ontario this is increased to 1.05. At the Chinguacousy site heat flows could be compared in several formations. No core was available for the borehole itself and so average conductivity values have been taken for the Precambrian across Southern Ontario and values for the Trenton-Black River Groups from the C.N.E. and U.S.Steel boreholes. Using the C.N.E. conductivities the Trenton and Black River formations give heat flows of 0.99 and 1.10 respectively whereas the Precambrian result is 1.08. With the U.S.Steel conductivities the results are significantly higher. However Caley (1940) describes the Trenton and pre-Trenton as rather shaley in comparison to the U.S. Steel core.

In the C.N.E. only two Trenton conductivities were measured and thus the value may not be reliable. This leads to an average heat flow in the Chinguacousy hole of 1.09, leading to Trenton conductivities of 5.7 and a Billings shale value of 4.0 . The Morristown hole has a low Precambrian temperature gradient which leads to a heat flow of 0.90. This latter result appears disturbed by water-flows and requires further work. A mean heat flow for the area is 1.07.

Region 3

The Trenton-Black River temperature gradients in these boreholes increase from north to south. Using mean conductivities from the C.N.E. borehole this leads to heat flows ranging from 0.90 to 1.16 whereas using the U.S. Steel results the values are 0.95 to 1.22. Usually the formation logs in this region do not subdivide the two groups in the subsurface. However Sanford (1961) shows isopach maps indicating a thickening of the Black River from north to south and a corresponding thinning of the Trenton. Since it was shown in Chapter VII that the Black River has a higher conductivity than the Trenton this would in fact increase the variation in heat flow even more. Comparison of G.N.T. logs between Haldimand Co. and Essex Co. certainly emphasises the increased shaliness of the former beds. Similarly the Black River appears shalier in the Haldimand hole than in the Chinguacousy hole whereas the Trenton appears quite similar. Obviously it would be very useful to have a complete core in the Niagara region but failing this a reasonable heat flow to adopt is probably the lowest values since U.S. Steel conductivities are similar to the undolomitised ones in

Essex Co. whereas the Niagara area would seem to be one of increased shale content, ie. lower conductivity. Thus 1.05 is adopted as the regional heat flow with indications of lower values in the north and higher in the south.

Region 4

One heat flow value from this area has already been published (Judge & Beck, 1967, Beck & Judge, 1970), giving a mean value of 0.78. However, as described in Chapter XI, this includes a zone of low heat flow with apparent water flows. The average heat flow in the lower section of the borehole below this flow and through the Queenston shales is 0.86. Only the Dunwich borehole penetrated the Trenton in this area but was cemented off partially through the Black River. This section of the hole together with the U.S. Steel conductivities give a heat flow of 0.99. However the Queenston section of the hole yields 0.85 using U.W.O. conductivities. Similarly the Guelph section using U.W.O. conductivities give 0.85. Subdividing the Salina into each of its constituent divisions and using the conductivities for each from the U.W.O. results yields a heat flow of 0.92. This value may be a little high since the U.W.O. hole is on the edge of a small remanent evaporite bank. Averaging the heat flows leads to a value of 0.91 for the Dunwich hole. In the South Walsingham borehole use of U.W.O. Guelph conductivities results in a heat flow of 0.87 compared with one in the lower Detroit River section of 0.88. Again the Salina was split up into subsections and an average thermal conductivity calculated using the U.W.O. results, which gave 0.95. However using the U.S. Steel values directly since it has a similar lithology to the Walsingham hole gives 0.90. Thus we have a resulting heat flow of 0.88 for this borehole and for the region.

Region 5

In the boreholes in the Becher, Kimball and Bickford areas the water-tables were quite low, below 240 m, the holes being cased into the Guelph. However temperature gradients through the Salina formation were measured. The heat flows are calculated by subdividing the Salina group and assigning conductivities to these sections on the basis of conductivities measured in the Argor hole and the Enniskellan, Plymton and MacGillivray boreholes. Calculated heat flows are 1.01, 0.87, and 0.98 in the Becher, Kimball and Bickford boreholes. Changing the composition of the Salina A1 and A2 sections to allow for the varying accurrence of limestones and dolomites only changes the heat flows by 5%. With only one formation it is difficult to assign any reliability to these values. However comparing heat flows in one section of the Salina, the B-salt, gives an average value of 0.96 which tends to verify the above results.

Heat flow results determined in the Riddel-Hanna borehole with conductivities taken from the same boreholes results in values of 0.88 in the Detroit River, 0.89 in the Bois Blanc and 0.88 in the upper Salina section penetrated. The heat flow results in region 5 range from 0.87 to 1.01 with an average of 0.94.

Region 6

The Gosfield township borehole is a very interesting one since the heat flows in the Queenston and the Salina average 1.01. However the Guelph formation has a very low heat flow of about 0.6 which is a rather similar situation to that in the U.W.O. borehole. The water record for this hole shows that salt water was struck at 373 m, close to the top of the Guelph, producing initially 5 barrels/hr. Thus the system was an artesian one with some pressure driving it. Unfortunately the Colchester South borehole is disturbed by the nearby presence of the lake together with some flow pattern which is very hard to make any sense out of. The two Malden holes give heat flow values in the Trenton-Black River of 1.17 using thermal conductivity values from the Colchester field. A further value in the Malden area was obtained using the Salina-Guelph section of one of the holes. This gives a heat flow of 1.20. The fact that these results are obtained from an industrial log, that they give rise to heat flows substantially higher than the Gosfield hole and a published value for a hole close to Detroit makes the values suspect.

Region 7

Leney (1956) reported a heat flow result of 0.8 close to Detroit using the combination of temperature measurements in a 152 m surface hole which was disturbed by a temperature inversion and a 66 m underground hole in a salt mine. Heat flow values can also be obtained for the Muttonville and Northville holes since they are only 60 km from the boreholes in Regions 5 and 6 for which cores and thus conductivities were available and they contain similar formations in the lower parts of the holes. For Muttonville, the conductivities from the completely cored drill-hole Argor #1 have been used. The mean heat flow through the Palaeozoic section from the top of the Detroit River to the base of the Upper Salina is 0.81 ± .02. A comparison value using the conductivity of rock-salt and the temperature gradient in the Salina B salt yields a heat flow of 0.81 also. Considering each formation individually and calculating the mean heat flow yields 0.82. To calculate the heat flow at Northville the conductivities of Region 5 plus those determined by Leney are used. The mean heat flow through the Palaeozoic section from the top of the Detroit River to the base of the Upper Salina is 1.1 ± .1. Calculating the heat flow for each formation yields a mean heat flow of 1.2 ± .1. A heat flow of 1.15 is adopted for this site. The heat flows in Region 7 vary from 0.8 to 1.2 with a mean of 0.92.

Using the adopted value for the heat flow at Northville, conductivities may be calculated for the Traverse and Antrim formations yielding values of 5.5 and 3.2 respectively. These values may be slightly too low since the temperatures in the shallow portions of the borehole are disturbed by an inversion.

Region 8

Although no conductivity measurements were made on core from the Overisel field, very comprehensive descriptions of the rock types were available and thus mean conductivities of similar rock types from Regions 1 to 7 could be assigned to the formations. The sections of the holes between a depth of 520 m and bottom hole at 825 m passed through formations similar to those in southern Ontario and eastern Michigan. Mean formation heat flows for these sections were calculated to be .90, .90, and .93 for Overisel holes # 157, #162 and #150 respectively. Since the respective temperature gradients varied by 15% whereas the calculated heat flows vary by only 4%, the technique would seem to be valid. A mean heat flow of 0.91 ± .01 is adopted for the region.

Again this enables conductivities to be calculated for the shallower geological horizons such as the Traverse, Antrim and Coldwater, which are 7.7 ± .8, 3.2 ± .2 and 4.2 ± .4 respectively. Since the formations are described as limestones, shales and a mixture of limestone and shale, the values are reasonable.

Region 9

This region in Central Michigan is the most difficult one to calculate heat flow for because none of the formations through which temperature measurements were made were encountered in southern Ontario. Hence no values for the thermal conductivities of the formations were available.

Several boreholes in Regions 7 and 8 however penetrated formations which are present in southern Ontario and also formations present in Region 9. The Bregg hole for example penetrates 122 m of Traverse limestone. In the Northville hole in Region 7 the thermal conductivity of the Traverse was estimated to be 5.5, giving a heat flow in the Bregg hole of 1.0. The Traverse conductivity was thought to be low so the above value gives a minimum value. Jodry (1957) describes the Traverse as a dense chemical carbonate and thus the conductivity should be much higher. The boreholes in the Overisel field of Region 8 penetrated the Traverse limestone, and the Antrim and Coldwater shales, all three of which occur in the Bregg hole. Using these estimates for the conductivity the heat flows are 1.4, 1.6 and 1.6 respectively. However both Jodry (1957) and McGregor (1954) point out the changes of character of these formations from east to west. The Traverse changes from a dense chemical carbonate to a limestone composed of reworked fossil material containing a higher proportion of dolomites and evaporites. Whereas in the west the Antrim is a grey-black calcareous shale with interbedded argillaceous limestone and dolomite grading into a black carbonaceous shale, in the east it is a dark grey to black carbonaceous shale with beds of calcareous grey shale. In the west the Coldwater is generally a shale with interbedded limestone and dolomite grading in the east to a siltstone and fine-grained sandstone with shale interbeds. Each of these descriptions could lead to slightly lower conductivities at Bregg in comparison with the Overisel values.

The other boreholes in Region 9 only penetrate to the Marshall sandstone of the 'shoestring' gas fields. Temperature gradients in the Saginaw and Michigan formations are similar to those in the Marion, Austin and

Billingsly so that if the Bregg hole is not in equilibrium and the heat flows are too high, then this should apply to all of the holes. The temperature gradient in the Michigan is higher than that in the Saginaw except in the Billingsly hole suggesting a higher conductivity in the latter. However the difference is not constant but varies between 2 and 20%. The latter suggests rather variable conductivities in these formations through the region. Adopting initially a mean heat flow of 1.5 for the region, the conductivity of the Michigan formation in the Bregg hole is 6.0 and that of the Saginaw slightly higher. Using these conductivities the mean heat flows in the other holes vary from 1.2 in the Billingsly hole to 1.6 in the Austin hole.

Ball et al (1941) describe the Michigan of the Austin field as being composed of shale with some gypsum and anhydrite bands and interbeds of brown dolomite and limestone. The Saginaw is also composed mostly of shales but with interbeds of micaceous sandstone. A conductivity of 6 is rather excessive for rocks of such a description; 5 would be a better value to use. The range of heat flows is then .9 to 1.4 with a mean of 1.2 ± .2 for six boreholes. In the Marion holes which are close to each other the range is 1.1 to 1.3. A mean heat flow of 1.2 is adopted (with reservations) for this region.

A further rough determination could be made using the bottom hole temperatures from the Rose City drill-hole discussed in VI-3A. Subtracting the extrapolated surface temperatures in Central Michigan gives temperature gradients of 16.3^{o}C/km to the base of the Guelph, and 14.7 and 18.5^{o}C/km into the Cambrian. The Cambrian is described as the Mt. Simon sandstone for which a conductivity of 11.0 to 12.0 has been determined in Chapter VII, yielding a heat flow of 1.1. Using the mean temperature gradient to the base of the hole together with a mean weighted conductivity of 6.6 for the formations penetrated determined from other boreholes and from conductivity results in Chapter VII, the average heat flow is 1.2. In shallow holes bottom hole temperatures taken immediately after drilling are unreliable because any error in them causes a large gradient error. However in such deep holes as this one an error of 2°C in the bottom hole temperature changes the heat flow by only 3%. Likewise errors in estimating the conductivity of any one formation become less significant. Thus a heat flow of 1.2 seems reasonable for Central Michigan.

VIII-3 DISCUSSION OF HEAT FLOW RESULTS

The heat flows and their variation in each of the regional subdivisions are compared in the table below:

Region	Area	Average Heat Flow	Range of Values	No. holes
1	Eastern Ontario & Western Quebec	0.90	0.74 - 1.28	9
2	Toronto Area	1.07	0.90 - 1.09	3
3	Niagara Peninsula	1.05	0.95 - 1.22	4
4	London Area	0.88	0.86 - 0.92	3
5	Northwest Section	0.94	0.87 - 1.01	4
6	Southwest Peninsula	1.12	1.01 - 1.17	3
7	Southeastern Michigan	0.93	0.82 - 1.15	3
8	Southwestern Michigan	0.91	0.90 - 0.93	3
9	Central & Northwest Michigan	1.2	0.9 - 1.4	6

Table VIII-1 REGIONAL HEAT FLOW MEASUREMENTS IN SOUTHERN ONTARIO AND MICHIGAN

Values are shown on FIG VIII-1 for southern Ontario and the rest on FIG VIII-2 which covers the Great Lakes region.

In several of the regions above, eg., Region 3, at least a part of the variation may be due to lithological changes and the heat flow is not as variable as calculated. However in most of the regions it was possible to determine the heat flow through several horizons and thus verify the values.

Excluding the values of Region 9, the mean heat flow in the area is 0.97 ± .13 with a range of 0.74 to 1.28. The mean heat flow at all of the sites is 1.0 ±.2. Restricting the observations to those west of the Frontenac axis and on a Grenville basement, the mean heat flow is 1.00 ± .12 in a range of values of 0.82 to 1.22.

These results are all uncorrected for the glacial effects discussed in some detail in Chapter XII. If it in fact is present in these boreholes at a magnitude above the errors, it should be observable in two ways; the heat flows measured at greater depths should be greater than those at shallow depths (but below the inversion zone) and heat flows should be greater in formations with a low conductivity than those with high conductivities. As will be discussed in Chapter XI there was some evidence for the former in the London borehole. It will also be shown in Chapter XII that the predicted temperature gradient correction is approximately constant across the region. There are several areas where these corrections can be examined using the above criteria. In Region 2 and in the northernmost hole of Region 3 the Trenton occurs at depth of 500 m or less whereas in the other three holes of Region 3 the formation top is at a depth of 800 m. However the mean gradients are 16.3°C/km and 17.7°C/km respectively. If the past surface temperature variation model of Chapter XII is correct the difference between these two gradients should be 3°C/km. In Region 4 a medium conductivity formation occurs at a depth of 1000 m whereas a high conductivity formation occurs at a depth of 300 m. The proposed surface temperature variation model of Chapter XII would result in a heat flow about 12% higher at the top than at the bottom. However when the proposed model correction is applied to two formations of similar gradient at depths of 500 m and 1000 m, it is of the correct order. In conclusion it is apparent that the model derived from Dreimanis' and Terasmae's time-scale, and which gives smaller corrections than the conventional models, is in better accord with the heat flow observations made. The required corrections are no larger than this model proposes but may be smaller.

Thus at least part of the observed variations in heat flow may be due to the disturbances caused by past surface temperature variations. Several of the shorter drill-holes give lower heat flows. A completely cored 1000 m would probably be sufficient to derive Pleistocene corrections.

Results of gravity and magnetic surveys in southern Ontario as discussed in XI-2 and -3 suggest that the basement across southern Ontario is very uniform in composition and that the variations in the gravitational and magnetic fields are explained by an extension of the Cincinnati Arch, and the subsequent folding of a backbone material, 0.04 gm/cm^3 denser and 2000×10^{-6} c.g.s. units lower in susceptibility than the surrounding basement, into Ontario. It is very unlikely that large differences in the radioactive heat production would result and thus the heat flow would be expected to be uniform.

In Michigan the higher heat flow of Region 9 and of Northville in Region 7 appears to correlate with a gravity high, and the lower heat flow of Muttonville in Region 7 correlates with a gravity low. However Region 8, also with lower heat flow, is on the edge of a gravity high. A magnetic high corresponds to the gravity high through Michigan. This is rather confusing since gravity highs are normally related to basic rocks which usually have lower heat productions than acidic rocks. Obviously far more sites and more analyses of other geophysical results are required in Michigan.

VIII-4 PREVIOUS HEAT FLOW MEASUREMENTS COMPARED

Previous heat flow measurements in the immediate area are limited, with one hole in the Canadian National Exhibition Grounds in Toronto (Misener et al, 1951) for which a heat flow of 1.03 $\mu cal/cm^2 sec$ was obtained and a further heat flow of 0.8 $\mu cal/cm^2 sec$ obtained near Detroit in Michigan by Leney (1956). The latter hole was rather shallow and may not penetrate the inversion zone discussed in Chapter XII. Quite a few limiting values exist from around the region particularly on the Canadian Shield. These results are shown in FIG VIII-2.

Heat flow values in the Canadian Shield, uncorrected for Quaternary surface temperature changes, vary between 0.69 and 1.07 $\mu cal/cm^2 sec$, with 50% of then falling within 0.9 ± 0.1 which seems to indicate a fairly uniform heat flow. Such observations agree with those for other shields (Kraskowski, 1961). Subdivisions of the values from the exposed shield into metamorphic provinces gives a mean heat flow for the Superior province of 0.90 ± .13 with a range of 0.7 to 1.07 in 11 values, and for the Grenville of 0.89 ±.16 with a range of 0.79 to 1.22 in 8 values. Clearly there is no significant difference in the mean values at this stage although both the minimum and maximum values are higher in the Grenville. As discussed in the previous section and in Chapter IX, all of the heat flow values excluding Regions 8 and 9 determined in this work and those reported by Saull et al (1962) in the St. Lawrence Lowlands are in areas underlain by a basement which is Grenville in age. The mean heat flow on the unexposed areas is $0.97 \pm .13$ with a range of 0.74 to 1.28 for 25 sites. There is no significant difference in the heat flows measured on the exposed and unexposed shield. Restricting the observations to the southern Ontario results west of the Frontenac Axis and as far west as the Grenville boundary discussed in Chapter IX, the mean heat flow is $1.00 \pm .12$ in 21 sites with a range of values from 0.82 to 1.22. Obviously many more results are required on the exposed portion of the shield to determine the significance of these results. FIG VIII-2 shows all available heat flows in the Great Lakes region.

Roy et al (1968) have shown that there is a consistant relationship, in a given geological province, between the measured heat flows and the heat production of the rocks. Most of the variation in heat flow in the eastern part of North America is explained by heat production variations in a layer about 10 km in thickness.

No heat production measurements were made as part of this work, however comparison with Roy et al's results in the Adirondack portion of the Grenville suggests variations in heat production of 0.2 to 5.8×10^{-13} $\mathrm{cal\,cm^{-3}sec^{-1}}$. Temperature distributions in the crust corresponding to these values are discussed in IX-9.

Comparison of the mean heat flow in other regions of North America is made in the table below:-

Geological Province	Mean Heat Flow	Range of Values	No. of Boreholes
Shield (exposed)			
Superior	0.9 ± .1	.7 → 1.1	11
Grenville	0.9 ± .2	.8 → 1.2	8
Shield (unexposed)			
Grenville	1.0 ± .1	.7 → 1.3	25
Interior Lowlands	1.2 ± .3	.7 → 1.8	22
Appalachian (metamorphosed)	1.5 ± .4	1.0 → 2.3	13
Cordillera	1.7 ± .5	0.7 → 2.4	9

Table VIII-2 MEAN HEAT FLOWS IN THE GEOLOGICAL PROVINCES OF NORTH AMERICA

The result from the Cordillera is quoted from Lee and Uyeda (1965) and as such is out of date. However more recent data has tended to subdivide the area into smaller zones without changing the range of values or the mean significantly. The result for the Interior Lowlands contains all the U.S. results south of the Great Lakes not in the metamorphosed Appalachian, but believed to be sediments covering a Precambrian basement, plus the Michigan results from Regions 8 and 9. This includes sites on both Grenville and Central Province basement. Since heat flows in the unexposed Grenville were similar to those on the exposed shield, the mean heat flow in the Central Province must be higher than that in Grenville. This is consistent with the conclusions of XI-4 that the former had undergone little erosion and metamorphism whereas the Grenville contained heavily metamorphosed material which probably originated deep in the crust and thus is less radioactive. However this in inconsistent with Roy's conclusion that most of the variation of heat flow is contained in a layer of similar thickness in the Grenville and the Central Provinces.

In general the area of most recent metamorphic activity, the Cordillera, has the highest heat flow of 1.7. Appalachia, metamorphosed in the Devonian, has an average of 1.5, whereas the Shield rocks of Superior and Grenville age average 1.0. A relationship between heat flow and basement age appears to exist, as has been suggested by Hamza & Verma (1969).

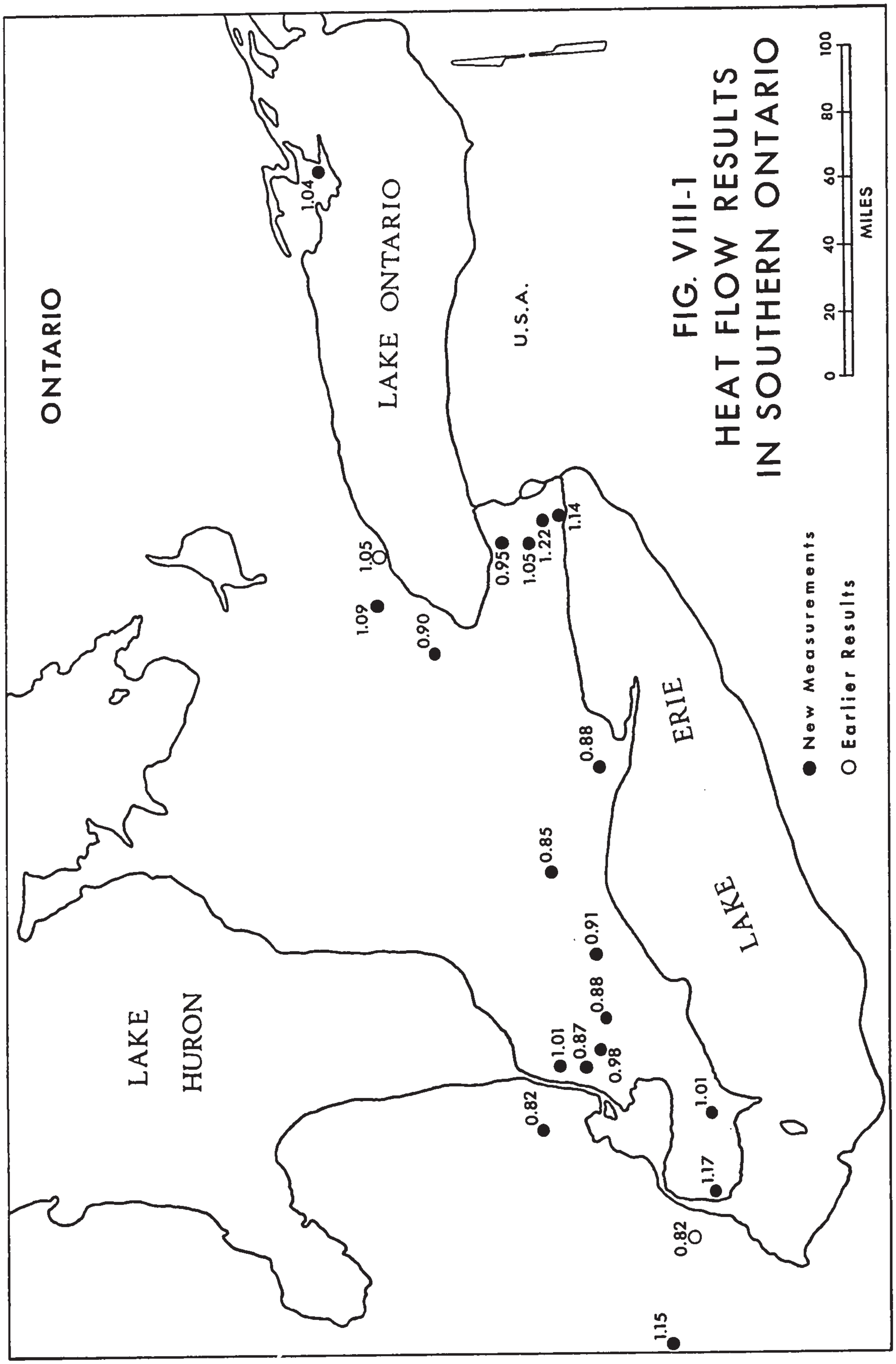


FIG. VIII-1
HEAT FLOW RESULTS
IN SOUTHERN ONTARIO

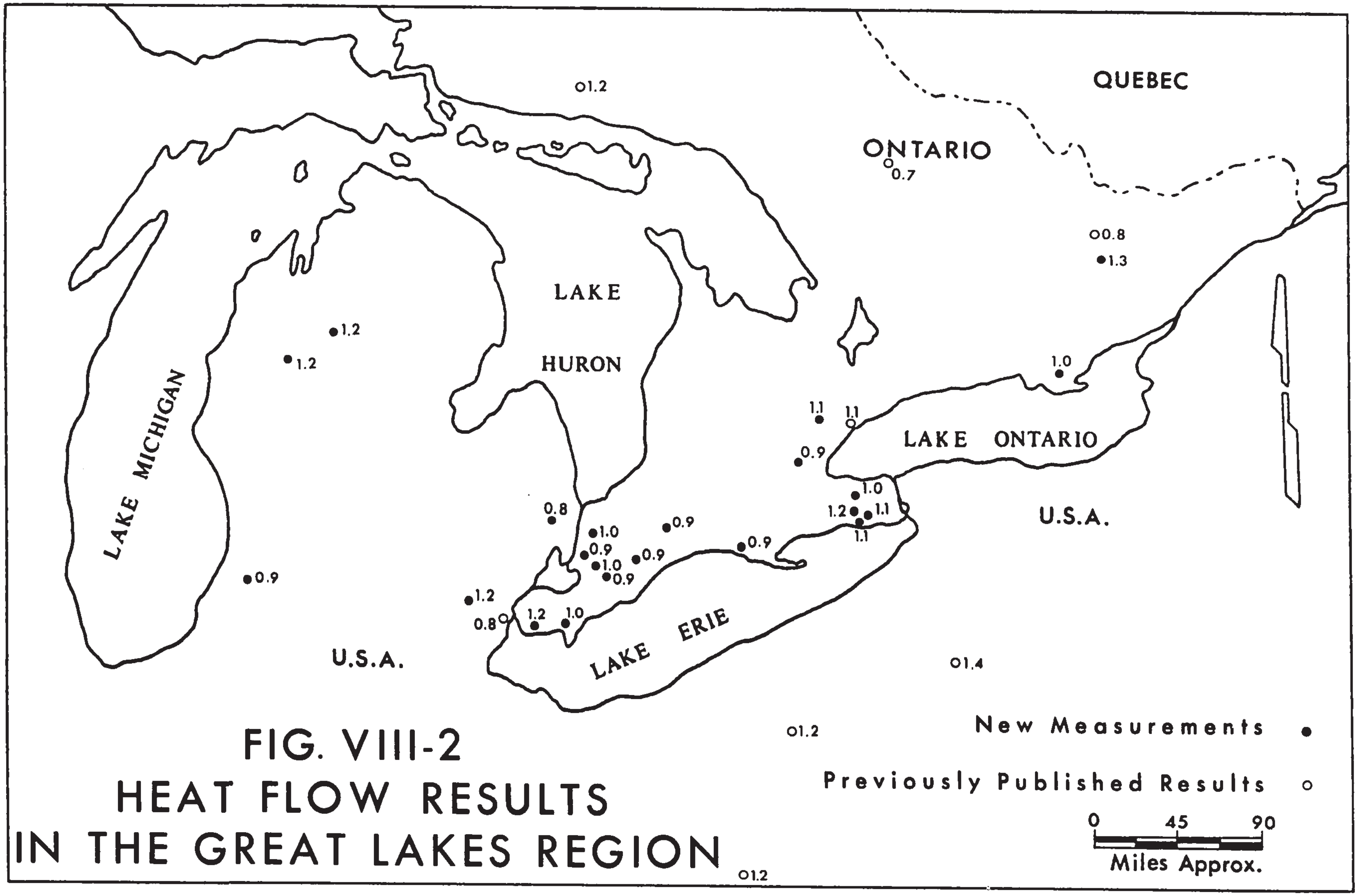


FIG. VIII-2
HEAT FLOW RESULTS
IN THE GREAT LAKES REGION

CHAPTER IX

CRUSTAL STRUCTURE

AND

THE TEMPERATURE FIELD

IN THE GREAT LAKES BASIN

PREFACE

In any attempt to interpret heat flow measurements over a large area it is of fundamental importance to know the thickness and the nature of the crust in that region. This need arises since the major contributor to the observed terrestrial heat flow is the heat generated by radioactive isotopes present in crustal rocks. However, different crustal rocks contain very different amounts of these isotopes. Since in only very few areas have holes been drilled to any great depths (eg., the deepest boring in sediments is approximately 8 km and in basement rocks about 3 km), it is necessary to derive most of this information by indirect methods. The most developed methods are those of seismic, gravity and magnetic measurements, and this chapter is an attempt to see what information has been derived by these or other techniques.

IX-1 SEISMIC EVIDENCE

During the course of this project, it became obvious that only the scantiest of information was available, with most of it included in Steinhart et al (1961). The results are summarised in FIG IX-1. The earliest crustal work was that of Schlichter (1951), using open-ended lines from Palmer, Michigan, running south, from Manistique, Michigan, to the west and from Sturgeon Bay, Wisconsin, to the southwest. Crustal velocities range from 6.16 km/sec to 6.26 km/sec, with a 4.6 km/sec section of Palaeozoic sediments above. Upper mantle velocities were consistently around 8.17 km/sec with mantle depths of 36 km. To the north of the region, beneath Lake Superior, the crust thickens to 50 km, but this is believed to be an anomalous situation, very restricted in extent (Steinhart et al, 1964); thus the Moho gradient is very steep in northern Michigan. Cohen and Meyer (1966) have summarized their own work in the Mid-Continent Gravity High and Schlichter's crustal thicknesses into a contoured map with a range of thicknesses of 36 to 46 km. To the east of the Great Lakes region Hodgson (1953) obtained a 37 km crust-mantle discontinuity with velocities of 6.25 km/sec above and 8.18 km/sec below for an open-ended line between Kirkland Lake and Ottawa. Katz (1955) found Moho depths of 36 km in the Adirondacks with velocities of 6.63 km/sec and 8.2 km/sec. To the north of Lake Superior and in the direction of Churchill, crustal thicknesses average around 30 km and thicken ino 40 to 50 km in the younger Churchill province (Mereu & Hunter, 1969), with crustal velocities of 6.23 km/sec and upper mantle velocities of 8.12 km/sec.

A brief analysis was made of the University of Michigan results from recording stations occupied on a line through the northern and southern peninsulas of Michigan during project 'Early Rise'. Plotting a travel-time curve suggested crustal and upper mantle velocities of 6.34 and 8.15 km/sec respectively. The crustal velocity is a little higher than much of the shield which agrees with Hinzes (1963) postulate that the gravity and magnetic anomalies in the region are caused by Keewanawan lavas. Crustal thicknesses along the line varied between 31 and 37 km using the time term analysis on the stations in the northern portion of the southern peninsula of Michigan. At greater distances from the shot point the interpretation becomes more complex but would suggest a crustal thinning in the vicinity of Howell, Michigan and then a thickening into Ohio.

IX-2 GRAVITY SURVEYS

Gravity surveys of southwestern Ontario have been conducted by Brant (1943), Prendergast (1952), Oldham (1954) and Thompson and Miller (1958). In Michigan, gravity results have been published by Bacon (1966) and Hinze (1963); and in Ohio by Heiskanen and Uotila (1956). These results are summarised in FIG IX-2. The Ontario results show, in general, a band of gravity highs between Wallaceburg and Mitchell reflecting the axis of the Algonquin Arch. Highs also occur in the Winsor area with values decreasing towards the Michigan Basin, suggesting a correlation with the Findlay Arch. Thompson also finds a high entering the Niagara Peninsula around St. Catharines. Gravity lows occur in Lake Erie off the Pelee Point region, in Lake Huron off Southampton and north of Toronto in the vicinity of Newmarket. Brant has noted that the average density of the Palaeozoics is about the same as the basement gneisses, hence anomalies greater than 0.3 milligals probably reflect changes in composition of the basement. He suggest two types of material of slightly different density with the denser material closer to the surface in the west.

Extension of the coverage to include the Michigan Basin broadens the picture and, whereas the largest anomalies in southern Ontario were -20 milligals, anomalies of +40 milligals are found. The average baseline of Bouger anomalies is approximately -30 milligals with localized lows southwest of Flint and around Rogers City in the north. Certainly the most prominent feature in the whole region is the gravity high extending from the northwest to the southeast right across the centre of the basin. Along this high, values range up to +10 milligals. As is shown in FIG IX-2, this anomaly is about 300 miles long and 50 miles wide. As discussed partially in the previous section, Hinze has suggested that this anomaly is caused by a belt of volcanics similar to those comprising the Mid-Continental Gravity High. Although no wells have penetrated basement in central Michigan, drilling in a positive gravity anomaly in Ohio indicated a basement composed of basic intrusive rocks. Gravity compilations by Woollard (1966) suggest that the region is close to isostatic equilibrium, so that it should be possible to interpret the anomaly with fairly simple ideas. Assuming the mantle to be uniform across the region and of the same depth, then the central anomaly may be explained by a density difference of 0.05 gm/cc in the crustal column. This same effect could be derived by including 5 km of volcanics in the crustal column. Probably, then, part of the crust in the Michigan Basin is composed of Keewanawan rocks and these give rise to the gravity anomaly in the central part of the basin.

On the basis of isostatic anomalies, Steinhart and Woollard (1961) deduced a crustal thickness of 35 to 39 km across the basin. Assuming greater densities in the crustal column would decrease these values.

IX-3 MAGNETIC EVIDENCE

Ground magnetic surveys in Michigan by Hinze (1963), in Ontario by Brant (1943) and the sea-borne surveys in Lake Huron by Nwachukwu (1965) and in Lake Erie by Peter and Wall (1961), as well as aeromagnetic surveys in peninsular Ontario by the Department of Energy, Mines and Resources (1965), complete the picture of the region. In Ontario the regional magnetic background suggests a lithologically homogeneous basement consisting of granitic rocks and meta-sediments. Larger anomalies in the Niagara peninsula suggest an association of horneblendic gneiss with the granitic rocks. The magnetic trends are northeast to southwest in keeping with the trends in the Grenville Province in general. Nwachukwu et al (1965) have attempted to use the directions of the magnetic lineations to determine the position of the Grenville Front in Lake Huron. On this basis they have deduced that it crosses in a fairly straight line between Killarney on the north shore and East Tawas in Michigan. Just east of the Michigan shoreline it appears to bend sharply to the south and passes somewhere to the west of Detroit. Further to the south in Ohio, generalizations may be made from Zietz et al (1966). Their aeromagnetic strip shows a highly disturbed region with anomalies up to 800 gammas in central Ohio which cut off very abruptly on the west side; this cut-off probably represents the position of the Grenville boundary. To the west of this region the pattern is much broader and smoother with very few intense anomalies.

The most apparent magnetic structure in the northern region, as shown in FIG IX-3, is the magnetic high running across central Michigan and coinciding with the location of the gravity high. However, the largest magnetic anomaly occurs over the Howell anticline.

IX-4 PETROLOGY OF THE BASEMENT

The basement rocks of southern Ontario consist mainly of mica or horneblende gneisses. In the Niagara peninsula the horneblende gneiss predominates over the granite gneiss. These are observations based on cuttings and cores from the top twenty feet of the basement. No detailed petrological work on this area has been done. Further to the east, marbles and quartzites occur at the top of the basement and in some areas massive limestones of the Mistassini series. Directions of lineations, nature of the basement and a few age dates indicate that the gneisses are Grenville in age.

Few wells have been drilled to the basement in Michigan and most of these are in the southeastern corner close to the Ontario border. Again on the basis of examination of a few well cuttings, the basement is composed primarily of granite gneisses and horneblende gneisses. Isotopic age dates put these in the Grenville orogeny, together with the Ontario section.

Only in Ohio and Indiana has detailed petrological work been done on basement rocks, and these are again limited to a few well cuttings because of the thickness of the Palaeozoic cover. This work has been done by McCormick (1961) in Ohio and by Bradbury and Atherton (1965) in Indiana. Summarizing their results suggests that to the east of the suspected orogenic boundary the rocks are generally mica or horneblende gneisses with some granites and marbles; the region close to the Front is part of the central region of the Ohio platform and this seems to be composed of granites in the north close to Lake Erie, and rhyolites and trachytes to the south. Well to the west of the boundary, the basement seems to be composed of potassium-rich granites, diabases and unmetamorphosed sedimentary rocks. According to Zietz et al (1966), no metamorphic rocks have been found in the region. A few age dates by Lidiak et al (1966) suggest that the boundary zone on the platform is divided into granites of Grenville age and rhyolites and trachytes of Central Province age, with ages of 1.2 million years. Rudman et al (1965), however, place all of these rocks, even with dates of 1.3 million years, into the Grenville and place the Front further west so that it almost bisects Indiana from northeast to southwest. This was the position suggested by McLaughlin (1954) on the basis of earthquake epicentres. However, the magnetic evidence and the geological descriptions would tend to support its going through Ohio. How wide the boundary zone is elsewhere and whether the Central Province in Michigan bears out the relationships in Ohio are matters for conjecture at present. The results are summarised in FIG IX-4.

An examination of the chemical analyses of shield rocks by Eade (1966) and Reilly (1968) suggests that the Superior Province in north-western Quebec is rather more basic than in north-western Ontario and that the Grenville is still more basic.

IX-5 BASEMENT TOPOGRAPHY AND STRUCTURE

Most of the Palaeozoic structures in the Michigan Basin have a prominent northwest to southeast trend; well-known amongst these are the Albion-Scipio trend, the Howell monocline and the Northville field. All of these trends are probably related to faults in the basement which were still active in post-Cambrian times. Thus the basement movement has played an important role in Palaeozoic sedimentation, and examination of it may assist in studying the crustal relationships, and indeed the temperature distribution. Structural lows at present consist of the Michigan Basin, containing 4 km of sediments, the Appalachian Basin, containing more than 5 km of sediments, and the Illinois Basin with about 5 km. Separating each of these basins are the so-called arches: the Cincinnati-Wisconsin (Kanakee) separating the Michigan and Illinois Basins, the Cincinnati separating the Illinois and Appalachian Basins and the Findlay-Algonquin separating the Michigan and Appalachian Basins. Each of these merge into an area known as the Indiana-Ohio Platform. FIG IX-5 shows the thickness of sediments and the major structural features. The apparent reationships of these features have certainly not been constant throughout geologic time. For example, thicknesses of Cambrian rocks increase into the Illinois Basin with no sign of the Kanakee arch, although the Algonquin and Findlay arches already show as positive features. Middle Ordovician deposition indicated a structural high where the Kanakee arch is, but shows little sign of the Algonquin arch, a situation which almost reverses in the Upper Ordovician. As can be seen, the present structural picture has certainly not existed throughout geological history, although since the end of the Devonian the relationships have not changed greatly. It is difficult to see how these differential movements can take place excepting between faulted blocks, indeed the Findlay and the Algonquin meet at the Chatham sag which is bounded on the north side by the Electric fault, and certainly basement faulting is far more common than was formerly believed, as is evidenced by the occurrence of the Clearville and Colchester fields which appear to be fault-controlled. Brigham and Winder (1966) have plotted the directions of joints in southern Ontario. The rose plot south of the Algonquin arch shows a very prominent mode almost directly east-west, and further important modes northwest and northeast. To the north of the axis, the modes are rather more confused with a north to north-west one and a south-southeast one. It must be admitted that these are based on very few observations, but the directions are very similar to those of the magnetic lineations in the Grenville and in Michigan. However, the prominent east-west direction is confusing.

IX-6 EARTHQUAKES

A further source of evidence concerning the behaviour of the crust is in the earthquake distribution. Epicentral distributions in the area have been determined by Smith (1962, 1966) and Bradley (1965). Neither of these authors cover very much of Michigan nor are they able to determine depths to epicentres. Examination of their maps yields a very striking feature, ie., that the most active zones, those of the Niagara peninsula and the Anna area of Ohio, lie on a linear extension of the very active earthquake region of the St. Lawrence Valley. These epicentres are shown in FIG IX-4. The Anna region of Ohio is very close to the probable Grenville boundary and to the junction of the Kanakee and Findlay arches and may represent a fault zone with differential movement as it acts as a hinge zone. It would certainly be very interesting to look at the isopachs of past sedimentation and then to analyze the direction of present movement to see whether this is indeed a hinge zone.

IX-7 MAGNETIC VARIATIONS

Rostoker (1963) and Whitham (1964) have discussed results from magnetic variometer stations between London, Ontario and Sherbrooke, Quebec. They report that no major electrical conductivity anomalies exist beneath the region and thus it may be considered tectonically stable. The area in question includes Ottawa and the Logan fault zone, which is perhaps surprising.

IX-8 RADIOACTIVE HEAT PRODUCTION

The basement is Grenville in age beneath the southern part of Ontario and Central beneath the Michigan Basin. Few radioactive measurements are at present available for any of the Canadian Shield; the specific results for small areas are given below, using conversion factors of Weatherill (1966) to obtain heat production:-

Province	Location	Heat Production $\mu cal/cm^2 km$
Superior	New Quebec	0.031
Superior	Val D'Or	0.007
Superior	Preissac	0.068
Superior	Minnesota	0.014
Grenville	Bancroft	0.025
Grenville	Adirondacks	<0.004 - 0.058
Southern	Elliot Lake	0.042
Central	Missouri	0.055
Central	Oklahoma	0.076

Table IX-1 HEAT PRODUCTION RESULTS FROM THE CANADIAN SHIELD

General analyses of uranium, thorium and potassium content of large numbers of Precambrian crystalline rocks have been made by Lambert & Heier (1967) for the Australian Shield and by Fahrig et al (1967) and Shaw (1967) for the Canadian Shield. Heat Production figures based on their values are given below:-

Rock Type	Location	Heat Production $\mu cal/cm^2 km$
Amphibolites	Canada	.034
	Australia	.054
Horneblende Granulites	Canada	.024
	Australia	.014
Total	Canada	.038
Range		0.022 → 0.056

Table IX-2 RADIOACTIVE HEAT PRODUCTION OF TYPICAL SHIELD ROCKS

The measurements of Fahrig were in Northern Quebec which appears highly metamorphosed and deficient in radioactive elements. Shaw's results are an average of several shield areas.

Using the values given by Clark (1966), the radioactive heat production of other common rock types is as below:-

Rock Type	Heat Production $\mu cal/cm^2 km$	No. of Samples
Mafic Igneous	.007	169
Diorites and Quartz Diorites	.026	47
Granodiorites	.036	92
Silicic Igneous	.062	156
Shales	.045	75
Carbonates	.018	103
Sandstones	.006	18
Halites	<.005	7
Anhydrites	<.005	3
Dunites	.0002	unknown

Table IX-3 RADIOACTIVE HEAT PRODUCTION OF COMMON ROCKS

IX-9 TEMPERATURE FIELD IN THE EARTH'S CRUST

The major factors governing the temperature field in the crust are the thermal conductivity and the heat production. In a region such as the Canadian Shield the crust may be considered to be in thermal equilibrium, with the exception of small temperature perturbations due to recent surface history, eg., fluctuations in surface temperature and surface loading. Southern Ontario and the Michigan Basin are underlain by shield rocks and show no evidence of major crustal activity or movement in the western section since the Devonian and in eastern Ontario since the faulting of early Tertiary times (Hewitt, 1962). Roy et al (1968) have shown a linear relationship to exist between heat production and heat generation. This suggests that most of the lateral variation in the heat-producing elements (U, Th and K) occurs in a layer a few kilometers thick at the top of the crust. In their 'Central Stable Region', the province most akin to the Canadian Shield, this layer is 8.0 km thick which is in agreement with gravity interpretations of the depth extent of plutons in the Superior Province. The heat-flow from below this zone is 0.76. However there are several heat-flow values, eg., Val D'Or, Kirkland Lake, Noranda, Timmins, Brent, Montreal, which are lower than this intercept value. Thus it must be assumed that either the heat production is very low at these sites, the relationship does not hold, or, that structural or water-flow disturbances are resulting in non-equilibrium heat flows which are low. However the heat production figures given by Roy for the Adirondack sites are $<0.4\times10^{-13}$ cal/cm^{3}sec whereas heat production figures close to Val D'Or are 0.7 (Ingham et al., 1951). The latter leads to a heat flow at the base of the variable layer of 0.64. If however a heat production figure of 0.7, similar to mafic igneous rocks, is used for the lower crust such that the total crustal thickness is 35 km, then the mantle heat flow is 0.40 which, according to Hyndman et al (1968), is the lowest level which can be tolerated for mantle heat flow from other considerations. Thus to be consistent in the shield it must be assumed that parts of the shield are deficient in radioactive materials in the lower crust when compared with the 'Central Stable Region'.

The effect of glacial disturbances on surface heat flows complicate this picture even further since a reasonable glacial correction for Val D'Or increases the heat flow to 0.84 but increases that in the Adirondacks to 1.05. Obviously these uncertainties in the equilibrium heat flow make enormous differences to a crustal temperature field. Carslaw & Jaeger (1959) give suitable solutions for the temperature field under these conditions. In FIG IX-6 the temperature field with depth is plotted for each of these cases. Temperature differences of 100°C result at a depth of 30 km. Thus while it may be valid to make intercomparisons between the results in the Adirondacks where glacial corrections are similar, it is dangerous to attempt gross comparisons over large areas or to attempt to calculate and compare crustal temperatures over them. Since climates were also changed to the south of the ice-sheets, these comments apply there as well. However we can allow reasonable ranges of corrections and use these as the limits, as has been done above for Val D'Or.

The Michigan Basin contains 4 km of sediments which on weighting for composition yield a heat production of 2.2. This is lower than the heat production figure of 3.8 yielded by Shaw's (1967) average composition of shield rocks but similar to measurements in the Grenville basement (Ingham et al, 1951). Assume that beneath the sediments is Roy's 7.5 km layer of variable heat production, with a heat production figure of 4.0 as the mean of Shaw's heat production for the Superior and Sass' (1968) value for granitic outcrops north of Elliot Lake. Assuming a lower value of heat production makes for difficulties in not exceeding a heat flow of .45 at the crust-mantle boundary. However the facts that the Basin is in isostatic equilibrium, has an average crustal thickness and is a deep sedimentary basin suggests that the crust may be more mafic than the average shield which ought to result in a lower heat production. However from Chapter IX-4 the Precambrian terrain in the Central Province is largely non-metamorphic and hence may be more radioactive. The mean conductivity for the basin in each of the models shown in FIG IX-6 was chosen using the temperature constraints placed by the well discussed in Chapter VI-3. As at Val D'Or, upper and lower heat flow limits have been chosen for cases with and without the glacial correction, with the exception that the mantle heat flow has been constrained to be the same as at the Shield site. This requires an increase of heat production in the upper layer from 4.0 to 5.2 and again results in a temperature increase of 100°C at the mantle interface. Obviously we need to know a great deal more about both the distribution of radioactive material in the crust and about the magnitude of past surface temperature variations.

Changes of surface conditions or changes in surface temperature, as well as disturbing the temperature gradients and thus the heat flow close to the surface, may affect the mantle-crust temperatures directly. Since the Palaeozoic seas retreated 200 million years ago the region has been subjected to erosion. Hough (1963) speculated that the Great Lakes were broad river valleys until the transgression of the Pleistocene glaciers. In the light of recent structural studies, the Lake Superior basin and perhaps others may be older than formerly supposed. Mean bottom temperatures vary between 2 and 3.5°C depending on the severity of the winter. However temperatures at comparable depths in wells several miles from the shorelines lead to temperature differences of 7°C or more. If the lakes have persisted for millions of year the temperature disturbance at the base of the crust is as shown in FIG IX-7. Refraction effects on a large scale may have substantial effects on crustal temperatures, particularly in this highly variable upper layer. In the Michigan Basin these effects are probably not sizeable since the conductivity contrast between basin sediments and the basement rocks is only 10 - 15%.

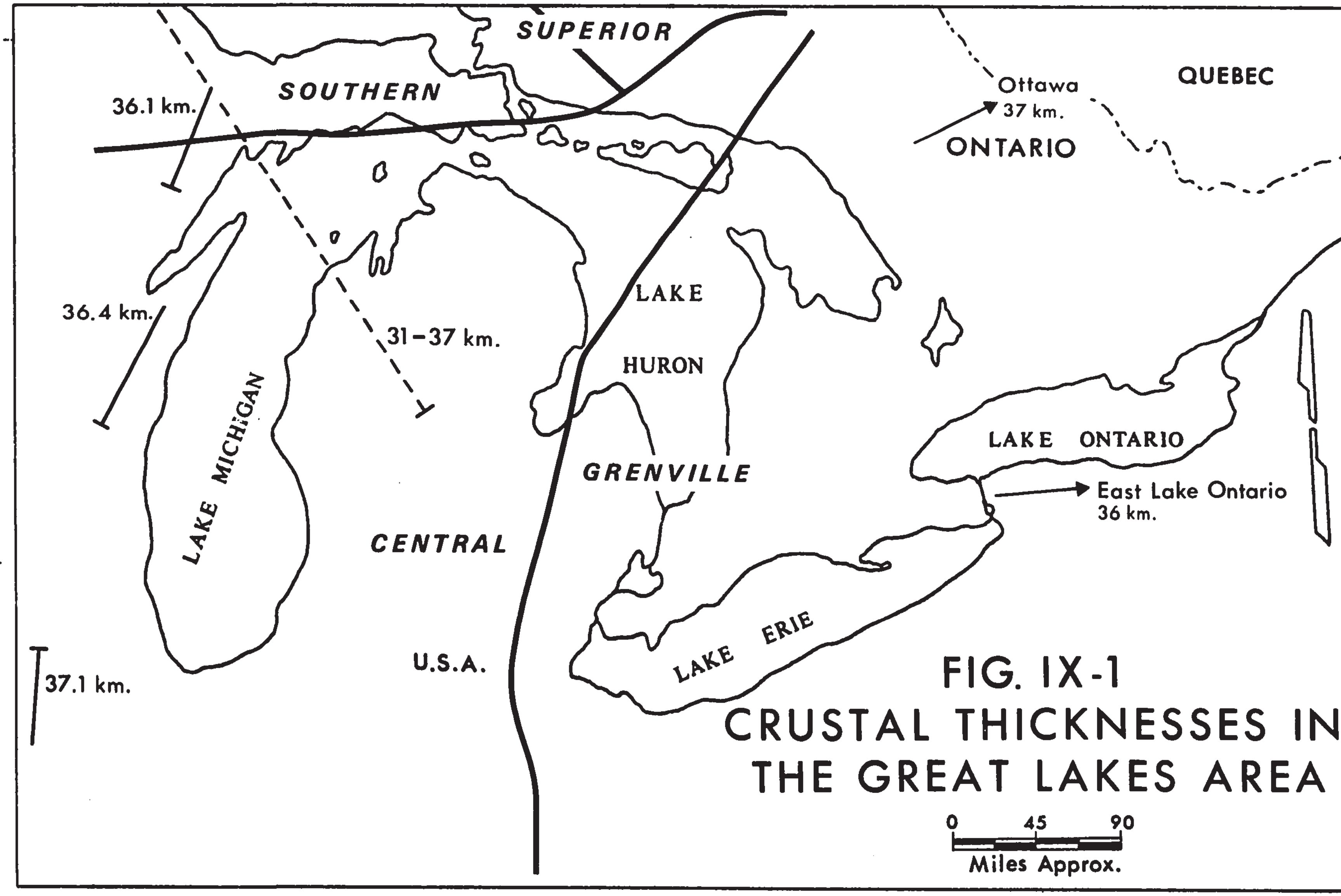


FIG. IX-1
CRUSTAL THICKNESSES IN THE GREAT LAKES AREA

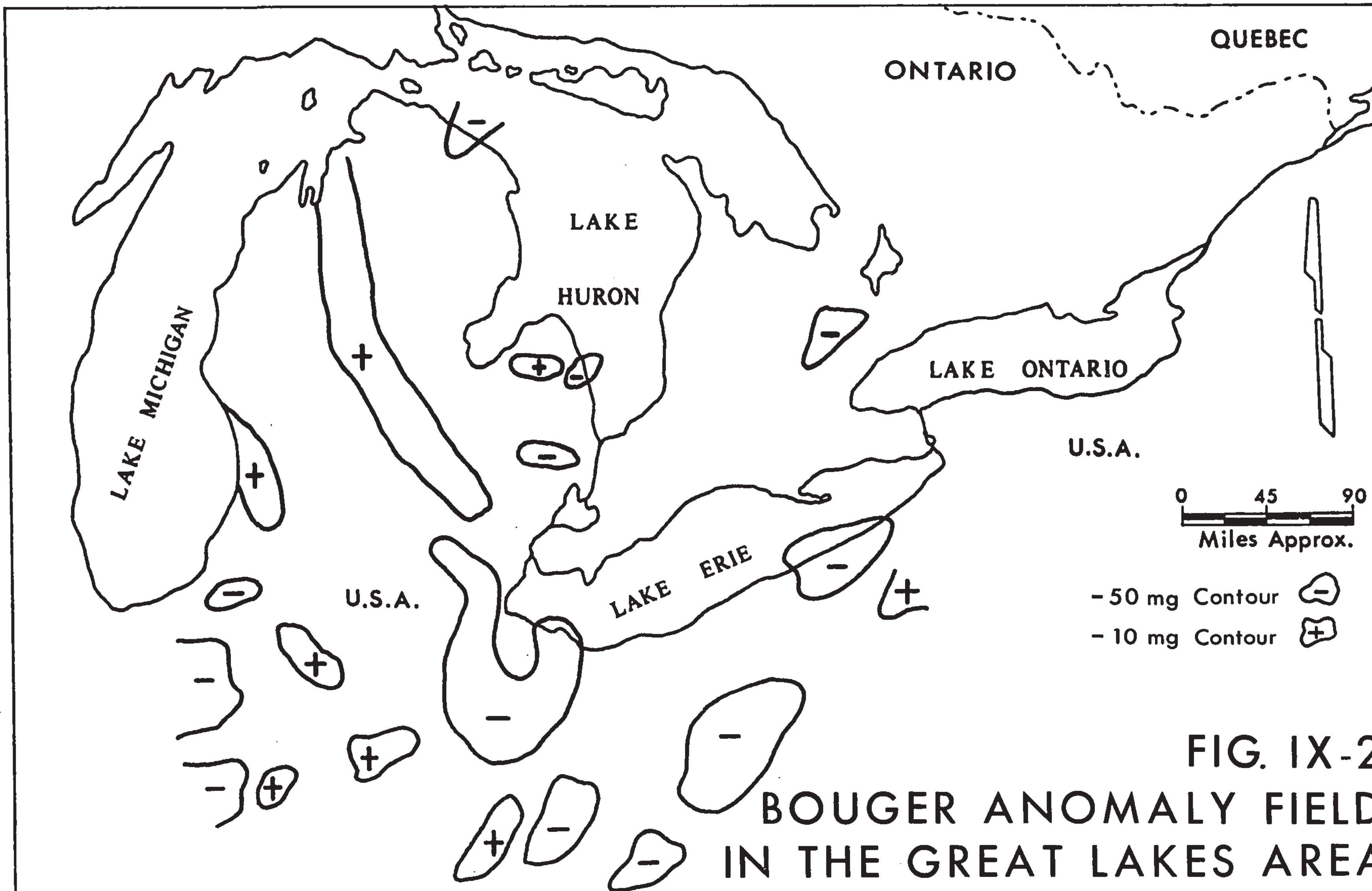


FIG. IX-2
BOUGER ANOMALY FIELD
IN THE GREAT LAKES AREA

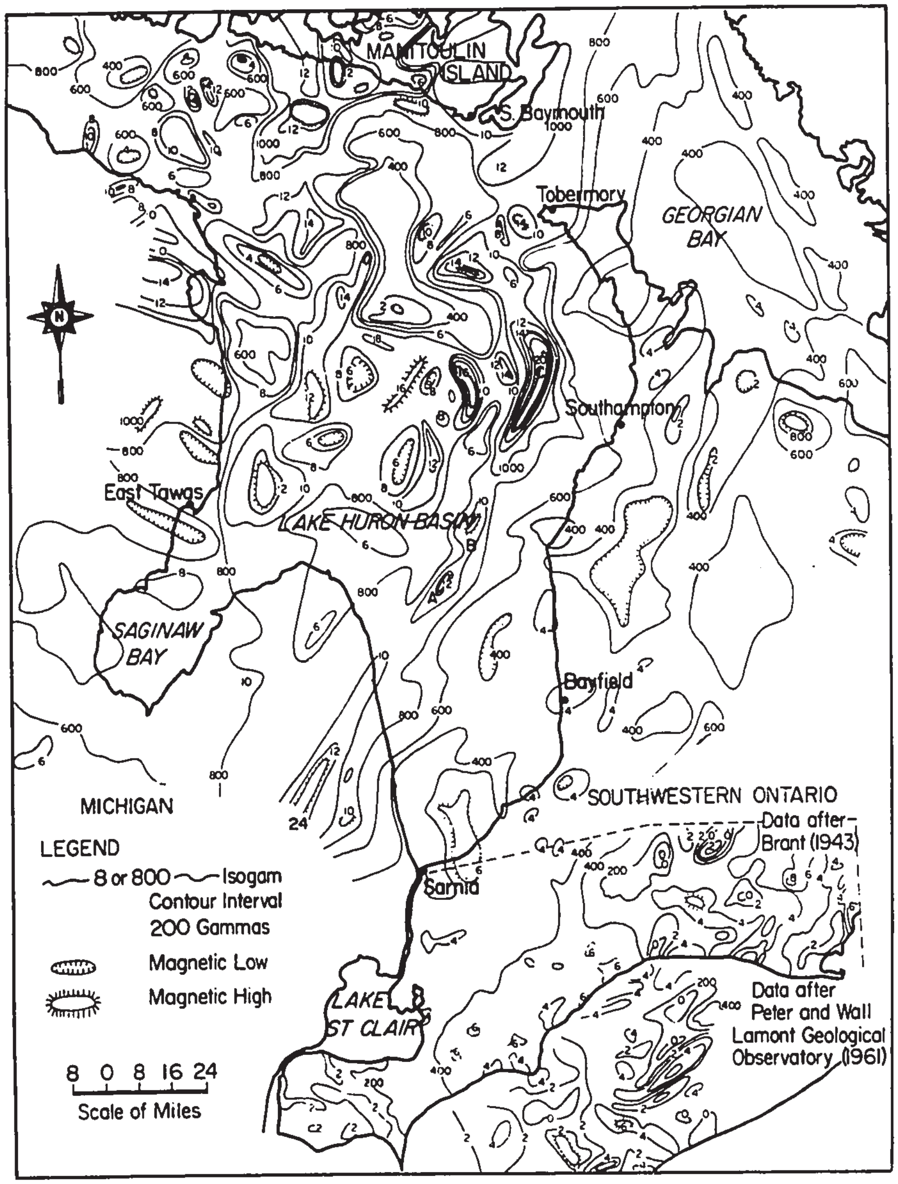


FIG. IX-3

MAGNETIC ANOMALIES IN THE GREAT LAKES REGION

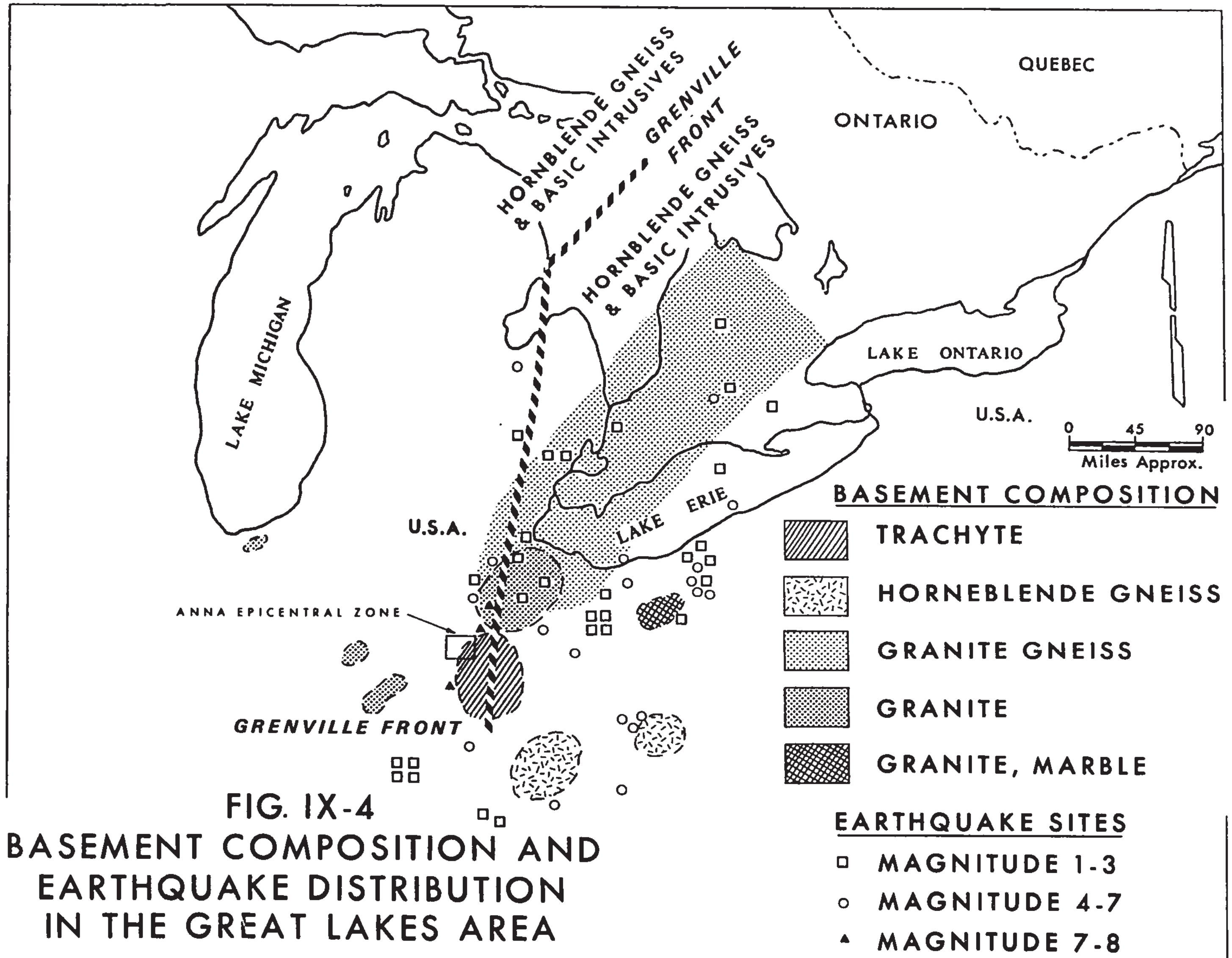


FIG. IX-4
BASEMENT COMPOSITION AND EARTHQUAKE DISTRIBUTION IN THE GREAT LAKES AREA

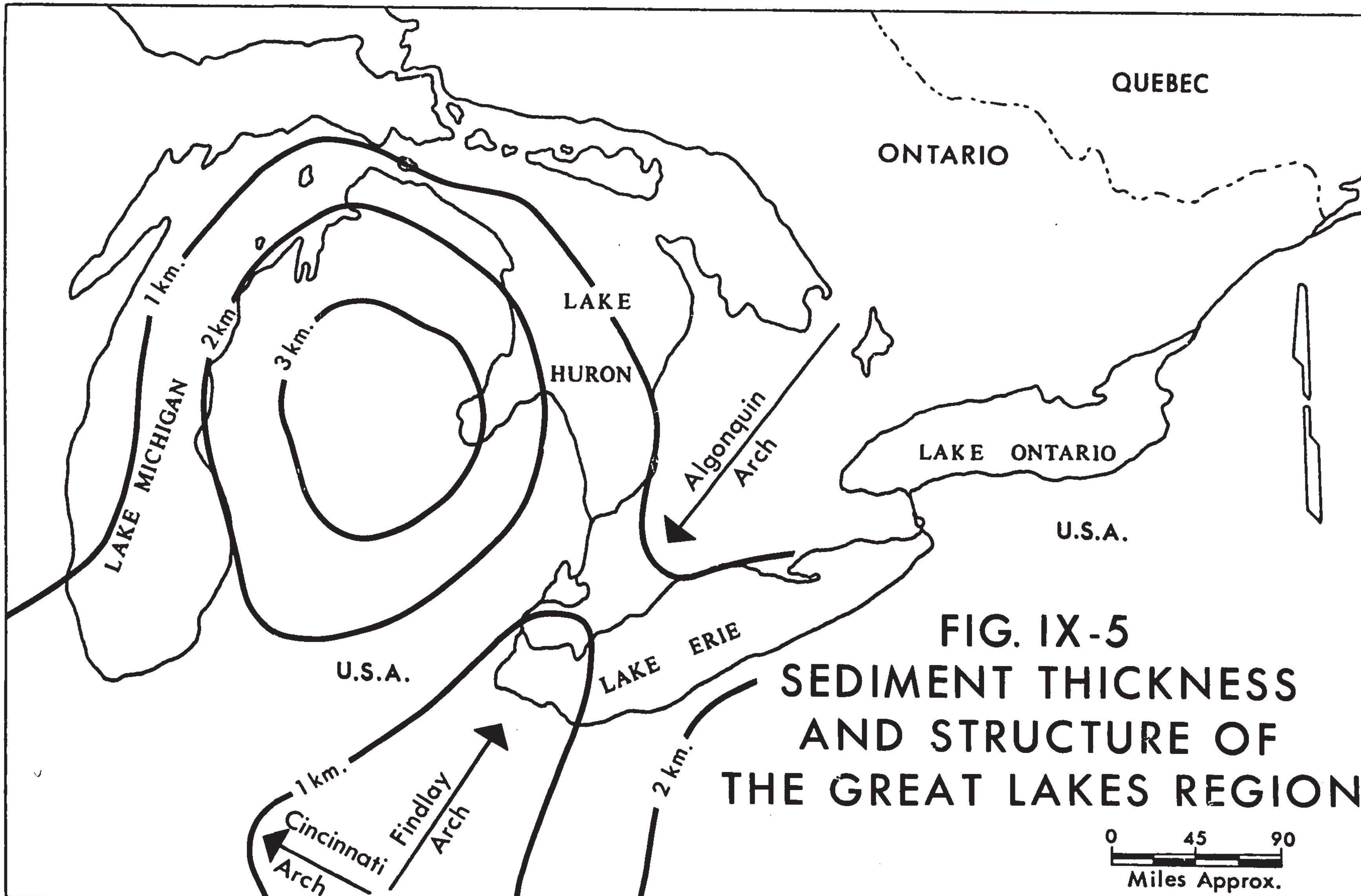


FIG. IX-5
SEDIMENT THICKNESS AND STRUCTURE OF THE GREAT LAKES REGION

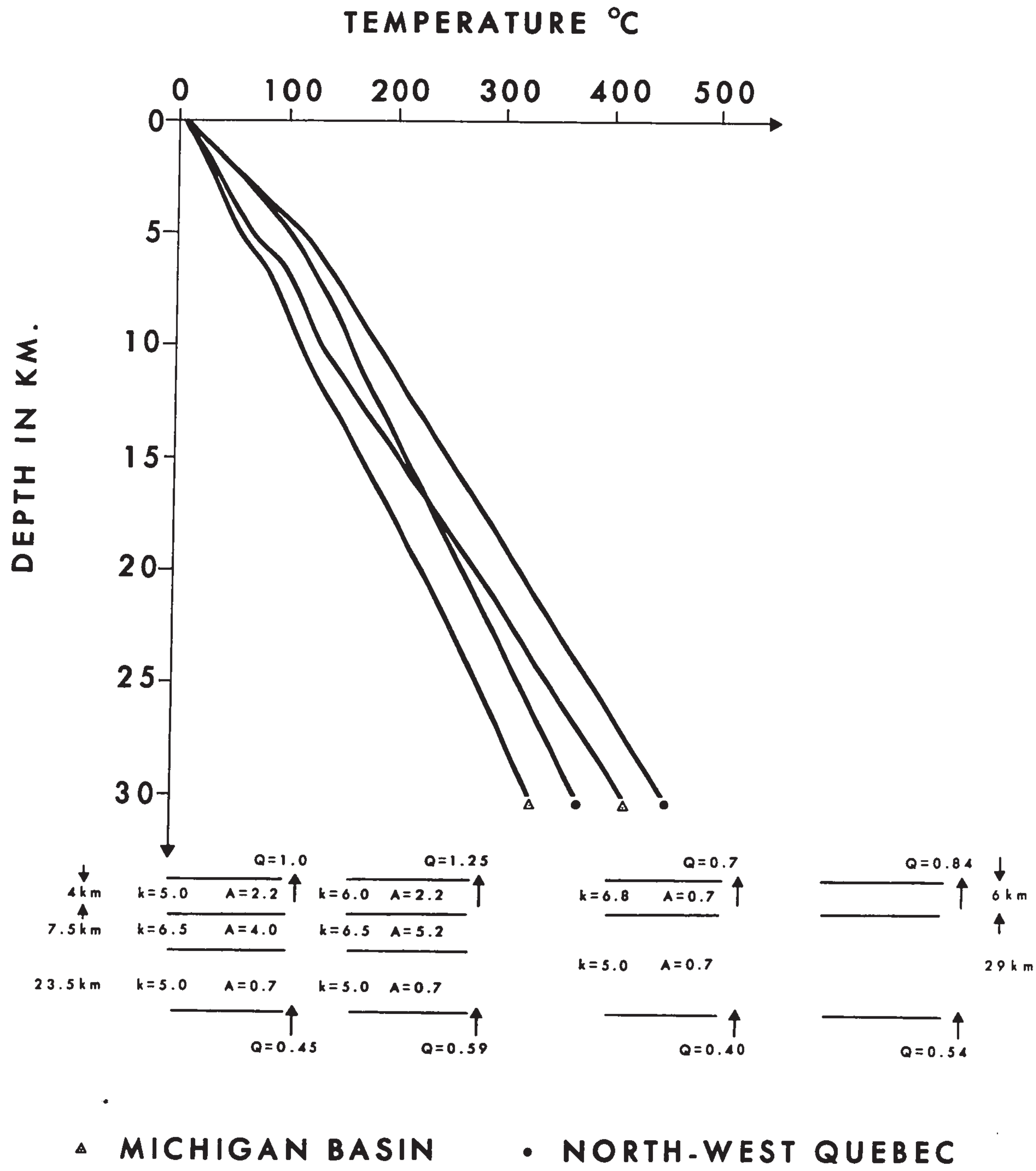


FIG. IX-6
CRUSTAL TEMPERATURE FIELD

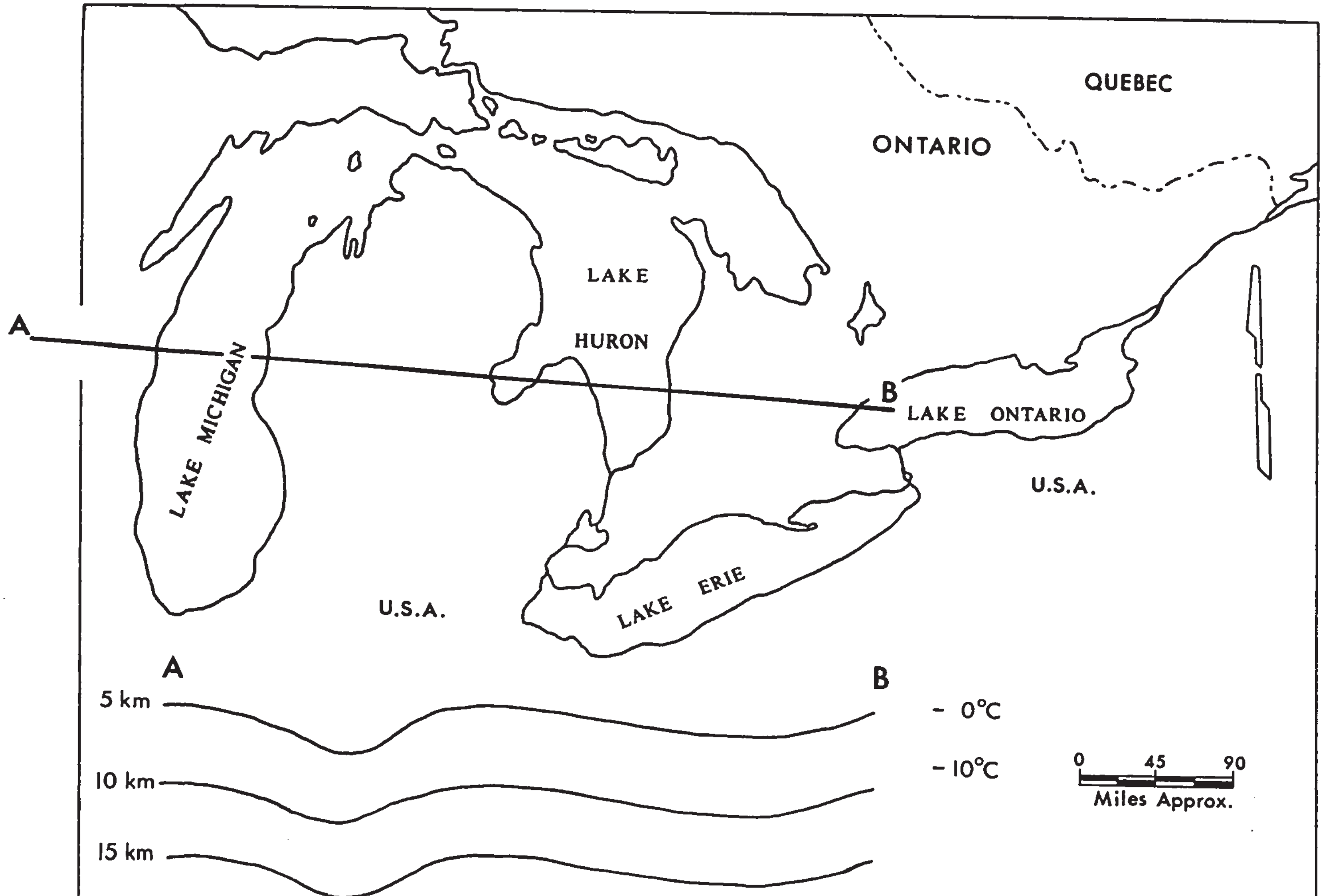


FIG. IX-7
DISTORTION OF THERMAL FIELD BY THE GREAT LAKES

CHAPTER X

DRILLING DISTURBANCE
IN THE LONDON BOREHOLE

X-1 INTRODUCTION

Determinations of terrestrial heat flux in the sedimentary regions of Canada are still fairly sparse in comparison to those on the shield. One of the major reasons for this is that national and provincial regulations require that boreholes be sealed at various levels and be plugged at the surface as part of the abandonment process. To suspend the plugging of a hole for several years and then to take a drill-rig back to the site later is prohibitively expensive for most groups doing geothermal research. Using holes that were drilled prior to the introduction of plugging programmes or which are suspended by companies pending further development is one method of obtaining geothermal data, but it is time-consuming. A method at present used by the Observatory group to obtain holes of the order of 600 m deep is to purchase this amount of casing or tubing from the operating company and have then place a cement plug at the base of this section in accordance with provincial regulations. This leaves only a surface plug to be placed at the completion of measurements.

An alternative method which would make many more boreholes available would be some method of obtaining the undisturbed geothermal gradient from temperature measurements made during halts in drilling. Most boreholes in sedimentary areas are drilled with large diameter rotary rigs using large quantities of drilling mud. Anglin & Beck (1965) have confirmed that bottom-hole temperatures measured in such holes shortly after a cessation of drilling are not reliable indicators of the equilibrium strata temperature. This is probably due to the settling of large volumes of drilling mud in the bottom of the hole. In this case the instrument does not penetrate to the base of the mud and measures only mud temperatures. Obviously in these circumstances a rather more specialised tool is required in which the weight of the drill-rods can be used to get the sensing head in contact with the rock below the mud. A tool of this kind would have obvious value, for example, in detecting permafrost thickness, information which is in desperately short supply in the north of Canada. However there is at present a great deal of interest among oil and gas exploration companies in using hard-rock diamond drill methods for cored stratigraphic test holes. The availability of these reopens the questions of either making temperature logs during breaks of several days in the drilling and/or measuring bottom-hole temperatures during shift changes.

X-2 THE BOREHOLE

One of the first holes drilled by the Dominion Observatory as a part of the Canadian Upper Mantle programme was a 600 m borehole on the campus of the University of Western Ontario. The hole was spudded on September 18th, 1963, and completed on October 23rd, 1963. Bedrock was reached at a depth of 41 m on September 22nd, below which the hole passed through Palaeozoic sediments. Core recovery in the solid rock was almost complete with the only significant loss being 3 m between 67 and 70 m. The hole was cased to 441 m with BX drill-rod inside BX casing to 34 m which was inside NX casing to 29 m and H casing to 3 m. Circulation loss occured at 55 m and 96 m where cementing was necessary. During drilling, records of input and output, water temperatures and water flow rates were kept and frequent stoppages made to make temperature measurements down the hole. Drilling was also suspended over several weekends to allow temperature measurements to be made. After completion the hole was logged at regular intervals until it had returned to equilibrium.

X-3 DRILLING HISTORY

The distance drilled on each successive day is shown in FIG X-1 commencing with the 5th day when drilling into bedrock began. Operations were suspended on the 7th, 17th, 24th and 31st days to enable temperature measurements to be made. Between these periods the drilling rate varied from little more than 20 m per day to greater than 40 m per day. There is some general agreement between drilling rate and density since denser rock is usually harder and hence drilling is slower. The drillers kept records of the times that they were drilling and pulling and noted the water meter readings when they commenced to pull. In this way it was possible to form a daily time of fluid circulation and a daily rate of fluid circulation, as shown in FIG X-2. Once drilling was under way in earnest the fluid rates were fairly constant, ie., between 27 and 36 litres/min, a fairly low rate when it is remembered that far greater mud volumes are normally circulated in rotary drilling.

X-4 TEMPERATURE MEASUREMENTS

X-4-1 During Halts in Drilling

A Bottom-Hole Temperatures

The most complete cooling record was obtained on the weekend of the 19th to 20th of October. Water circulation was stopped at 2315 hours on the night of October 19th and the rods pulled up 20 m from the bottom of the hole. The latter action was to be sure of no end-effects from the steel rods and to prevent the rods from jamming permanently. Bottom hole was at a depth of 582 m when drilling ceased for the weekend. Temperatures at the bottom of the hole were recorded with a thermistor cable as usual. They changed very rapidly initially and were recorded on a Rustrak recorder from which temperatures and rates of recovery could be interpolated. Observations were continued at intervals until the Monday morning, some 2000 minutes later. A plot of the temperatures versus the time since circulation ceased (FIG X-3) suggested an inverse relationship which is verified by the linear nature of a plot of temperature against the reciprocal of time.

This variation is consistent with the temperature decay from an instantaneous linear source. If the time of creation of a source of strength Q units of heat per unit length is taken as the time of cessation of drilling, then by Lachenbruch and Brewer (1959):-

$$\theta(t) - \theta_r = \frac{Q}{4\pi k} \cdot \frac{1}{t} \qquad \text{i)}$$

where k is the thermal conductivity of the medium, θ_r the undisturbed temperature and t the time elapsed since the drill-bit first reached the depth. However this is a very simple means of representing a finite cylinder. A further step is to consider a cylinder of radius a, in which the region r<a was initially at a temperature θ_o when the circulation ceased, and r>a at temperature θ_r. The former region can be considered to be a perfect conductor since the column of water can transfer heat by convection. Then, using notation as before, by Carslaw and Jaeger (1959):-

$$\frac{\theta(t) - \theta_r}{\theta_o - \theta_r} = F(h, \alpha, \zeta) \qquad \text{ii)}$$

where $\alpha = \frac{2\rho c}{\rho_c c_c}$, $h = \frac{k}{aH}$, $\zeta = \frac{\kappa t}{a^2}$,

ρ and c are the density and specific heat of the rock, ρ_c and c_c those of the fluid column, κ is the thermal diffusivity of the medium and H the reciprocal contact resistance per unit area between the fluid and surrounding strata.

$$F(h,\alpha,\zeta) = \frac{4\alpha}{\pi^2} \int_0^\infty \frac{\exp(-\zeta u^2)du}{u\Delta(u)}$$

where $(u) = \{uJ_o(u) - (\alpha - hu^2)J_1(u)\}^2 + \{uT_o(u) - (\alpha - hu^2)Y_1(u)\}^2$,

J_o and J_1 are Bessel Functions of the First Kind and Y_o and Y_1 of the Second Kind.

For the case of no contact resistance h=0, Jaeger (1956) lists results to $\zeta = 20$ and $\alpha = 8.0$. In the case in hand the water column gives $\alpha = 2\rho c \approx 1.0$ and ζ varies from 3.7 to 160. Now Blackwell (1954) has shown that for large ζ:-

$$F(h,\alpha,\zeta) \approx \frac{1}{2\alpha\zeta} + \frac{(4h - \alpha)}{4\alpha^2\zeta^2} - \frac{(\alpha - 2)}{4\alpha^2\zeta^2}\left\{\ln\frac{4}{1.781} - 1\right\}$$

which on putting h = 0 becomes:-

$$F(0,\alpha,\zeta) \approx \frac{1}{2\alpha\zeta} + \frac{1}{4\alpha\zeta^2}\left\{\ln\frac{4\zeta}{1.781} - 2\right\}$$

If $\zeta > 1$, then:- $F(0,\alpha,\zeta) \approx \frac{a^2}{4k} \cdot \frac{1}{t}$ . . . . iii)

As can be seen from the above, at long time intervals the function behaves in a similar fashion to the instantaneous source solution, and is fairly consistent with the relationship suggested by the data. However a closer look at the data shows a few interesting features. It can be seen in FIG X-4 that the linear plot can be subdivided into two straight lines with the same gradient but different intercepts. The changeover time occurs between 70 and 80 min. after the cessation of drilling and represents a systematic change of 0.035°C in the measured temperatures. Whether or not this merely represents a change in thermistor or cable characteristics is difficult to ascertain since such problems were encountered during many of the measurements at that time. At very long times it is apparent that the inverse relationship no longer holds and the return to equilibrium appears to be faster than that predicted by these models. If we assume that these models do hold through the range, then a time for return to equilibrium can be predicted which will be slightly longer than the true one, and thus has a built-in safety factor. Six hours after the drilling has ceased about 1.6% of the temperature disturbance at bottom hole remains and after sixty hours this is reduced by a factor of ten, which in this instance means that, after six and sixty hours respectively, the temperature should be within 0.1°C and 0.01°C of the final temperature. In fact however, as mentioned above the extrapolated intercept temperatures are too low. Taking the curve for a period of less than 80 minutes, the results are too low by 0.1°C and for the period 80 to 170 minutes by 0.05°C when compared with the temperatures recorded on the final log in 1967. If the curvature is taken into account at long times, the extrapolated temperature is 0.03°C lower than the final temperature. This difference may represent a combination of depth and calibration differences between measurements made in 1963 and those made in 1967.

Several bottom-hole temperature measurements were also made during the period of cessation of drilling from the 4th to the 6th of October at a depth of 341 m. Although data was only recorded at two intervals it is apparent that, since the temperature at the bottom is within 0.05°C of the equilibrium temperature after a period of 36 hours, the return is again very rapid.

On frequent occasions when the rods were pulled, temperature measurements were made down the borehole and readings taken at the bottom. These however have little value since, although the rate of cooling and the time are known, these factors need to be defined in several points before any useful calculations can be made.

B <u>Temperatures Along the Hole</u>

On the weekend of October 19th and 20th the hole was completely logged on three separate occasions in addition to the bottom-hole measurements. If these temperatures, made at 16, 37 and 57 hours after cessation of drilling, are plotted together with the equilibrium temperatures made in 1967 (FIG X-5), it is very striking to note that, for depths below 425 m and more than 30 m above the bit, the temperature gradient measured 57 hours after cessation of drilling is within a few per cent of the equilibrium gradient. Yet these depths represent times of active circulation of 0.4 to 7.5 days, since the bit first reached a particular depth. The rocks penetrated in this interval consist of low permeability red shales of the Queenston formation. The upper section of the hole, reflecting source times of 8.8 to 22.6 days, is slower approaching equilibrium. To some extent this might be due to the more permeable rocks acting as a larger source volume, but this is discussed later. It is also worth noting that above 366 m the temperature gradient increases with time in the return to equilibrium whereas below this depth it decreases.

Depth (m)	Temperature Gradient (^{o}C/ 30 metres)				Predicted Equil.
	16hrs	37hrs	57hrs	Equil.	
152					
	.083	.130	.150	.182	.39
183					
	.081	.091	.095	.198	.14
213					
	.145	.194	.235	.333	.39
244					
	.117	.135	.145	.200	.20
274					
	.161	.187	.200	.237	.27
305					
	.155	.159	.167	.195	.21
335					
	.166	.163	.163	.158	.20
366					
	.248	.219	.210	.184	.12
396					
	.389	.316	.285	.210	.16
427					
	.564	.547	.550	.563	.56
457					
	.476	.466	.437	.444	.43
488					
	.525	.488	.500	.474	.49
518					
	.623	.484	.438	.485	.33
549					
	.407	.150	.305	.461	
579					

↑ Gradient increases with time

↓ Gradient decreases with time

Table X-1 RETURN TO EQUILIBRIUM IN LONDON BOREHOLE OVER WEEKEND

In treating the problem of the return to equilibrium with a mathematical solution, we must consider sources created at some instant in time and then study the decay. At depths above bottom-hole the drilling fluid has circulated for fairly considerable lengths of time. Thus it is necessary to take the instantaneous source solution and integrate it over the time of persistence. Thus from i):-

$$\theta(t) - \theta_r = \frac{1}{4\pi k} \int_0^s \frac{q}{t - \zeta} \cdot d\zeta, \qquad t > s \qquad . \quad . \quad . \quad . \quad iv)$$

where s represents the time elapsed from the day the bit first reached the depth in question until the operation ceased. If the source is treated as a constant for this period then:

$$\theta(t) - \theta_r = \frac{q}{4\pi k} \log_e \frac{t}{t - s} \qquad \ldots\ldots\ldots \quad v)$$

This solution is not strictly correct since the diameter of the well is finite and the source strength is not strictly constant through the drilling period. If this equation is used to force a least squares fit through the temperature data obtained at each depth, it should be possible to estimate the equilibrium temperature θ_r and the source strength q, if we put in the average thermal conductivities at each depth. The numerical results are shown in the table below and FIG X-5 shows the measured temperatures, the calculated equilibrium temperatures and the equilibrium temperatures measured in 1967. Examining the results, the predicted equilibrium temperatures are within 0.1^{o}C of the final ones over the entire range of 183 to 518 m but the result deviates seriously at 549 m. It must be recalled that bottom-hole at this time was 583 m. Over the interval 305 to 396 m the prediction is better than 0.02^{o}C. In general the final predicted equilibrium temperature is less than measured. A very apparent feature of the curves is that the gradient of the temperature curve after 57 hours and below 427 m is as good a predictor of the final gradient as the gradient of the predicted equilibrium temperature curve.

From the least squares data it was also possible to estimate the source strength q at each depth and from this make an estimate of the heat transfer coefficient.

DEPTH m	TEMP. 66/67 °C	T (∞) calc. °C	$\frac{q}{4\pi k}$	$q \times 10^{-3}$ cals cm^{-3}
152	9.38	9.19	.305	32.
183	9.57	9.50	.260	29.
213	9.75	9.64	.248	21.
244	10.09	10.03	.175	15.
274	10.29	10.23	.151	25.
305	10.52	10.50	.118	19.
335	10.72	10.71	.100	15.
366	10.88	10.91	.113	17.
396	11.06	11.03	.150	21.
427	11.27	11.19	.249	17.
457	11.83	11.75	.276	19.
488	12.28	12.18	.325	18.
518	12.76	12.67	.392	22.
549	13.24	13.00	.661	45.

Table X-2 SOURCE STRENGTHS AND PREDICTED EQUILIBRIUM TEMPERATURES DURING WEEKEND DRILLING HALT

The entire hole was warmed by the drilling fluid. There was some variation in the temperature of the drilling fluid but an average value for these calculations of 14.4°C is used. Using equation v) for the return to equilibrium, the predicted heat released and equilibrium temperatures are calculated, again for the weekend of the 19th and 20th of October, as shown in the table above. Once the source strength is calculated then the heat transfer coefficient is given by:

$$\lambda = q/\pi d \Delta T \qquad \text{vi)}$$

where d is the diameter of the borehole and ΔT the temperature difference between the borehole fluid and that of the undisturbed rock. This gives a very coarse measurement of the coefficient since some heat is lost vertically as well as radially. There is also much uncertainty as to the value to adopt for d since this is really the diameter of the source rather than that of the borehole. A rough calculation based on an average loss of 20% of the fluid over the entire drilling period indicates a total source radius of 14 cm. compared with a borehole radius of 4.5 cm. Where zones of low permeability are encountered this figure will be smaller and closer to the actual borehole diameter. Thus it is not possible to make more than an order of magnitude calculation. The average value of the coefficient is about .0007 ± .0002 cal/cm^{2}sec$^{\circ}$C. Variation is due to combinations of porosity, permeability, temperature gradients, which affect the heat source and its diameter, and the heat exchange system of the drill-rods. Without more measurements of the coefficient it is difficult to say much about it except that it is small.

A conclusion to be drawn from the above sections is that, with care, some reasonable values of temperature gradients may be measured at the end of a weekend halt in drilling and, what is perhaps more useful, very accurate predictions made of bottom-hole temperatures.

It is not always possible to close down a rig for an entire weekend and thus it is also useful to look at temperature logs made during shift changes. Generally temperatures at any depth were still changing linearly over the 15 or 20 minutes of observation and thus it is difficult to extrapolate these results to equilibrium. A knowledge of the time of cessation of circulation, the time of measurement, the souce strength and persistance allows a rough estimation of bottom-hole temperature. The errors however are about 20% of the total temperature decay to equilibrium and thus the method is not recommended. Preferably the return to equilibrium should be followed for a period of several hours.

Plotting the temperatures measured along the borehole during these brief periods shows up very well the reduced disturbance close to the bottom of the hole; a selection of these are reproduced in FIG X-6 to illustrate this effect.

An important feature of these and other weekend logs obtained is the temperature inversion that develops just above bottom-hole. For example in the measurements of the weekend of the 4^{th} and 5^{th} of October, the temperature had reached a maximum about 24 m above bottom-hole after 19 hours and this still persisted 16 hours later with the inversion pivoted about bottom-hole temperature and decreasing in size with time. The temperature gradient near the bottom was $-0.5^{o}C/30m$ after 19 hours. Thus large errors could result from the probe not being quite at the bottom of the hole.

X-4-2 After Completion of Drilling

The borehole was completed on the 21^{st} of October, was flushed with water on the 22^{nd}, with water circulated for an entire shift. On the 23^{rd} the rods were pulled back to 442m and set in place as casing. Observations were then commenced on the long term recovery of the borehole to equilibrium, and continued intermittantly for a period of 4 years. During the early part of this return a great deal of trouble was encountered in the construction of probes that did not leak or show changes in characteristics with time. However there are sufficient results to enable an analysis of the return to equilibrium to be made. FIG X-7 shows

temperature measurements made between 5 and 36 days after completion of drilling compared with equilibrium logs made in November 1966 and July 1967. The coordinates are plotted to represent temperature versus $\ln(\frac{t}{t-s})$ as suggested by equation v). A linear relationship passing through the equilibrium temperature verifies the equation as representing the return to equilibrium. It is interesting that the relationship is followed to a depth of 366 m but below this depth the equilibrium temperature is consistently higher than the extrapolated equilibrium temperature. Since the sets of logs taken in 1966 and 1967 agree to better than .01°C there is no reason to doubt their validity. However the same cannot be said of some of the earlier logs and thus the possible suggestion of a more complex return to equilibrium in the lower and uncased section is not proved. The differences gradually increase between 427 and 579 m to .06°C yet a log made in November 1965 lies on the usual equilibrium decay curve. One explanation which seems to remain and which involves measurement technique rather than a physical mixing process is a switch in cable types used for logging. The discrepancy may be caused by errors in stretching corrections for the cables. Since it only occurs in the lower section this is the most likely explanation, although it is also true that it only occurs downhole from the section in which a horizontal water flow was postulated to exist (see XI-9). Accepting the former explanation it is apparent that, as was shown for a weekend stoppage, the return to equilibrium is faster in the lower sections of the borehole where the source persistance time is much shorter. In FIG X-7 the t=3s line is also shown, since this represents the time after which the temperature should be within .05°C of the equilibrium temperatures.
Of note is that in the lower sections of a diamond-drill hole this represents only one month after borehole completion and the complete hole fits this

condition within three months of completion. As has been shown previously bottom-hole temperatures were within .05°C of the final temperature after only 5 days.

Of more concern in heat-flow measurements are the temperature gradients rather than the absolute temperatures. Here more data can be used since several of the temperature logs showed obvious drifts in the absolute temperatures recorded although not in the gradients since the probes and cables had not leaked. It is apparent from the table below that the gradients are constant to within a few percent after the 10th of November 1963 log, or 18 days after completion of the well.

DEPTH(m)	TEMPERATURE GRADIENTS AFTER COMPLETION (°C/30m)							
	1963					1964	Means	1966/67
	28/10	30/10	10/11	28/11	/12	/6		
91	.09	.12	.11	.12	.11		.11	.12
122	.07	.10	.12	.14	.11		.11	.17
152	.14	.19	.21	.18	.19		.18	.19
183	.20	.13	.14	.16	.13	.14	.15	.19
213	.22	.27	.31	.32	.32	.34	.30	.24
244	.22	.20	.16	.20	.17	.19	.19	.20
274	.19	.27	.24	.22	.22	.24	.23	.23
305	.18	.12	.18	.19	.23	.19	.18	.19
335	.17	.16	.17	.16	.15	.15	.16	.17
366	.18	.18	.17	.17	.19	.17	.18	.18
396	.33	.23	.21	.20	.20	.22	.23	.21
427	.40	.53		.52	.61	.54	.52	.56
457	.47	.44		.45	.42	.44	.44	.45
488	.47	.46		.52	.55	.47	.49	.48
518	.46	.45		.47	.47	.46	.46	.48
549	.48							.46
579								

Table X-3 TEMPERATURE GRADIENTS MEASURED AFTER COMPLETION OF BOREHOLE

The implication is that whilst absolute temperatures cannot be accurately measured until three months after completion, accurate gradients may be obtained after only one month. Comparison of these mean gradients with the 1966/67 logs agree well in the lower portion of the hole. In the upper section discrepancies occur in two intervals and are probably related to the zones of cementing.

Calculations of the heat transfer coefficient using the return to equilibrium data obtained after completion of the borehole yield results of the same magnitude as those obtained during drilling pauses. However the source strengths and the coefficients are slightly lower. This is particularly apparent close to the top of the hole.

DEPTH(m)	TEMP(^{o}C) 66/67	T(∞)(^{o}C) calc.	$\frac{q}{4\pi k}$	$q\times10^{-3}$ (cal.cm^{-3})
91	9.09	9.08	.248	23.
122	9.21	9.21	.246	18.
152	9.38	9.37	.209	22.
183	9.57	9.57	.208	22.
213	9.75	9.74	.193	18.
244	10.09	10.09	.137	9.
274	10.29	10.26	.150	27.
305	10.52	10.50	.137	24.
335	10.72	10.69	.141	19.
366	10.88	10.87	.128	18.
396	11.06	11.02	.164	22.
427	11.27	11.22	.230	16.

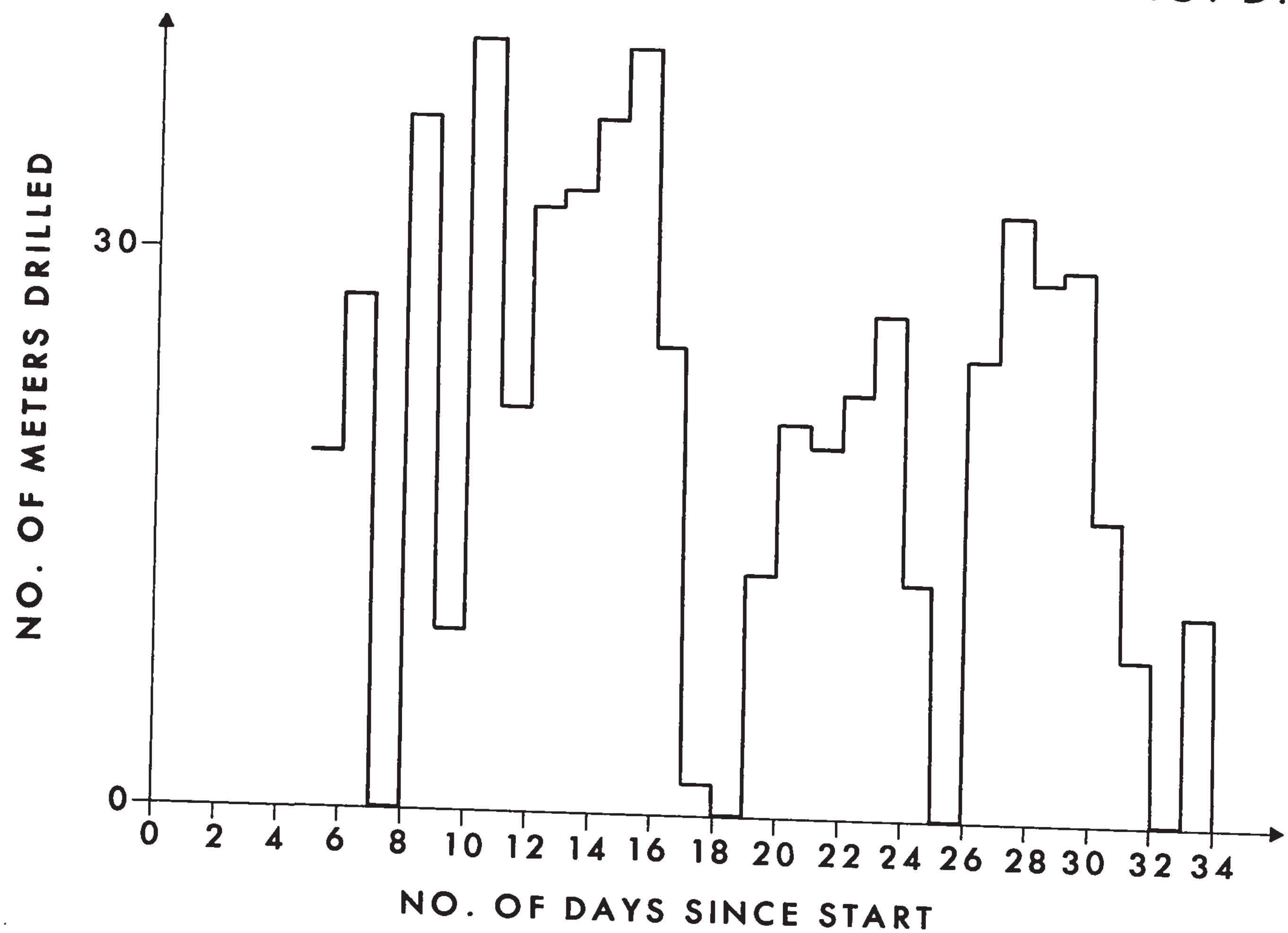
Table X-4 SOURCE STRENGTHS AND PREDICTED EQUILIBRIUM TEMPERATURES AFTER COMPLETION OF BOREHOLE

The predicted equilibrium temperatures, shown together with the souce strength in the table above again show that whilst the predicted equilibrium temperatures are very close to the final temperatures in the top 305 m, the predictions suggest too low a temperature at greater depth. The reason for this is probably in the corrections for cable stretch between two different types of cable.

X-5 CONCLUSIONS

Many of the estimates of the return to equilibrium of boreholes have tended to discourage the taking of measurements either during breaks in drilling or immediately after completion and thus many possible heat-flow sites have been missed. This analysis suggests that many of the slim-holes currently being drilled by diamond-drill rigs in sedimentary regions might adequately be used as heat-flow sites merely by making temperature logs when breaks of one or two days occur during drilling or at completion while other tests are being conducted prior to final abandonment. A similar argument applies to cable-tool holes in which, as in diamond-drilling, low fluid circulation rates are used.

FIG. X-1
DISTANCE DRILLED EACH DAY IN U.W.O.-D.O. D.D.H. #1



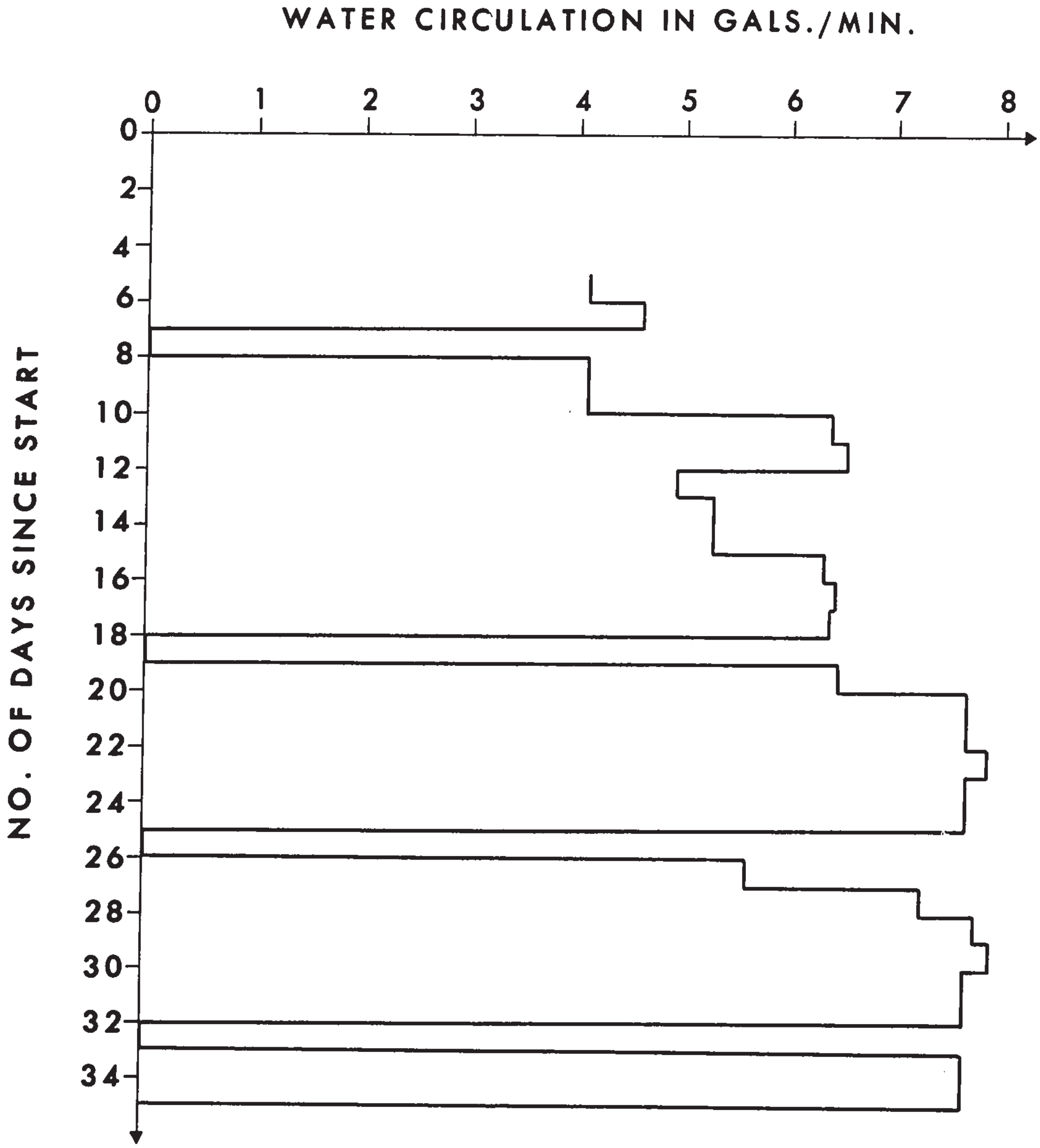


FIG. X-2
RATE OF FLUID CIRCULATION

FIG. X-3 RETURN TO EQUILIBRIUM AT BOTTOM OF DRILL-HOLE

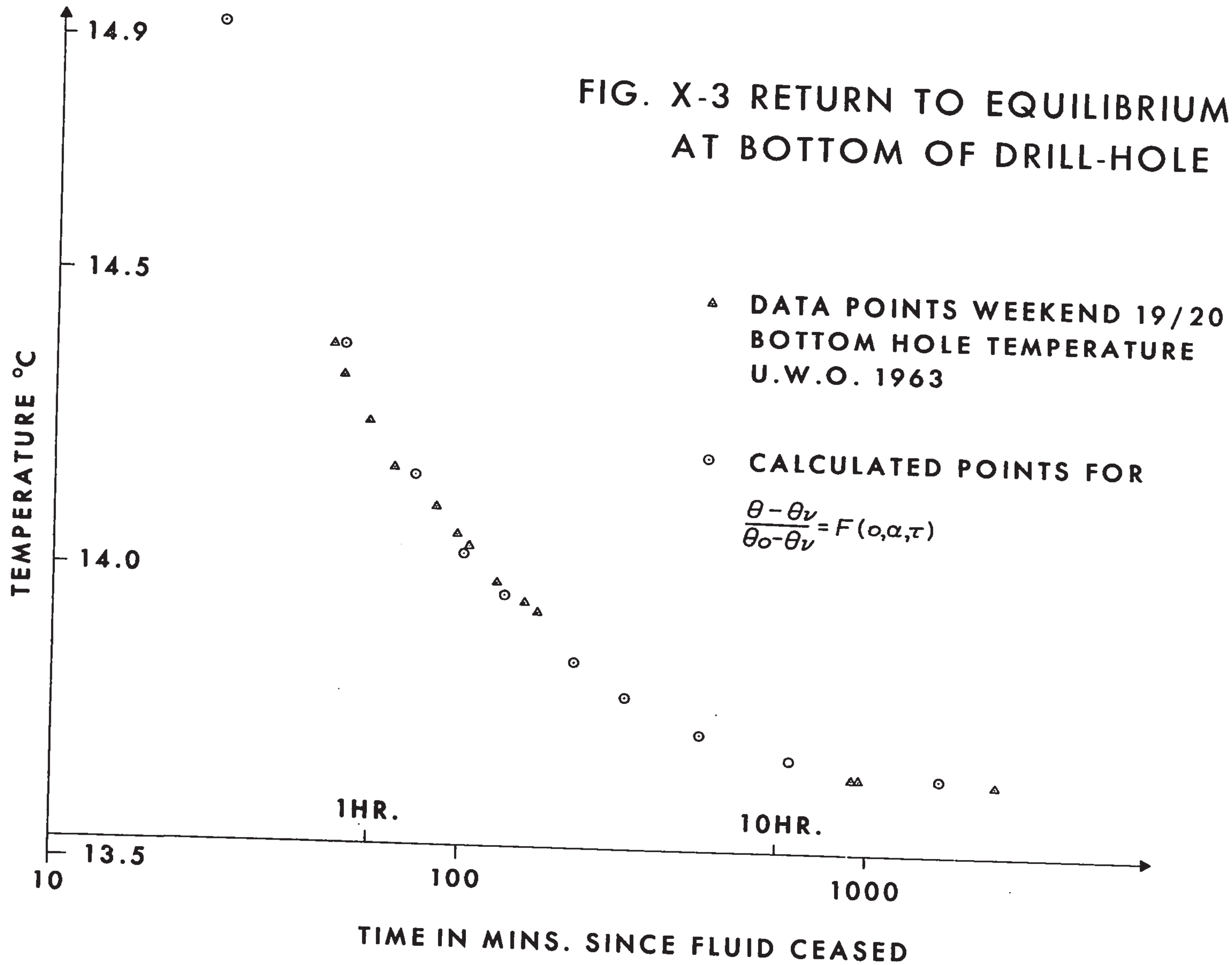


FIG. X-4 RETURN TO EQUILIBRIUM AT BOTTOM HOLE

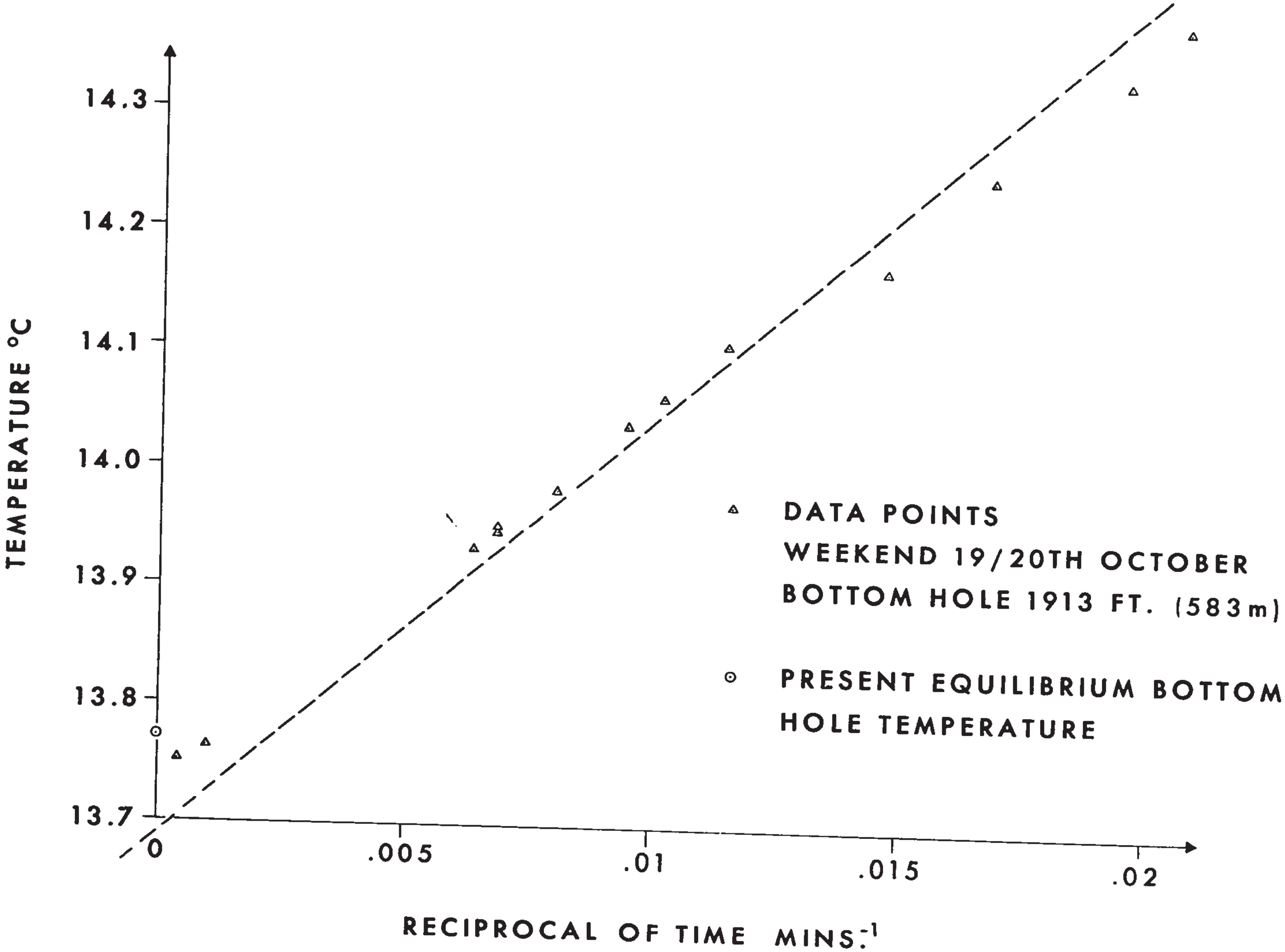
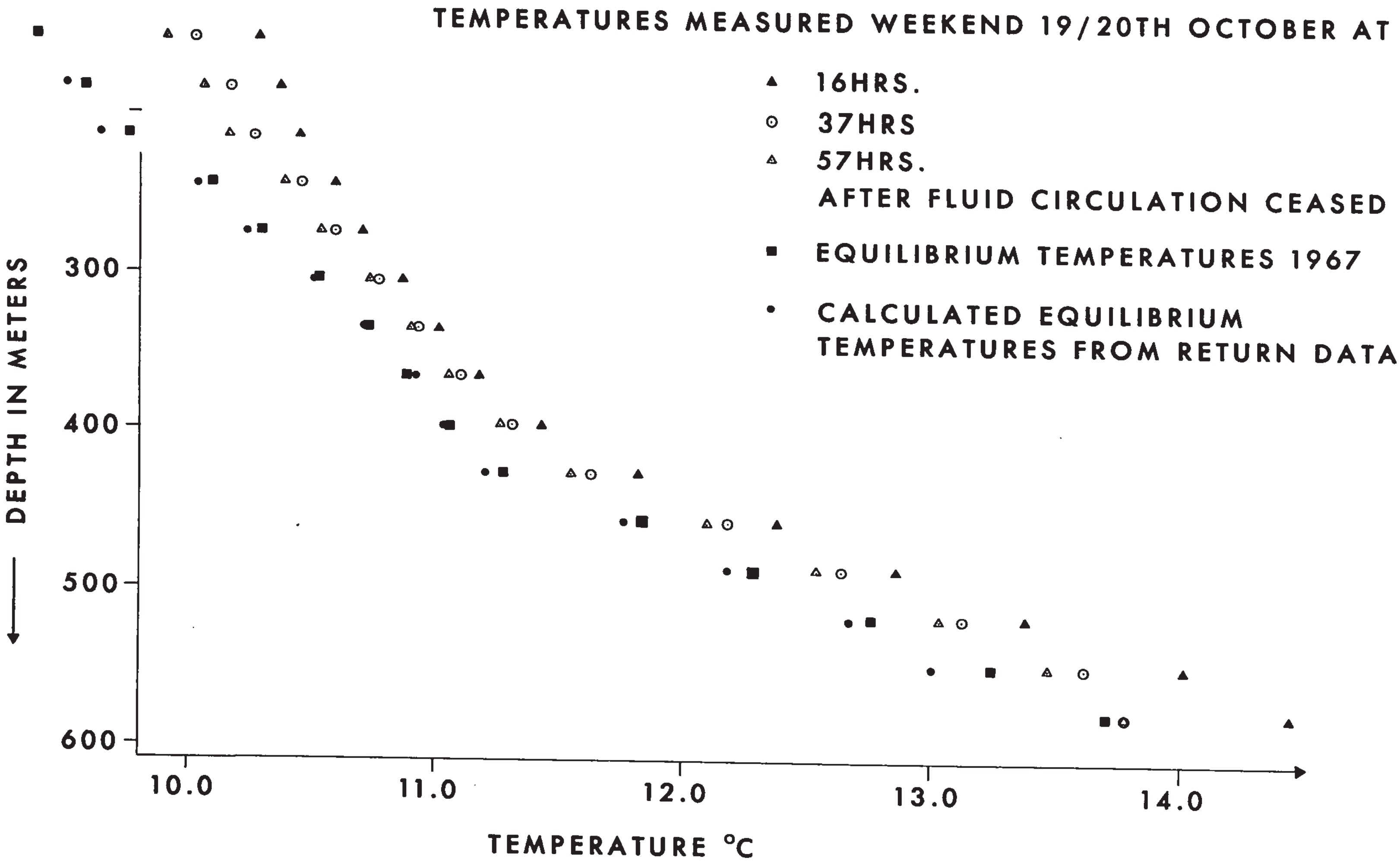


FIG. X-5 RETURN TO EQUILIBRIUM OF COMPLETE BOREHOLE



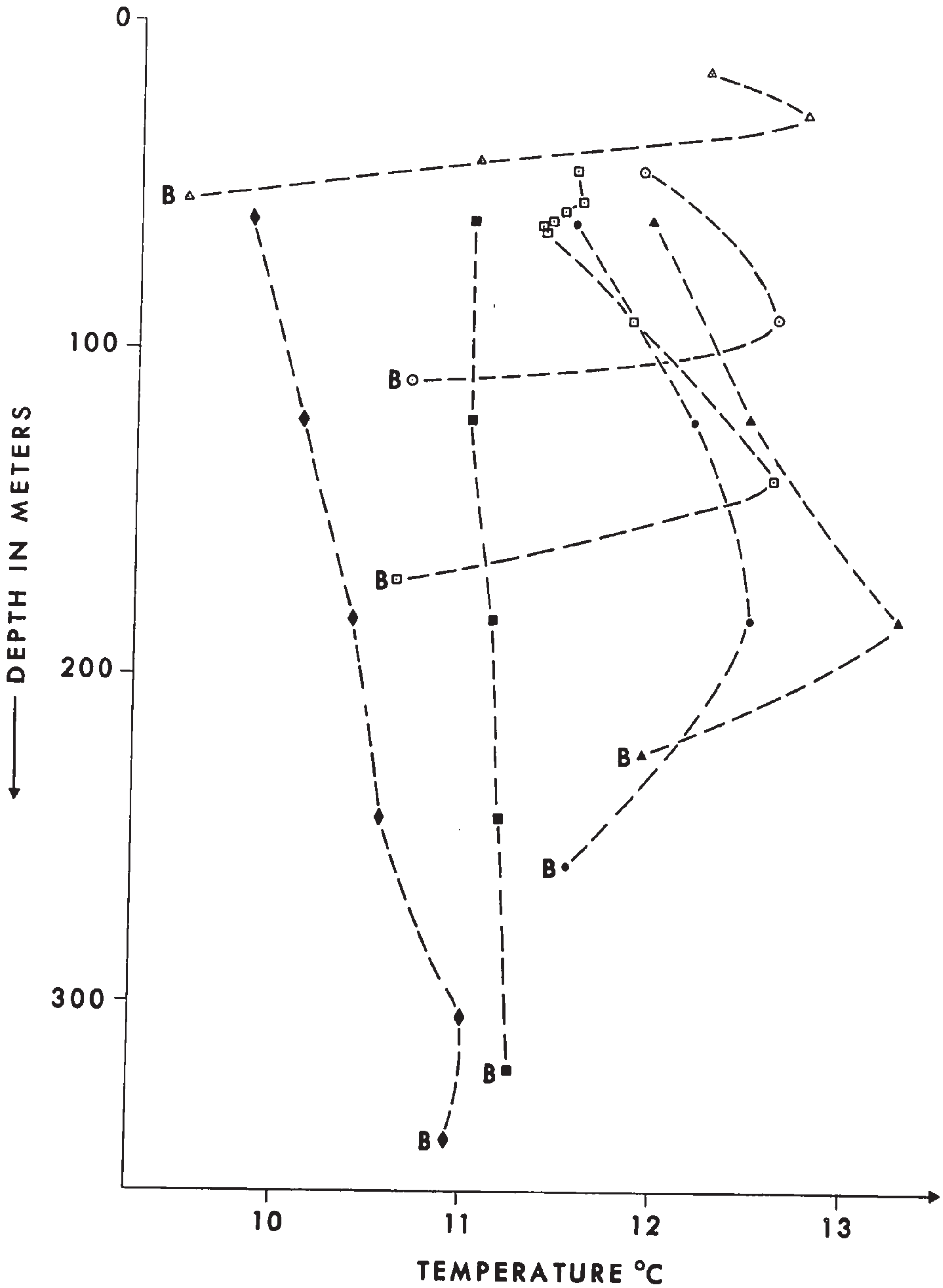


FIG. X-6
INVERSE TEMPERATURE GRADIENTS
IMMEDIATELY AFTER STOPPING FLUID CIRCULATION

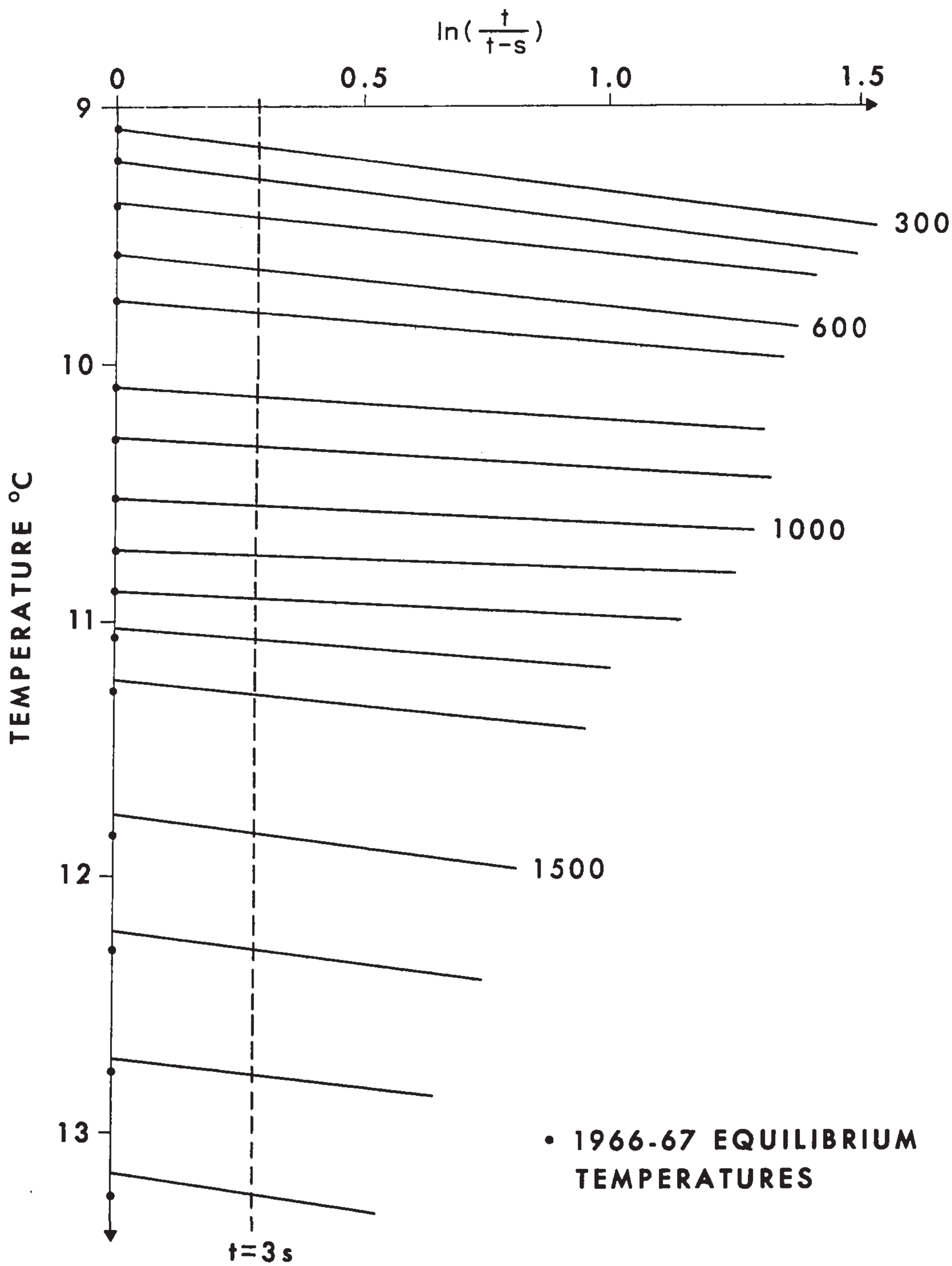


FIG. X-7
LONG-TERM RETURN OF
LONDON HOLE TO EQUILIBRIUM

CHAPTER XI

HEAT FLOW MEASUREMENTS

IN THE LONDON BOREHOLE

PREFACE

The previous chapter discussed the drilling history and the return to equilibrium of the borehole drilled on the University campus. This chapter considers the equilibrium temperatures once the drilling disturbance had been dissipated and attempts to determine the terrestrial heat flux along the borehole by a variety of methods, to compare the results obtained and then to explain the observed variations in terms of sampling and measurement accuracy, of local variations of structure, of possible chemical reactions, of climatic history and of water movements.

XI-1 EQUILIBRIUM TEMPERATURE MEASUREMENTS

Equilibrium temperature curves are taken by comparing the logs of November 1965 at 6 m intervals, November 1966 at 3 m intervals and June 1967 at 3 m intervals. Comparison suggests that the borehole has been in equilibrium below 125 m since 1965 and below 50 m since 1966. In the last log more careful measurements were made of cable stretch and thus it is believed that combined temperature and depth errors are equivalent to no more than $\pm 0.003^{\circ}C$ or for precise measurements of depth this would represent less than 30 cm depth error in sections with the lowest gradient. However these temperature errors in the sections of low gradient represent some 20% of the temperature gradient. Thus the gradient curve might be expected to be 'noisy' in these intervals. For this reason the curves have been smoothed in units of 3, 5, 10, 15, 20 etc. A cursary look at the temperature curve in FIG XI-1 reveals some interesting features as follows: a near surface temperature inversion with a minimum at a depth of 80 m.

The significance of this inversion is dealt with in Chapter XII-5. Below this the temperatures increase with depth, to 10.5°C at a depth of 300 m and 13.9°C at a depth of 600 m. The temperature gradients shown in FIG XI-1 are positive below 80 m and are divisable into two major sections, above and below 421 m. In the upper section the highest gradient, of 12°C/km occurs in the interval 213 to 232 m and in the lower just below 436 m where it rises to 22°C/km coinciding with the bottom of the cased section. Average gradients over the upper section are 6.8°C/km and in the lower 15.6°C/km. Regional gradients and temperatures are compared in Chapter VI-2-2.

XI-2 LITHOLOGY OF BOREHOLE

A lithological log of the borehole was prepared by B.V. Sanford and W.E. Koepke of the Geological Survey of Canada. The essential features are shown in FIG XI-2. Essentially the borehole contains Palaeozoic sediments commencing at 41 m with the Devonian-Delaware formation and finishing at the top of the Ordovician Meaford-Dundas formation. The regional dip of the formations is less than 2°. The rock types vary from the shales of the Meaford-Dundas and Queenston to the dolomites of the Guelph, with some formations such as the Clinton-Cataract containing mixtures of shale, limestones and dolomites. A more detailed lithology is given in Chapter VII-2, as each formation is dealt with in determining the thermal conductivity. Each position of a thermal conductivity measurement is shown on FIG XI-2 by an asterisk.

XI-3 THERMAL CONDUCTIVITY, POROSITY AND DENSITY

The three physical parameters measured on the core during this work are compared in FIG XI-3. Thermal conductivities range from 3.0 to greater than 14.0 mcal/cm.sec.^{o}C, with a mean of 8.3. Densities range between 2.20 and 2.94 gm/cc and the porosity variation is 0.1% to 14.4% by volume. Mean values for all samples are 2.58 gm/cc and 4.4%. For each formation penetrated, the means, standard variation and rock types are listed in the table below:-

Formation	No. of Samples	k	ρ	σ	Rock Types
Delaware	2	7.3	2.68	1.6	Limestone
Detroit River	22	7.3 ± 1.3	2.39 ± .09	8.8 ± 2.6	Limestone
Bois Blanc	14	8.5 ± 0.8	2.46 ± .14	7.7 ± 4.1	Limestone
Bass Island	9	10.0 ± 2.5	2.59 ± .22	5.1 ± 3.5	Limestone
Salina	34	10.8 ± 2.7	2.75 ± .12	2.2 ± 2.2	Limestone, Shale, Dolomite
Guelph-Lockport	21	11.8 ± 0.4	2.72 ± .07	2.7 ± 1.1	Dolomite
Clinton-Cataract	16	5.4 ± 1.6	2.57 ± .10	3.4 ± 1.0	Limestone, Shale
Queenston	32	5.3 ± 1.0	2.57 ± .06	4.0 ± 1.8	Shale
Meaford-Dundas	1				Shale

Table XI-1 PHYSICAL PARAMETERS MEASURED ON THE CORES FROM EACH FORMATION IN THE LONDON BOREHOLE

The formations with the highest average thermal conductivities are the Salina and Guelph-Lockport in which large sections of the Salina and all of the Guelph are composed of dolomites. These sections also have the highest densities and the lowest porosities. The formations with the least variation in conductivity are the Guelph-Lockport and the lower section of the Salina. Conductivity and density variations are also low in the

Queenston formation composed of reddish shales.

A careful examination of FIG XI-3 reveals that in general the density and thermal conductivity follow each other quite well: where conductivity increases so does the density. It is also apparent that densities are usually higher where porosities are lower. However it may be more revealing in terms of these parameters to take a closer look at several selected formations; this is done in Chapter VII-4.

XI-4 THERMAL DIFFUSIVITY OF SEDIMENTS

Several thermal diffusivity and conductivity measurements were made in the London borehole using an 'in-situ' cylindrical heat source, (Beck et al, 1971). Their results are summarised in the following table together with the thermal capacities calculated as the ratio K/κ and the calculated specific heats using the disc densities. The specific heats are calculated according to the relation $c = K/\rho\kappa$.

Depth (m)	$\kappa \times 10^3$ cm /sec	$K \times 10^3$ cal/cmsec$^\circ$C	ρc cal/$^\circ$Ccm	c cal/gm$^\circ$C	Formation Name
65 - 87	12.2	8.2	0.66	.26	Detroit River
192	18.7	11.7	0.63	.23	Bass Island
238	9.1	5.9	0.65	.24	Salina F
284	17.9	12.5	0.70	.26	Salina E
424	22.8	12.9	0.57	.20	Gasport
438	8.4	6.2	0.74	.28	Rochester

Table XI-2 THERMAL CONDUCTIVITY AND DIFFUSIVITY MEASUREMENTS DETERMINED 'IN-SITU' IN THE LONDON BOREHOLE

The rock types sampled cover shales, limestones and dolomites and yet the specific heat varies between only .20 and .28 cal/gm^{o}C, and the thermal capacity between 0.57 and 0.74. Whereas the variation in specific heat is 40% that in the heat capacity is 30%, slightly smaller. In general, as the specific heat of the rocks increases the density decreases, making the thermal capacity of the rocks more constant than either the specific heat or the density. However to make a reasonable estimate of the thermal diffusivity of these sediments it is sufficient to assume a mean value for the specific heat and to use the routinely acquired measurements of conductivity and density to calculate the thermal diffusivity.

Beck et al (1971) have commented that the thermal capacity determined from several line source measurements on cores from the Detroit River formation in the London hole was 0.65 with a variation of only 20%.

Further values of the specific heats and densities of rocks given in the Smithsonian Physical Tables (1956) are reproduced below:-

Rock Type	Temp (^{o}C)	Specific Heat (c) cal/gm^{o}C	Density gm/cm^3	ρc cal/^{o}Ccm^3
Basalt	12 - 100	.20		
Dolomite	20 - 98	.22	2.83	.62
Gneiss	17 - 99	.20	2.64	.53
Granite	12 - 100	.19	2.67	.51
Limestone	15 - 100	.22	2.61	.57
Marble	0 - 100	.21	2.69	.56
Sandstone	——	.22	2.50	.55

Table XI-3 SPECIFIC HEATS AND THERMAL CAPACITIES OF SEVERAL ROCK TYPES

In the preceeding table it is apparent that the thermal capacity varies rather more than the specific heat. However since there a few measurements of the specific heat of rocks available it is true to say that, although it is not valid to always assume a thermal diffusivity of .01 cm^2/sec as is often done (the mean value of the U.W.O. measurements is 0.015), it would seem reasonable to assume a specific heat of about .20 for igneous and metamorphic rocks and about .23 for sediments and then to calculate the thermal diffusivity. Mean diffusivities calculated for each formation in the London hole are as given in the table below:-

Formation	Mean Thermal Conductivity mcal/cmsec°C	Mean Thermal Diffusivity cm^2/sec
Overburden	2.3	.003
Delaware	7.3	.010
Detroit River	7.3	.015
Bois Blanc	8.5	.014
Bass Island	10.0	.017
Salina	10.8	.015
Guelph-Lockport	11.8	.020
Clinton-Cataract	5.4	.010
Queenston	5.3	.009

Table XI-4 MEASURED FORMATION THERMAL CONDUCTIVITIES AND CALCULATED FORMATION THERMAL DIFFUSIVITIES IN THE LONDON BOREHOLE

However it is very apparent that thermal diffusivity measurements ought to be made routinely on various rock types.

These are the two physical parameters which govern the conduction of heat. In the steady-state the thermal conductivity will govern the temperature distribution with depth whereas disturbances, such as the effect of past surface temperature changes, will be governed by the thermal diffusivity. If other forms of heat transfer occur such as water movement these will be governed by yet other coefficients.

XI-5 CALCULATION OF HEAT FLOW

The heat flow is calculated as a combination of the temperature gradient and the thermal conductivity of the rocks. The most straightforward approach and perhaps the commonest is to divide the borehole into lithologic sections and for each of these sections to take the product of the least squares temperature gradient and the arithmetic or harmonic mean thermal conductivities. This is a measure of the heat flow, if the borehole is vertical, in a vertical direction. A weighted mean heat flux can then be calculated for the complete borehole by weighting each section dependent on its length.

An alternative method is to again use the product but to perform this process over intervals of three or more measurement units. In this case the least squares gradient is calculated through three successive temperature points and the mean conducvitivity calculated for all samples within the interval. The product is the interval heat flow. This can be calculated as a 'running mean' process. Smoothing this process over larger intervals reduces the noise due to measurements and sampling errors, leaving the gross variations.

A different approach again is to combine the observations in the form suggested by Bullard (1939):

$$T_z = T_o + q \sum_i D_i/K_i$$

where T_z is the temperature at a depth $z = \sum_i D_i$, K_i is the thermal conductivity of the i^{th} homogeneous section of thickness D_i, and T_o is the surface intercept temperature. The mean heat flow is calculated from the slope of a plot of T_z versus $\sum_i D_i/K_i$.

Most authors find some slight differences associated with the use of different methods of calculation. Some analysis of the U.W.O. borehole has already been published (Judge & Beck, 1967; Beck & Judge, 1970) and copies of these papers constitute Appendix VIII. However certain parts of the analysis were either not mentioned in those papers or only dealt with briefly and therefore are dealt with in the remainder of this chapter.

XI-6 HEAT FLOW IN THE U.W.O. BOREHOLE

As is shown in FIG XI-4 there is a similarity in the variations, with depth, of thermal resistivity and temperature gradient. This does not however lead to a constant heat flow with depth, excepting the decrease and reverse of heat flow at shallow depths and a sharp increase at 446 m. The latter is due to a combination of high conductivity and the end of the casing. Smoothing the data over five units and then twenty-five units as shown in FIG XI-5 brings out the variations of heat-flow with depth very well, and these are discussed in a later section. It is fairly difficult to put error figures on the interval heat flows since the errors on the single disc conductivities can only be estimated in a fairly rough fashion. However the standard deviations from the least squares fits over at least 10 data intervals give some idea of the variation.

Heat flows have also been calculated by the thermal depth method. This form of calculation is not suitable for examining the variation with depth but presents a good method of averaging sections. The errors calculated by this method are usually lower than those derived by the interval method. However as is shown in the table below the mean heat flows are not significantly different.

Section (m)	Heat Flow		Number of Data Points
	Interval μcal/cm²sec	Bullard μcal/cm²sec	
167 - 241	.71 ± .02	.67 ± .008	17
241 - 314	.81 ± .008	.80 ± .004	19
314 - 387	.67 ± .007	.67 ± .004	20
387 - 454	.76 ± .06	.74 ± .01	25
454 - 527	.79 ± .004	.75 ± .005	23
527 - 596	.87 ± .004	.86 ± .004	19
192 - 596	.76 ± .04	.76 ± .002	110

Table XI-5 HEAT FLOWS MEASURED IN THE LONDON BOREHOLE

The errors given on the interval 387 - 454m are larger than for other intervals partly because of the rapid formation changes and the inherent sampling problems and partly due to the effects of the end of the casing.

Heat flow, as was mentioned in the previous section, may be calculated by lithologic section. The heat flows in the upper three units are disturbed by the shallow temperature inversion.

In the table following, the range of heat flows below this inversion is .62 to .97 averaging .76 ±.10. Sectional heat flow is lowest in the Guelph and highest in the alternating shale and limestone sequence of the

Cataract group. This is the section which also contains the end of the casing which is accompanied by a large increase in the temperature gradient.

Formation	Section Heat Flow	Number of Conductivity Samples
Detroit River	.10	27
Bois Blanc	.49	14
Bass Island	.59	9
Salina G & F	.76	7
Salina E	.83	8
Salina C	.75	7
Salina A_2	.79	6
Salina A_1	.66	5
Guelph	.62	10
Lockport	.69	10
Rochester	.69	5
Reynales Cabot Head Manitoulin	.97	11
Queenston	.82	33

Table XI-6 MEAN FORMATION HEAT FLOWS IN THE LONDON BOREHOLE

According to Bullard the thermal effect at the end of the casing would be negligible if the distance from the top of the casing is greater than 50a, where a is the radius. In this case one depth interval has an abnormally high gradient since it encompasses the zone less than 50a. In this short zone the interval heat flow rises to 1.40, almost twice as high as the average heat flow. However, as mentioned previously the values are also disturbed by a short interval of high conductivity and therefore it is not possible to make any detailed estimates of the casing effect. Weighting the heat flow according to length of each section yields a heat flow of .78 below the Bass Island formation.

XI-7 VARIATION OF HEAT FLOW WITH DATA INTERVALS

It is of very great interest to determine the effect of reducing the number of temperature and conductivity data points. Many of the measurements listed in Lee & Uyeda (1965) have only 15-20 conductivity samples associated with them. In this borehole, if heat flow is calculated for data intervals of 8, 16 and 30 m, the table below lists the resulting mean interval and Bullard heat flows below 192 m.

Data Intervals	Interval	Bullard	Number of Data Points
4m	.76 ± .04	.76 ± .002	110
8m	.78 ± .04	.74 ± .002	55
16m	.77 ± .06	.71 ± .005	28
30m	.80 ± .05	.71 ± .009	14

Table XI-7 VARIATION OF HEAT FLOW WITH DATA INTERVAL IN LONDON BOREHOLE

As the number of points of temperature and conductivity information is diminished the error limits on the heat flow values increase and the differences in the calculated values between the Bullard and Interval methods increase. Certainly the differences between the 4 m and 8 m intervals are not very significant and thus perhaps a fewer number of measurements would have been sufficient.

XI-8 CLIMATIC CORRECTIONS TO HEAT FLOW

An examination of the smoothed heat flow curve in FIG XI-5 shows a regularly increasing heat flow in the section of the hole below 457 m. A possible explanation of at least part of this increase could be the onset and retreat of the Pleistocene ice-sheets with other superimposed variations. The greatest portion of this effect would be due to the most recent glacial period in North America, the Wisconsin, as is described in Chapter XII. As a crude assumption, consider the period of ice-cover to be a single step function with a sudden onset 100,000 years ago and a sudden retreat 10,000 years ago. The ground temperature at onset was suddenly reduced to -2^{o}C and was as at present both before and after the removal of the ice. This correction to the borehole temperatures increases the heat flow in all sections of the borehole but does not improve the overall variance by a great deal. Varying the temperature step difference and the times of onset and retreat does not seem to give one solution which is very much better than any other. Failing this approach the model derived in Chapter XII is believed to be the best one and the heat flows have been corrected using both the Wisconsin history and also the Wisconsin together with the post-Wisconsin history. The interval heat flows calculated below 180 m are shown in FIG XI-6 and in the table belww for the two diffusivities which probably bracket the true diffusivity range of the rocks. In the figure, the inclusion of the post-Wisconsin history is seen to overcorrect the heat flow since it then decreases with depth below 200 m. However the Wisconsin correction alone using an assumed diffusivity of .015 gives the best fit and yields an equilibrium heat flow of 1.06 ± 7% compared with an uncorrected one of .76 ± 8%.

Model Used For Correction	Heat Flow Interval
Measured temps	.76 ± .06
Wisconsin + Post $\kappa = .01$	1.13 ± .09
Wisconsin + Post $\kappa = .015$	1.10 ± .09
Wisconsin alone $\kappa = .01$	1.10 ± .08
Wisconsin alone $\kappa = .015$	1.06 ± .07

Table XI-8 HEAT FLOW IN THE LONDON BOREHOLE CORRECTED FOR PLEISTOCENE GALCIATION

The real uncertainty in the correction of a measured heat flow to obtain equilibrium values is of course very much higher. Many of the problems associated with these corrections are discussed in Chapter XII and until these are resolved the estimation of a glacial correction is something of an academic exercise. The large number of papers published recently - eg., Jessop (1968), Crain (1969), Horai (1969) and Beck (1970) - point out the great importance of these corrections; but only measurements in 3000 m boreholes and a greater understanding of ice dynamics will finally resolve the problem.

XI-9 HEAT FLOW VARIATION WITH DEPTH

Examination of any of the heat flow plots (FIG XI-5) reveals variations of heat flow with depth. Ignoring the Pleistocene correction and the region above 190 m, which is dealt with in Chapter VIII, a minimum heat flow of .62 occurs between 365 and 395 m and a maximum of 1.03 occurs at the bottom of the hole. Again the heat flow spike at the end of the casing is ignored. There appear to be a variety of explanations for this phenomenon. First, possible sample bias must be discussed. It is not unusual in the selection of samples, particularly of shales, for the most competent samples to be chosen since many of the less competent ones will not yield sufficient length for the preparation of discs. Thus since the selected samples usually are more calcareous or more highly compacted they yield higher thermal conductivities than would more representative samples. If the mean heat flows in the shale sections are higher than those in the carbonate sections it is suspected that this could have occurred.

Indeed if a table were prepared of carbonate versus shale heat flows the situation would look bad since the low heat flow zone occurs in a carbonate sequence. However an adjoining carbonate and shale sequence, ie., the Lockport dolomite and the Rochester shales, give the same formation heat flow. Thus sampling is probably not a significant error.

Failing this as an explanation, four natural physical phenomena remain to explain the results. First, underground waterflows. These are bringing cooler or warmer water into various regions of the borehole depending on the value taken as the normal heat flow. Secondly, endo- or exothermic chenical reactions are taking place in certain formations. Thirdly, certain sections of the borehole may contain larger amounts of heat-producing radioactive materials than others. Fourthly and finally, the borehole may pass through or near geological structures which distort the heat flow lines.

The waterflow explanation was dealt with briefly by Judge & Beck (1967) who took the mean heat flow value as the normal one. The Salina A_1 and the upper Guelph units are more porous and permeable than the formations directly above and below them, and the minimum heat flow correlates well with the maximum in the porosity curve. These formations outcrop between 100 and 160 km north and east of London at elevations several hundred metres higher than that of London, which would produce a 400 to 500 m head of water beneath London. Seepage of cooler rainwater and snow melt waters into the formation at outcrop would seem to be a reasonable explanation. Velocities of flow can be estimated fairly simply. Consider a cube of rock of 1 cm side, which contains a mass of water $\rho_w\sigma$ where ρ_w is the density of the water and σ is the porosity of the rock. If the thermal gradient in the horizontal plane is $\Delta T/\Delta x$, then the rate of maintenance of a 1 cm^3 sink is $c_w \rho_w \sigma v_x \Delta T/\Delta x$ where c_w is the specific heat of water and v_x is the horizontal velocity. If the difference in heat flow between the disturbed zone and the normal heat flow is ΔQ over a section of thickness H then;

$$\Delta Q = Hc_w \rho_w v_x \sigma \Delta T/\Delta x$$

Since the temperature in the Guelph section is about $10^{\circ}C$ and the ground temperature in the outcrop area is $8^{\circ}C$, a temperature difference, ΔT of $2^{\circ}C$ seems reasonable. ΔQ is 0.14 $\mu cal/cm^2 sec$ with respect to the mean heat flow over a thickness H of 30 m, porosity σ is 5% and the distance Δx is 100 km. Using these figures a velocity v_x of 10 cm per day is obtained. Bullard and Niblett (1951) have pointed out earlier that underground water movements of a few inches per day with a few degrees C temperature difference are all that is necessary to affect the heat flow significantly. Permeabilities in the Guelph might be expected to vary considerably depending on the degree of reefing present, but assuming a typical permeability of 1 md and unit cross section, a hydrostatic head of 30 m will produce a flow of 6 cm/day. The regional dip is very small, of the order of 2°, thus the picture is one of cooler water flowing horizontally along the upper section of the Guelph absorbing heat as it does so. The length of time for the water to flow from the outcrop to the London area would be about 60 years which suggests that using water-dating methods, it might be possible to verify the theory.

Now this simple seepage of meteoric water into a sedimentary basin presents serious difficulties since the recharging basin must be discharging and the in-situ waters are generally highly saline and therefore of higher density than the seepage water. A mechanism discussed by Beck & Judge (1970) which might have significance in this regard is that of seepage waters slowly leaching out salt and moving down hydraulic gradient as they gain in density. This mechanism may well apply to the London area. As is shown in FIG XI-7 London lies in the marginal zone of the Salina salt; the salt being fairly continuous in the downbasin direction to the west but having been largely leached away to the east. However several isolated salt masses remain, one of them a few kilometres to the east of London, near the village of Crumlin, and up hydraulic gradient. Thus meteoric water seeping into the region of this salt body may increase in density and then move down-formation until stopped by an impermeable formation such as the Rochester and then down hydraulic gradient displacing less saline waters. Obviously to verify this mechanism as the cause of the low heat flow zone would require a great deal of careful measurement in the local area. It is apparent that this process must have been going on for a long time and thus the effect is not concentrated in the high porosity zone but has been disperesed by conduction processes. FIG XI-8 shows the interval heat flow, smoothed over several data intervals, plotted against the porosity of conductivity samples.

Little is known about many geological processes such as dolomitisation of limestone. Such processes are probably exothermic in nature and it would be intriguing to suppose that at the boundary of penetration of water a chemical change may be occurring producing heat. However at the lower boundary, the high heat flow is in the Clinton-Cataract which is primarily composed of shales so this explanation is not really very convincing.

It is possible that the pattern of heat flow is caused by a series of exothermic and endothermic chemical reactions but this is thought to be highly unlikely.

Possibly, as was mentioned before, there are significant differences in heat-producing radioactive materials in sedimentary rocks but these should have formed an equilibrium pattern in the borehole which would bear little resemblance to the observed heat flow distribution. A causal relationship will almost certainly exist between the heat flow distribution and the radioactivity of the rocks since the shales corresponding to zones of higher heat flow probably contain higher potassium and thorium than the dolomites corresponding to the low heat flow zone. The differences in content however are unlikely to be sufficient to cause the depth distribution of heat flow.

The last possibility is of an unknown structure close to the borehole distorting the heat flux. While on a regional scale the geology of the area is well known and relatively simple, it is also known that various beds and sub-units pinch out, particularly in the Salina and the Clinton-Cataract. A north-south section including the U.W.O. hole is shown in FIG XI-9 which illustrates this pinching out effect. Unfortunately the U.W.O. borehole is the only one of any significant depth in the immediate vicinity of London so only hypothetical structures might be considered. Obviously more detail is required on the geology in a 1 kilometre radius about the U.W.O. borehole. The effects of some simple structures on the temperature gradient are discussed in Chapter VI-6.

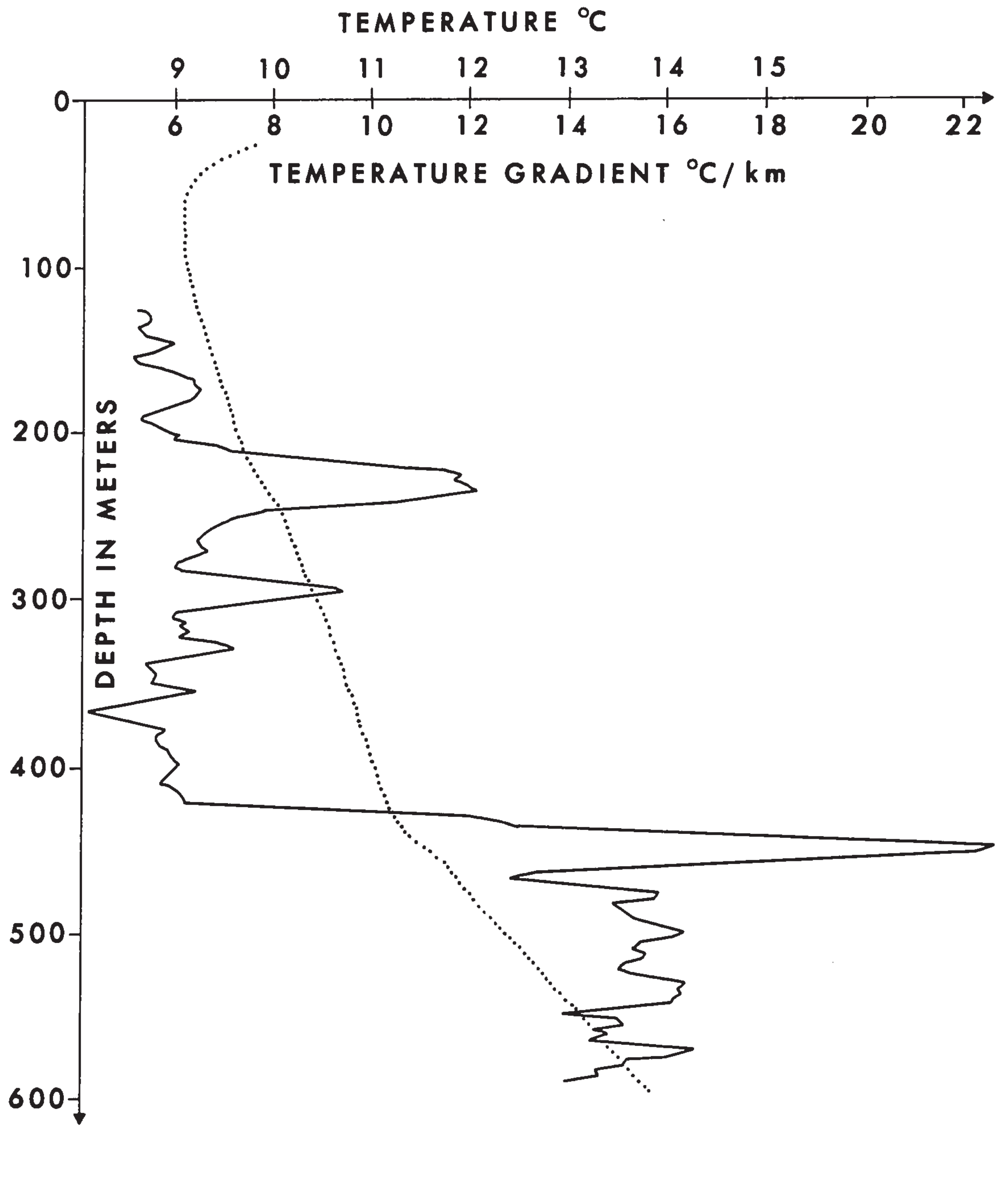


FIG. XI-1
TEMPERATURES AND TEMPERATURE GRADIENTS
IN U.W.O. BOREHOLE

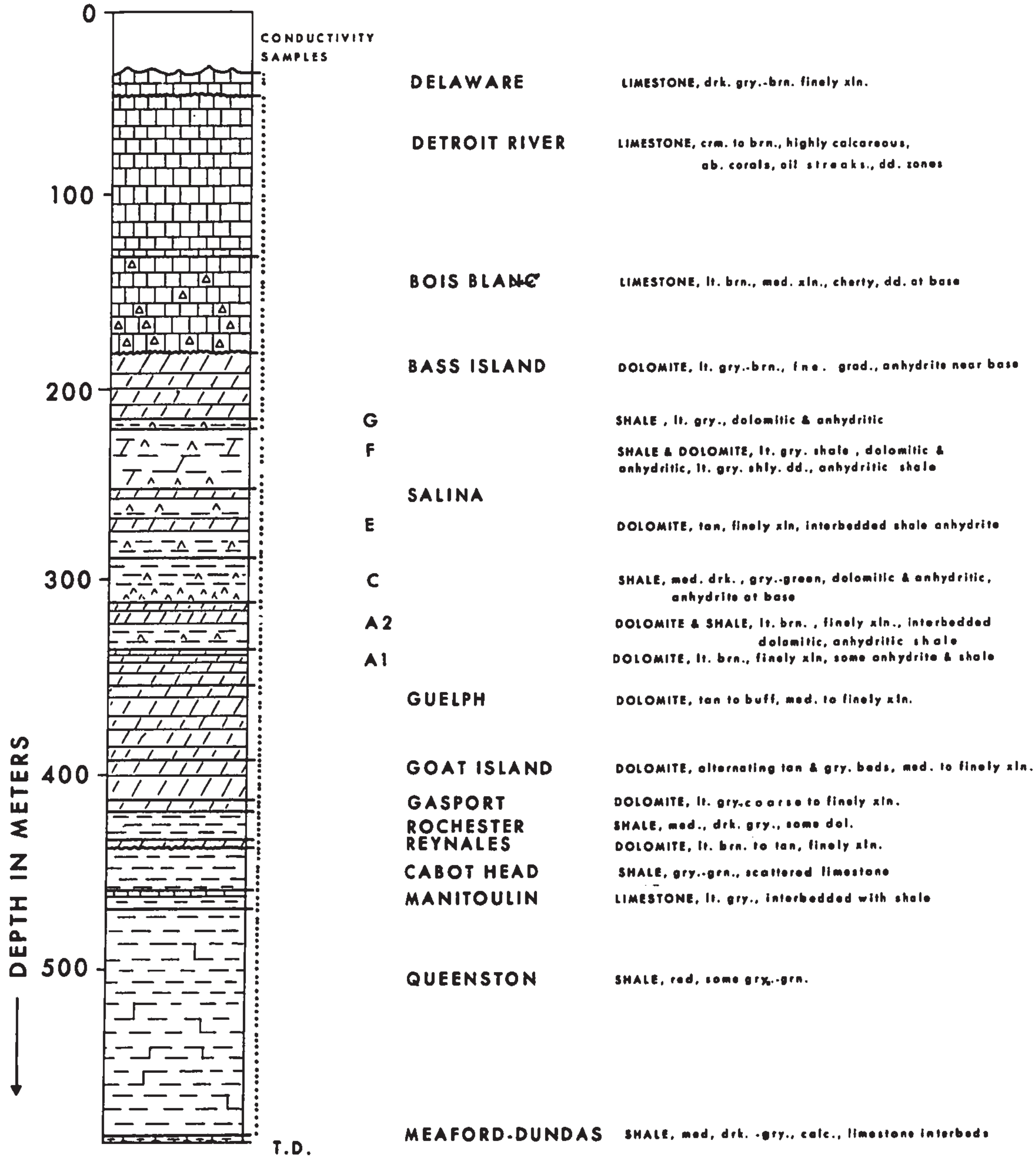


FIG. XI-2 LITHOLOGY OF THE LONDON BOREHOLE

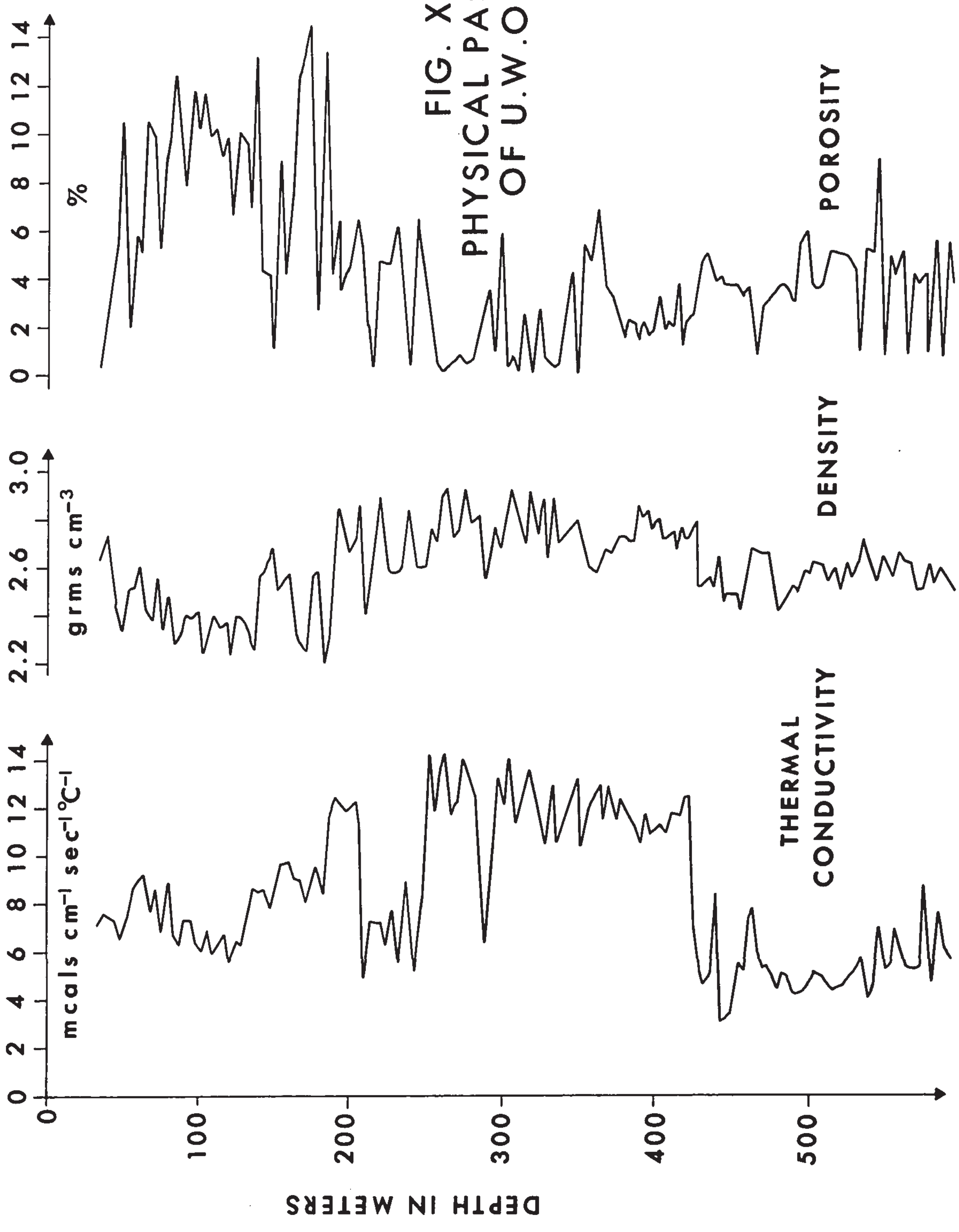


FIG. XI-3
PHYSICAL PARAMETERS
OF U.W.O. CORE

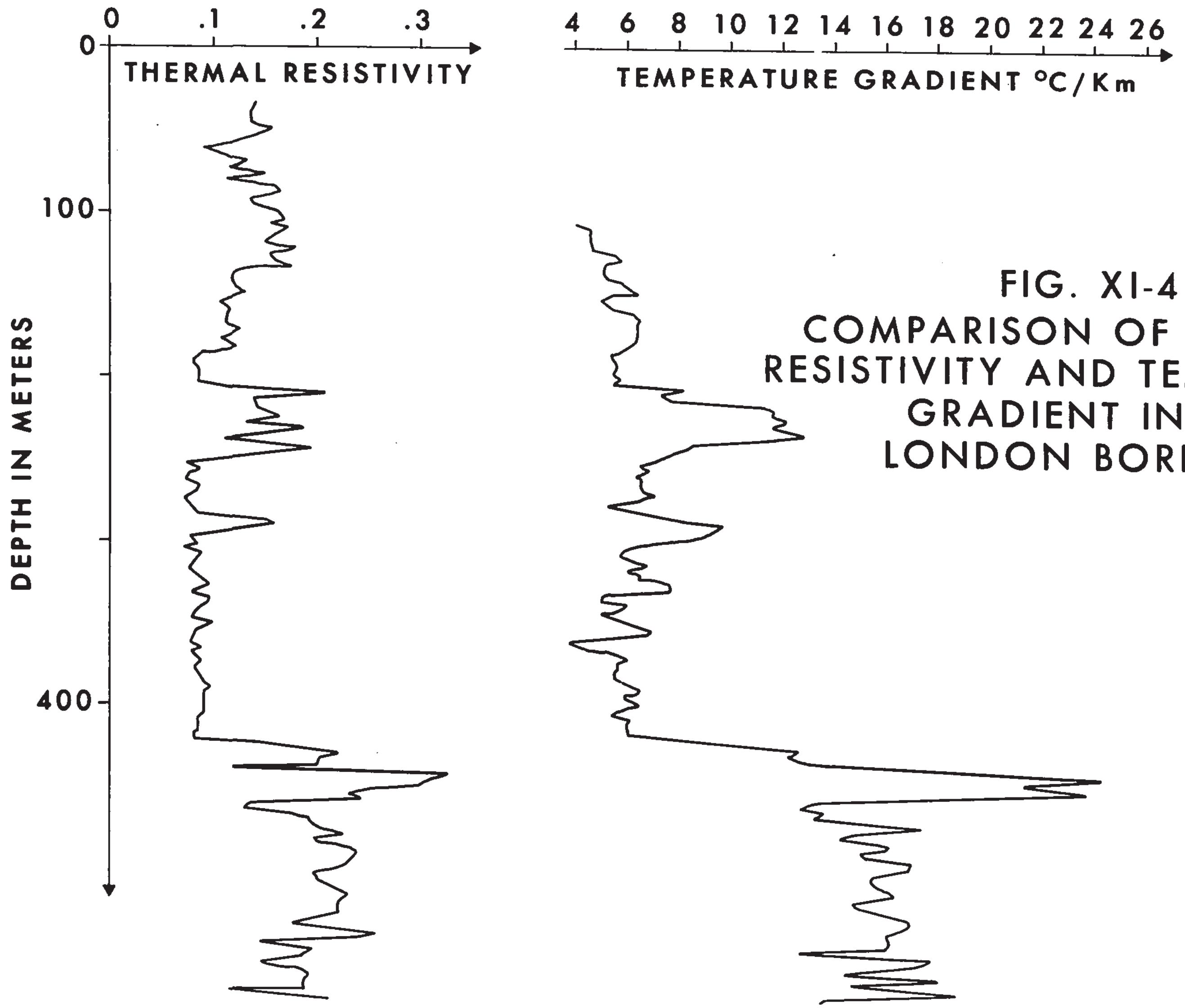


FIG. XI-4
COMPARISON OF THERMAL RESISTIVITY AND TEMPERATURE GRADIENT IN THE LONDON BOREHOLE

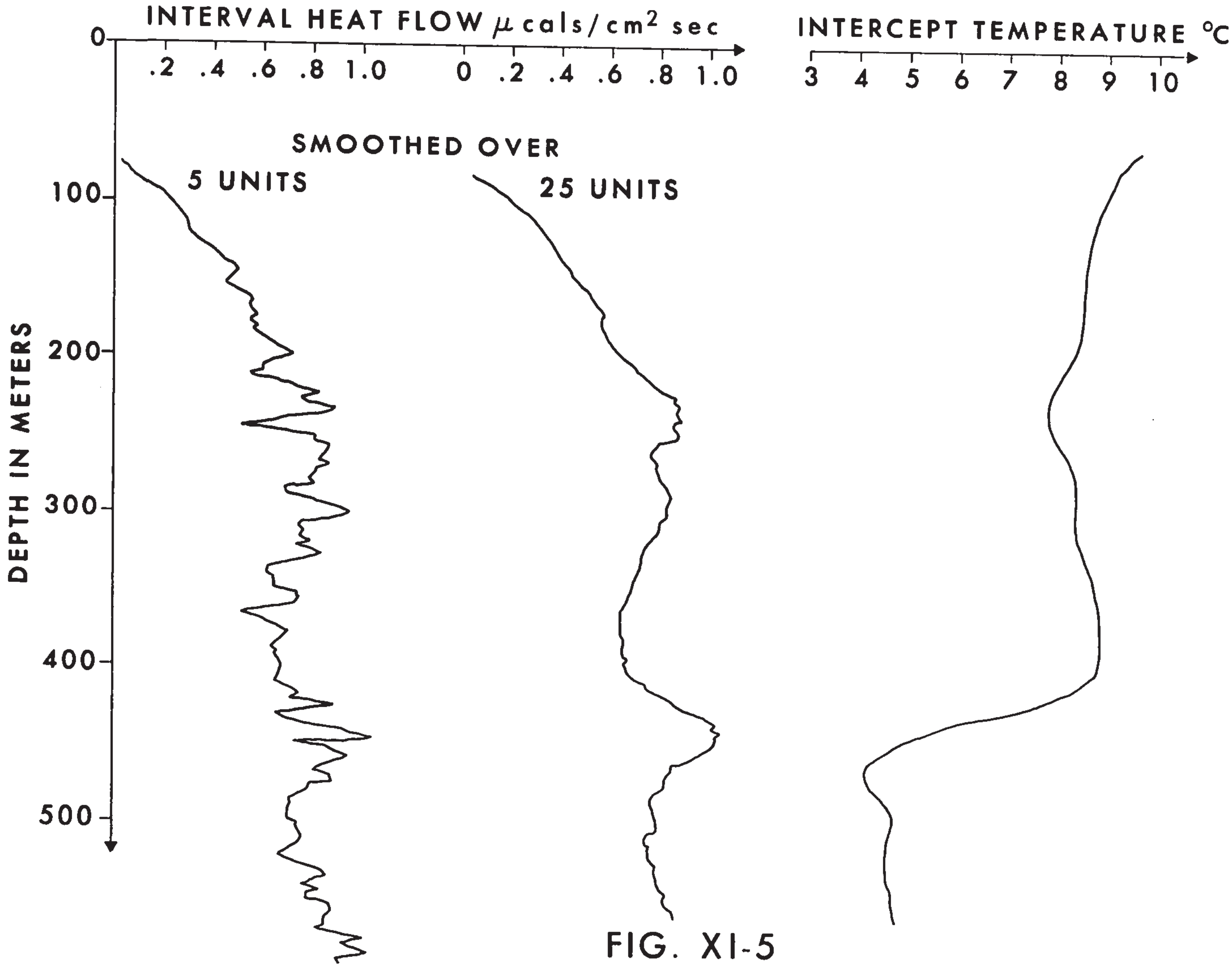


FIG. XI-5
VARIATION OF HEAT FLOW WITH DEPTH

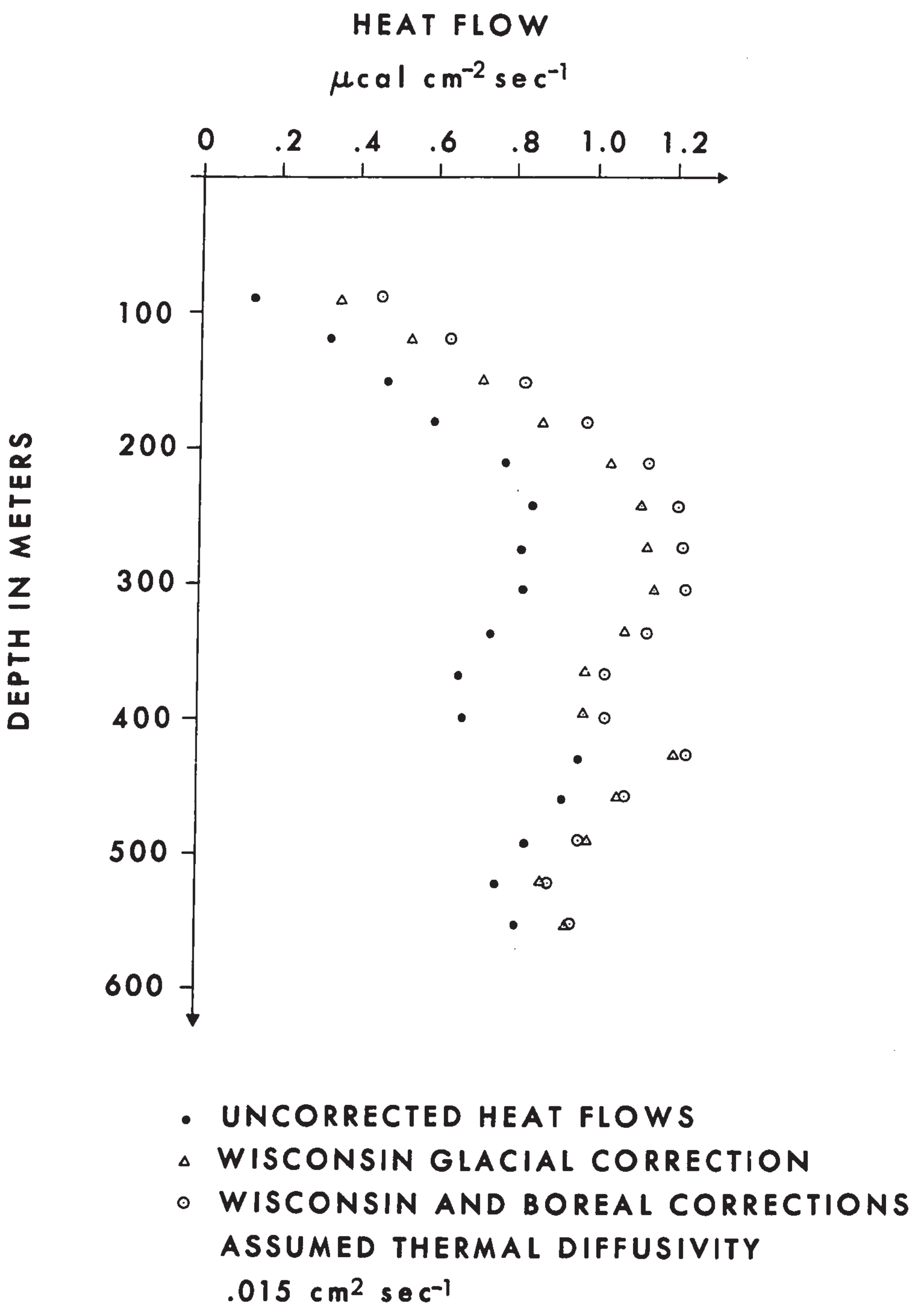


FIG. XI-6
INTERVAL HEAT FLOWS CORRECTED FOR QUATERNARY CLIMATE CHANGES

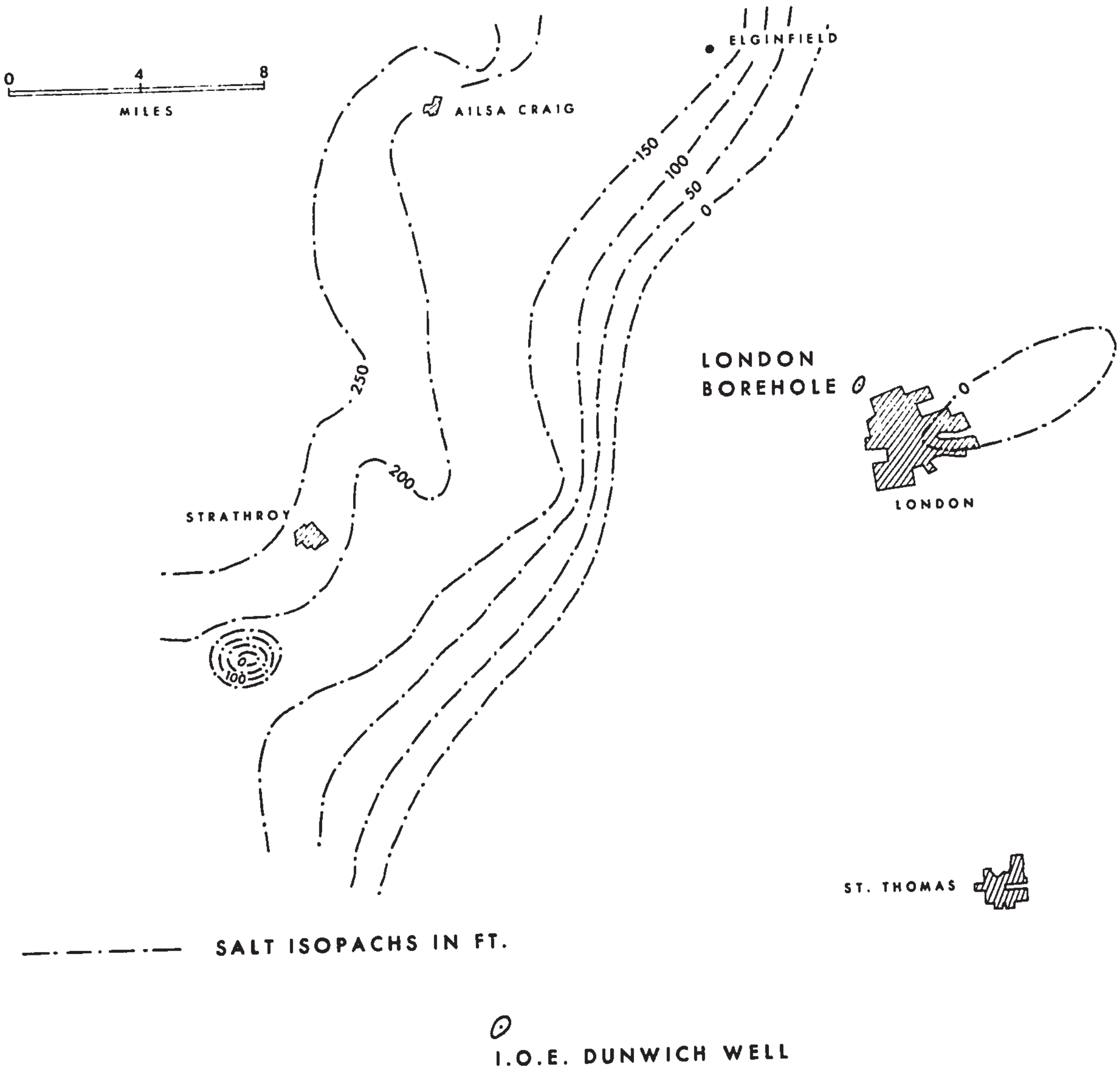


FIG. XI-7 POSITION OF THE LONDON BOREHOLE AND UPPER SALT ISOPACHS

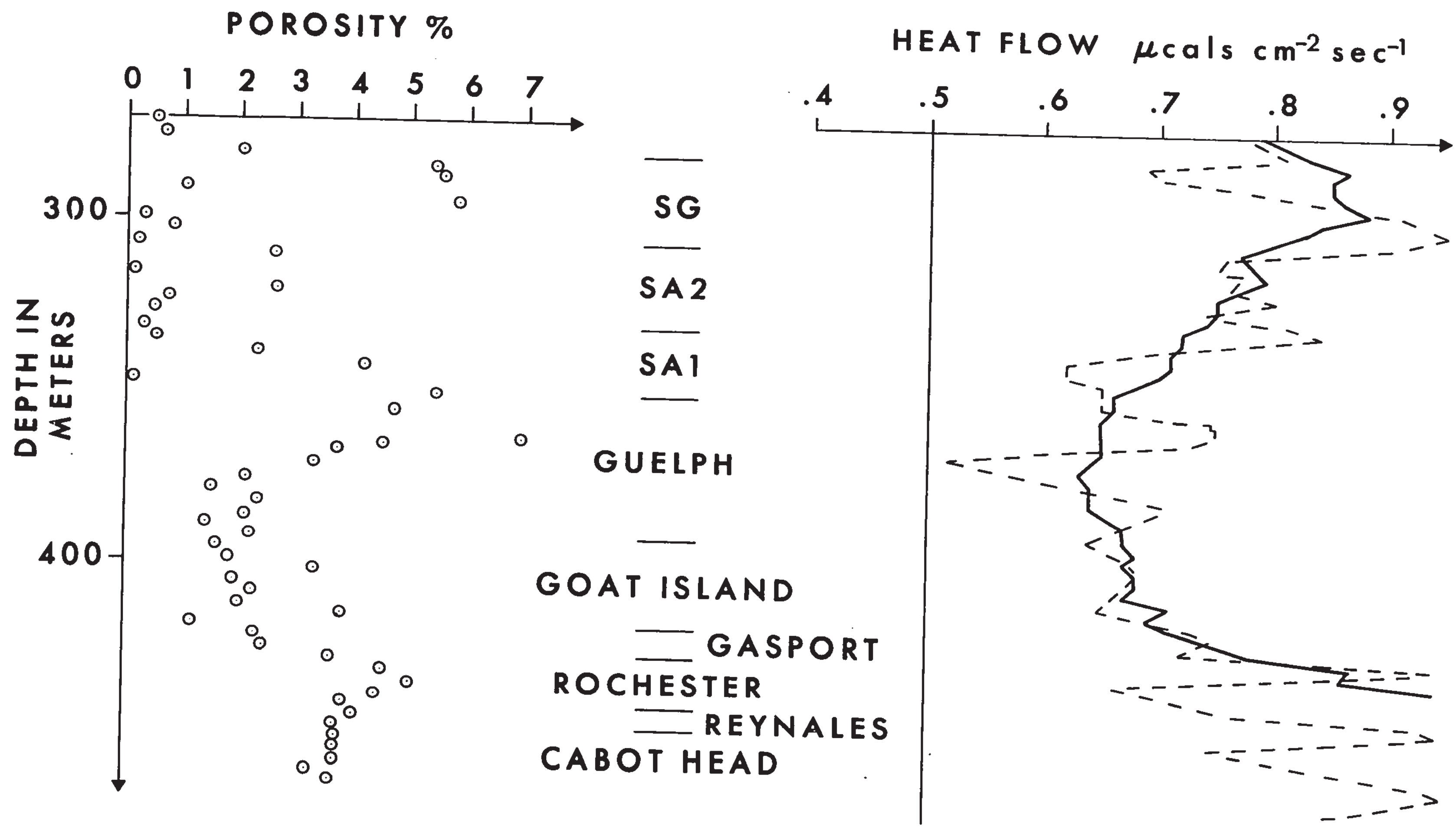


FIG. XI-8 POROSITY AND HEAT FLOW IN THE LONDON HOLE

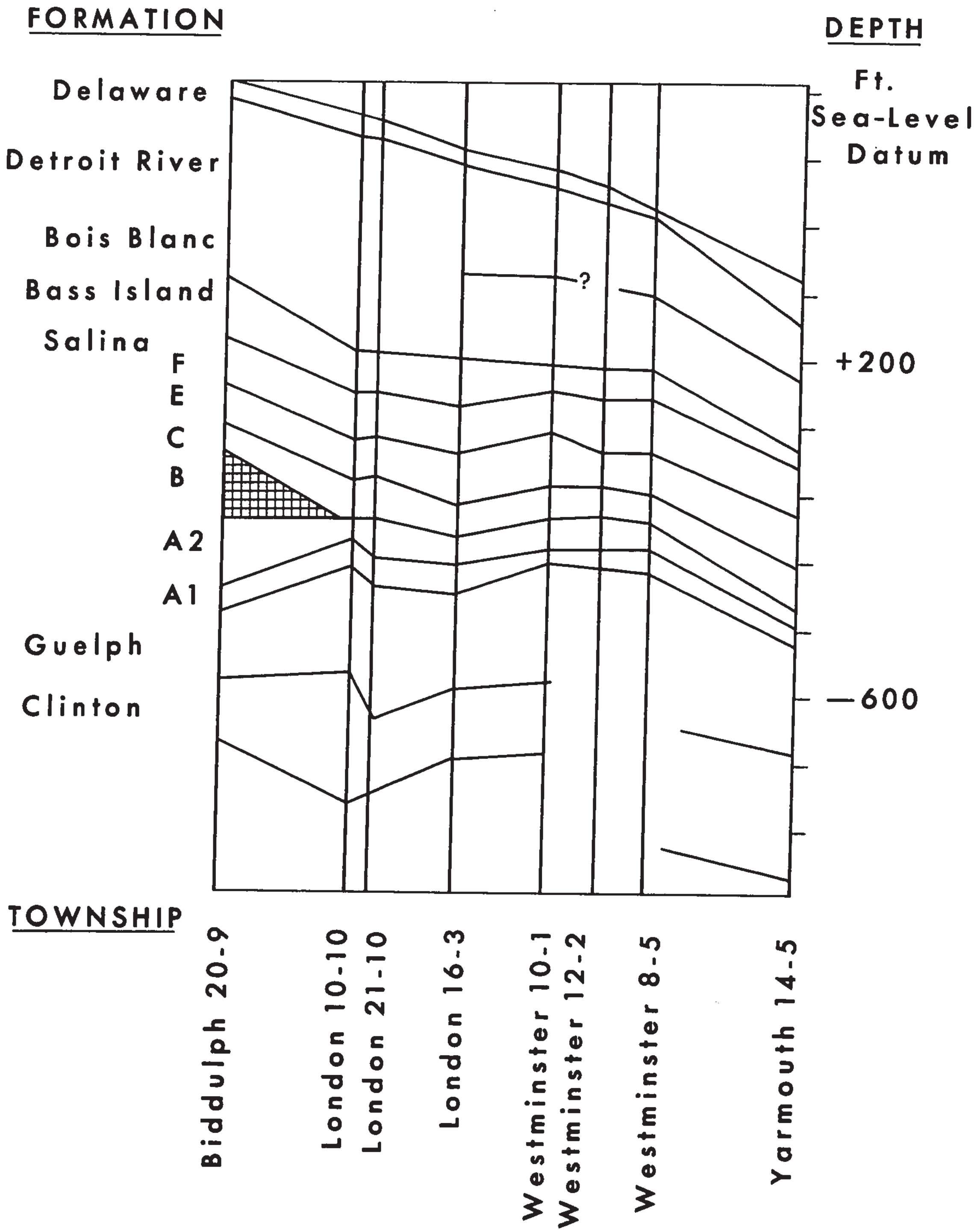


FIG. XI-9
STRUCTURAL RELATIONSHIPS
IN THE LONDON AREA

CHAPTER XII

SURFACE TEMPERATURE CHANGES

AND HEAT FLOW

PREFACE

The equilibrium heat flow is dependent on there being constant boundary conditions for long periods of time, even by geological standards. It is possible that the temperatures at the inner boundary which is at some depth in the mantle have remained essentially constant for long periods of time. However, the outer boundary, the air to surface boundary, is highly variable in incident radiation and hence in the surface temperature. This of course is also the region in which measurements of the terrestrial heat flux can be made. The problem becomes one of a steady state equilibrium heat flow with a series of transient disturbances superimposed on it. In order to find the former, which is the value reflecting thermal conditions in the crust and mantle, it is necessary to determine the latter either by assuming reasonable models or by an analysis of the heat flow variations in boreholes.

This chapter is concerned with methods of determining the influence of past climatic changes on the shallow thermal field of the earth. Surface temperature changes from the very recent to those which might have occurred 2 million years ago are considered.

Since the phenomenon is that of conducted heat it is useful to consider a simple temperature variation over varying periods of time. The table following reproduces the depth at which various temperature cycles are reduced to 0.2% of the total change at the surface for a diffusivity of .01. This will give some idea of the depths which need to be considered to resolve past surface temperature changes. As may be noted, the effects of the Pleistocene changes have had time to penetrate to considerable depths.

Period	Depth of Penetration
1 yr	20 m
100 yr	200 m
10,000 yr	$2x10^3$ m
1,000,000 yr	$20x10^3$ m
10,000,000 yr	$63x10^3$ m
4,500,000,000 yr	$1330x10^3$ m

Table XII-1 PENETRATION OF TEMPERATURE CYCLES OF VARYING PERIODS

XII-1 SURFACE TEMPERATURES

The subsurface temperature field is dependent not only on the heat production and heat flux from greater depths but also on the thermal history of the surface. A change in the surface temperature will be propagated downwards until a new equilibrium temperature field is established. One of the first problems which arises is that of determining the surface temperature.

Conventional meteorological stations record data at about 1.2 m above ground level in an area well away from buildings or trees and preferably over ground covered by short grass (Instructions to Observers in the Meteorological Service of Canada, 1930). The thermometer is in the shade and provided with some form of forced ventilation. Comparative studies of air and shallow subsurface temperatures over long periods of time are rare, and thus the surface temperature is fairly difficult to estimate from the conventional meteorological results. Temperature gradients as high as $5^{\circ}C/km$ have been observed within ±1 cm of the atmosphere-surface interface. Budyko (in Odishaw, 1964) has observed ground temperatures warmer than air temperature by up to $5^{\circ}C$ in summer-time in desert and Arctic coastal regions. The energy balance at the earth's surface involves, as primary components, the short wave radiation from the sun, long wave radiation from the earth and sky, transfer of heat through the ground, transfer of heat through the air and the contribution of the latent heat of evaporation or condensation. During the day-time there is a gain of radiant energy by the surface, the surplus being used to heat the soil, to heat the air and in evaporational cooling. At night-time there is a net loss through radiation by the surface. To balance this effect, heat flows upwards through the ground, downward from the air and by condensational heating (dew formation). There is an overall storage of heat in the ground during the summer-time and a loss during the winter-time. It is to be expected, then, that surface temperatures will depend on such factors as soil cover, winter snow cover, amount of vegetation, depth of weathering, position of the water-table and topography. A few examples are: the surface temperature may vary by 2°C between north- and south-facing slopes (Budyko, 1964); during measurements made by the author on microclimates at Lake Hazen, temperatures in the Dryas on ice hummocks were observed up to 10°C greater than those measured on open ground; and mean temperature differences of about 4°C between a bog surface and a granitic surface can develop during the summer due to evaporation effects. Measurements of soil temperatures by Penrod, Elliott and Brown (1960) for a period of five years near Lexington, Kansas, showed a mean annual ground temperature of 0.6°C below the air shelter measurements. Lovering et al. (1963) suggest from a series of measurements in shallow boreholes in Tintic, Utah, that ground temperatures are at least 1.7°C below the mean annual air temperature. In more southerly, fairly dry areas, the surface temperature appears to be lower than the air temperature. However, these areas may be controlled by summer evaporation since they have no snow cover in the winter. In snow covered areas the mean soil surface temperatures may be higher than the mean annual air temperature because the soil is partially insulated from temperature extremes in the winter-time. For example, in Yellowknife the ground temperature is almost 5°C above the mean annual air temperature, whereas in Ottawa and Guelph the difference is 3°C. These differences correlate reasonably well with the annual number of days of snow cover, as is shown in the data below:-

Station	Air Temp. °C	Ground Temp. °C	Difference °C	Number of Days With Snow Cover (Approximate)
Fort Simpson	-3.9	0.9	4.8	200
Fort Vermillion	-2.1	4.3	6.4	160
Ottawa	5.0	8.2	3.2	120
Toronto	8.8	10.7	2.0	80
Guelph	5.6	8.6	3.0	120
Harrow	8.9	10.4	2.5	<80

Table XII-2 COMPARISON OF MEAN ANNUAL AIR AND GROUND TEMPERATURES FOR VARIOUS PARTS OF CANADA

However, as the comparison of the records for Forts Simpson and Vermillion shows it depends also on snow thickness during this period of cover.

Station	Temp. Difference °C	Snow Cover Oct-Dec cm.	Snow Cover April-May cm
Fort Vermillion	6.4	15.0	28.4
Fort Simpson	4.8	8.1	7.4

Table XII-3 EFFECT OF SNOW COVER ON DIFFERENCES BETWEEN MEAN ANNUAL AIR AND GROUND TEMPERATURES

This short comment does little more than illustrate a problem in correcting heat flow measurements and indeed in using disturbances in temperature gradient with depth for palaeoclimatic analysis. A great deal more work is required on this problem.

XII-2 SHALLOW TEMPERATURE INVERSIONS - OBSERVATIONS

Many boreholes in Canada and the United States show thermal inversions in their temperature-versus-depth profiles with minima ranging between 30 and 122 m. These phenomena have often been noticed in the past and have generally been attributed to ground-water movements (Birch, 1954). In many of the older measurements a zero or near zero gradient in the first 152 m below the surface was observed (Van Orstrand, 1935). However, these measurements were made with maximum thermometers and near-surface inversions would have been logged as a constant temperature with zero gradient. With the advent of more accurate methods of logging, these shallow inversions have been recorded on numerous occasions (Diment, 1964, Saull, 1961, Mustonen, 1967, Judge and Beck, 1967) and have been attributed to ground-water movement or to the clearing and settlement of the land by pioneer farmers. In Puerto Rico, Diment (1964) attributed the inversion to the burning off of the bush prior to the original planting of sugar cane. These explanations, while certainly being mechanisms by which the surface temperature would be changed, seem to localize a phenomenon that occurs commonly throughout much of the world, and thus the inversions might be explained as natural surface temperature changes in the recent past. One of the great importances of any correlation is whether heat flow determinations may be made in shallow boreholes of a total depth of 150 m or less. Lovering and Goode (1963) have concluded such determinations are possible in holes of less than 30m and they found no inversions or curvatures in temperature logs of a 300 m hole in the area.

In the table below, the depths of the minima in the inversions are shown for holes in southern Ontario and Michigan, a few holes from other parts of Canada. Generally, the depths range from 61 to 120 m with drift cover varying from non-existent to 183 m. From a study of the data of the Great Lakes region, an apparent relationship of deeper reversals in areas of thicker drift cover suggests a mass transfer process plays a role, since the thermal diffusivity of soils is generally lower than that of bedrock material. This could be due to ground-water or water vapour movement. Several of the Overisal group of holes in the Michigan Basin show offsets from each other in their temperature curves at depth, while at shallow depths the curves converge to similar temperatures. These particular instances must be due in part to ground-water movement, since the temperature at which the curves coincide is the temperature of the ground-water used industrially in Holland several miles away and where the reservoirs are at a shallower depth. However other differences occur in areas where several boreholes occur close to one another. For example, although four holes in the Sudbury area show very similar inversion depths, and the undisturbed gradients are nearly equal in three of them, the intensity of the minimum varies considerably. This suggests that, although the time of increase of temperature did not vary much, the value of the increase did.

Site		Depth of Temperature Minimum	Extrapolated Gradient	Apparent Surface Temperature Increase	Bedrock Cover
Name	Location	m	$^oC/km$	^oC	m
CANADA					
London	S. Ontario	70	5.6	1.1	34
Ottawa	E. Ontario	70	13.0	1.1	30
Russel	E. Ontario	55	37.2	0.55	nil
Sudbury	N. Ontario	69	13.1	0.3	nil
Kapuskasing	N. Ontario	91	10.3	0.9	nil
Lake Dufault	W. Quebec	84-107	*10	*0.9	nil
Wallace	Nova Scotia	61	14.5	very small	-
Providence	N.W.T.	curvature no min.	75	2.2	34
Winter Harbour	N.W.T.	curvature no min.	48	3.0	18
U.S.A.					
Marion	Michigan	114	21.3	1.3	18
Northville	Michigan	76	36.1	0.8	128
Wayne Co.	Michigan	56	8.2	1.1	21

Table XII-4 DEPTH OF MINIMA IN BOREHOLES SHOWING SHALLOW TEMPERATURE INVERSIONS

XII-3 RECENT CHANGES OF AIR AND GROUND TEMPERATURE - OBSERVATIONS

There are a few meteorological readings of temperature and precipitation dating back to the early period of settlement in Canada; however, in this section only the instrumental records of weather stations will be discussed. The first station in Ontario to make daily observations was the Magnetic Observatory in Toronto, established in 1840. These records constitute one of the longest series of observations in North America. In 1871 the Meteorological Service of Canada was organized and in 1873 began to publish detailed daily weather observations.

Before discussing the data that has been accumulated over the years, it may be useful to set some limits on the value of the data. Many apparent climatic changes may be attributable to a relocation of the instruments or to a change in instrumental or observing techniques. Changes of up to 0.75°C may result from a very slight site relocation or from removal of bush near the site, whilst even larger changes may result when new buildings are located nearby. A further real effect is due to urbanization in the past century. This amounts to almost 2.0°C, as a result of the increase in artificially created heat and pollutants in the atmosphere. Papers on these phenomena have been written by Mitchell (1961) and by Summers (1967).

Studies of the changes in the world's climate have shown an increase of at least 0.5°C in the air temperature between the late nineteenth century and 1940 (Callender, 1961). Studies of oxygen isotope ratios in the Greenland ice suggest a climatic optimum in the 1920's and 1930's with a relatively cool period prior to this which began in the 1820's (Daansgaard & Johnson, 1969). Throughout much of the Northern Hemisphere records show a cold spell in the 1880's, followed over much of Canada by a rapid warming in the 1890's. Since this thesis is concerned for the most part with Southern Ontario, the variations for this region are considered first. The 10-year running means show that, after a rapid increase in air temperature just before the turn of the century, only minor variations occurred for the next 30 years, followed by higher temperatures in the decade ending in the late 1930's. Mean values then decreased in the 1940's, but take on an increasing trend to the end of 1955 when then reached a maximum. In comparison, air temperatures in northeastern Ontario show a steady decrease after the first decade of the century and rise again only at the end of the 1940's. Since that time the trend has been towards increased temperatures, but the level is still no higher than it was fifty years ago. Northwestern Ontario shows a decrease over the last decade, in common with western Canada. In Ontario the average increase in temperature over the recorded period is approximately 1.5^{o}C. The table below summarizes the changes in Ontario on a regional basis.

	Temperature Changes in Various Parts of Ontario Change in ^{o}C for Period							
Region	1880-1890	1890-1900	1900-1910	1910-1920	1920-1930	1930-1940	1940-1950	1950-1960
Southeast	no records	+0.8	-0.28	+0.25	0	0	+0.28	+0.28
South Central	partial records	+1.0	0	0	-0.28	+0.28	0	+0.42
Southwest	partial	+1.0	0	0	0	+0.42	0	+0.28
Northeast	no records	partial	0	-0.56	-0.28	0	+0.42	+0.28
Northwest	"	+1.25	+0.8	-0.42	+0.42	0	-0.28	0

Table XII-5 RECORDED CHANGES IN MEAN ANNUAL AIR TEMPERATURE FOR VARIOUS AREAS IN ONTARIO extending this summary to other parts of Canada, air temperature changes in the Prairies derived from Thomas (1964) are :-

Region	1880-1890	1890-1900	1900-1910	1910-1920	1920-1930	1930-1940	1940-1950	1950-1960	1960-
Alberta	no records	partial	+0.8	+0.25	0	-0.5	+0.2	-0.8	+
Saskatchewan	"	partial	+0.7	+0.6	+0.3	-0.3	-0.4	-0.5	+
Manitoba	"	none	+0.9	+0.1	+0.3	0	+0.3	-0.6	+

Table XII-6 RECORDED CHANGES IN MEAN ANNUAL AIR TEMPERATURES ACROSS CANADA

At 35 of 43 stations in British Columbia, Powell (1965) has observed upward trends in air temperature varying from 0.1°C at Kamloops to 1.2°C at Chayoquot.

In the east, in Nova Scotia and Prince Edward Island, temperatures have shown a general increase since the 1890's with a slight decrease during the 1940's and this is repeated in the Middle St. Lawrence area (Thomas, 1955, Longley, 1953).

In general the air temperatures in eastern Canada have increased over the last 80 years by several degrees, whereas this increase has been very small over much of the west. Boreholes in eastern Canada generally have shallow inversions in their temperatures, whereas holes in British Columbia and the Yukon, while often possessing slight curvatures in temperature, do not show inversions. Thus it would seem that the evidence for an inter-relationship between the occurence of the minima and the increase in air temperature is very strong.

If this were a simple correspondence, it might be expected that the curves in a given region would have similar characteristics. However from inversions recorded in the Sudbury Basin of Northern Ontario and in the Michigan Basin, it is apparent that some variation does occur in the characteristics even in a limited region. Thus a detailed study of surface environment together with temperature measurements over a long period of time is necessary to solve these discrepancies. The nature of the surface, whether a water-saturated bog, open grassland, forest, outcrop etc., obviously plays an important role in the evaporation history and in the ground penetration of winter minimum temperatures and thus in the freezing and thawing history and associated latent heat as do microclimatic and local topographic effects. However the apparent relationship between long term changes in air temperature and shallow temperature inversions in boreholes, at too great a depth to be ascribed to annual variation, warrants a closer examination as is done in the next section.

XII-4 PENETRATION OF THERMAL CYCLES

Periodic changes of temperature at the surface of the earth produce similar temperature changes within the surface but with an amplitude decreasing exponentially with depth. Let the surface be $z = 0$, then for a surface variation of the form;

$$V(o,t) = V_o + \sum_{n=1}^{\infty} V_n \cos (n\omega t) \quad , \quad . \quad . \quad . \quad . \quad . \quad .. \quad i)$$

the variation of temperature with depth will be given, from Carslaw and Jaeger (1959) by

$$V(z,t) = V_o + \sum_{n=1}^{\infty} V_n \exp (-h\sqrt{n}z) \cos (n\omega t - h\sqrt{n}z) \quad . \quad . \quad . \quad .. \quad ii)$$

where $h = \sqrt{\frac{\omega}{2\kappa}}$, ω is $2\pi/P$, where P is the period of the surface disturbance and κ is the thermal diffusivity of the homogeneous medium. The wavelength of the disturbance λ will be given by

$$\lambda = 2(\pi\kappa P)^{\frac{1}{2}} = 2\pi/h \quad \ldots\ldots\ldots\ldots \quad \text{iii)}$$

and the phase lag, which may be expressed as a velocity of propagation of the disturbance, is v where

$$v = 2(\pi\kappa/P)^{\frac{1}{2}} \quad \ldots\ldots\ldots\ldots \quad \text{iv)}$$

For a surface diffusivity of .01 cm^2/sec, the diurnal variation penetrates to a depth of a metre, unseasonal temperature spells to a few metres and even the annual variation only penetrates to some 20 metres for an amplitude decrease of $\exp(-2\pi)\Delta V$ or$(.0019\Delta V)$where ΔV is the annual range of surface temperature. The annual range at the surface at London, Ontario, is about 27°C; thus, for a diffusivity of .01, the annual range would be reduced to about .05°C at 20 metres depth.

Since the amplitude attenuation is so large, higher harmonics rapidly drop out, as do other terms, in the expansion as a series in n, hence this form of equation can be used to investigate longer term fluctuations. If the disturbance is written in the form $V_o \cos(\omega t)$, then the temperature at depth z is given by

$$T(z) \;=\; T_o + gz + V_o\, e^{-kz}\cos(\omega t - hz). \quad \ldots\ldots\ldots\ldots \quad \text{v)}$$

Differentiating this equation twice leads to expressions which can be solved to determine P, t and V_o as has been presented by Beck & Judge (1970).

In a similar fashion a further way to consider surface temperature changes is as a step function -

$$T(z) \;=\; T_o + gz + V_o \,\mathrm{erf}\, \frac{z}{2\sqrt{\kappa t}} \quad \ldots\ldots\ldots\ldots \quad \text{vi)}$$

where T_o is the undisturbed surface temperature, g is the undisturbed temperature gradient in homogeneous rocks of diffusivity κ, and V_o is a step amplitude of temperature change at a time t in the past. Differentiating the equation twice again enables solution for t and V_o and this method has also been used by Beck & Judge (1970).

A slightly different way of deducing some of the factors is to take advantage of the position of the minimum in the curve, ie., the point at which $\frac{\partial T}{\partial z} = 0$.

$$g = -\frac{V_o}{\sqrt{\pi\kappa t}} \quad e^{-z_M^2/4\kappa t} \quad , \quad \ldots\ldots\ldots\ldots \quad vii)$$

Rearranging gives an expression for the depth of the minimum z_M, since for a minimum to occur V_o must be negative.

$$z_M^2 = 4\kappa t(\tfrac{1}{2}\ln \pi\kappa t - \ln g - \ln V_o) \ . \quad \ldots\ldots\ldots\ldots viii)$$

Since g is determined, it may be used to relate κt and V_o. For z_M to be real,

$$\ln V_o > \tfrac{1}{2} \ln(g^2\pi\kappa t),$$

which becomes:

$$|V_o| > g\sqrt{\pi\kappa t} \ . \quad \ldots\ldots\ldots\ldots \quad ix)$$

The immediate point which comes from this solution is that only in situations where g is low or V_o is relatively large will an actual inversion be found. If these conditions are not fulfilled, although an inversion will not perhaps be found, a curvature of the temperature: depth curve is to be expected. FIG XII-1 shows a plot of some of the curves to be expected for different temperature gradients and sizes of disturbances. The table in section XII-2 shows approximate thermal gradients taken from below the disturbed region. To the extent that the disturbance may penetrate two or more different formations, these gradients may not be representative. Examination of the depth of minimum against the thermal gradient does not indicate any simple relationship connecting them.

Rather than analysing as a simple step or sine function, it is also possible to use a series of steps of small magnitude. However this type of solution is not unique and so it is necessary to begin with the possible changes of air-temperature and attempt to fit the magnitudes to the recorded temperature-depth variations.

Further mathematical functions which can be used fairly simply for curve fitting are the ramp and a general polynomial approach which are dealt with in Carslaw and Jaeger (1959, p.63). However to obtain a simple picture these methods are not used in this script.

The assumption of a homogeneous half space of constant diffusivity and conductivity is also not a very valid one in reality. While it is not hard to calculate the equations for a two layer model such as usually exists if overburden is present, they are algebraically more difficult to handle.

XII-5 INVERSIONS IN SOUTHERN ONTARIO - ANALYSIS

The only hole in southern Ontario accessible continuously for scientific tests was the one drilled by the Department of Energy, Mines and Resources on the University of Western Ontario campus. Detailed temperature logs were made at various intervals of time and the data used for a detailed study of heat flow variations in a borehole. The temperature versus depth curve showed a shallow inversion, thus presenting an ideal opportunity to test the various hypotheses to explain such effects.

Initially, as suggested in Judge & Beck (1967), it can be examined as a drilling effect. The circulating drilling fluid is warmed as the hole becomes deeper and, since there are very porous zones in the upper portion of the hole, some of the warmer fluid may penetrate and warm the upper layers. Furthermore, when drilling stops either temporarily or permanently, convection brings warmer fluid to the top of the borehole. Thus the upper portions of the hole take more time, and the lower portions less time, to return to equilibrium than is predicted by the theory based on the simple conduction model of Jaeger (1955). If this explanation is valid, then the negative gradient section which is acting as a remanent source should show evidence of decay back to equilibrium, although the rate of decay will be dependent on the source size. Examination of the temperature records since completion of drilling shows that the decay does not show signs of being any greater or any more rapid in the section of the hole above the inversion. Thus it is concluded that this mechanism is not the cause. Roy (1963) examined the return to equilibrium of a borehole at White Pine, Michigan in which, although the appearance of the inversion changed, it did not disappear.

A second possible explanation is that of ground-water movement. Such water movements at these depths should be surface-temperature dependent, ie., cooler in spring during the melt season, for which there is no evidence. A thermistor probe was left suspended in the well at a depth of 60 m and temperatures recorded over a period of several months. While there were fluctuations of temperature of almost 0.1°C the variations are not large enough to support the inversion. It is not the object here to state that there is no ground-water movement, but merely that it is not the principal cause of the inversion. Many of the air-conditioning plants in large buildings and factories in the London region withdraw water from a depth of approximately 46 m and pump it back after use as a coolant. This water is of the H_2S type. The very presence of sulphide waters suggests that there is no rapid ground-water movement on a large scale at this depth. Industrial use of the water should result in positive anomalies and discontinuities in the temperature:depth curve, since the use is recent in origin. Thus it seems valid to investigate the inversion as a change in surface temperature.

Much of the area around London was heavily forested at the beginning of the nineteenth century and fairly rapidly occupied and cleared soon afterwards. FIG XII-1 shows a step function fit which would also correlate with land clearance and settlement being the cause of the inversion. The fit is not very good unless a very low diffusivity, in direct contrast with 'in-situ' measurements, is assumed since the land was cleared more than 150 years ago.

As discussed in the previous sections, the air temperature has increased over much of Canada in the last 100 years. A step function to the borehole data fit suggests that the time of change of the surface temperature was less than 50 years ago.

The step and sinusoidal changes in surface temperature have been tested on the inversion zone in the U.W.O. borehole. The table which follows, reproduced from Beck and Judge (1970) shows how well the two models could be fitted. It is very apparent that the best fit is obtained for a periodic function with a maximum amplitude of 1.1°C. The change in air temperature at London and other meteorological stations in the same climatic area is well-documented and amounts to a maximum change in air temperature of 0.5°C. How changes in air temperature are translated into surface temperature changes is uncertain being dependent, as discussed earlier, on such factors as cloud cover, annual precipitation, snow cover in winter etc. Possible explanations for the discrepancy would be increased snow cover for a longer period of time during the winter, perhaps accampanied by increased cloud cover and thus a reduction of evaporation cooling in the summer. The present meteorological stations measuring soil temperature have been in existence for only 10 years and it is therefore difficult to determine any definite trends. Comparing the results from the periodic function analysis in the borehole with the changes in air temperatures suggest a very good fit in the form of the curve, in the period and the interval, but in terms of the amplitude the air temperature change is only half of the required surface temperature change. Since, as mentioned above, certain meteorological factors may seriously affect the amplitude while having no effect on the period, the fit to the curve is very encouraging and suggests further such analyses should be made.

Table XII-7

RECENT CLIMATIC CHANGE AS EVIDENCED IN LONDON BOREHOLE

Depth (meters)	Step Function		Periodic Function
	t (years)	V_o (^oC)	V_o (^oC)
39.1	7	-3.24	-1.72
42.1	8	-3.08	-1.56
45.0	9	-2.76	-1.43
47.9	10	-2.59	-1.33
50.8	11	-2.34	-1.25
53.8	12	-2.04	-1.20
56.7	13	-1.76	-1.15
60.1	17	-1.48	-1.13
63.5	21	-1.29	-1.11
67.2	24	-1.17	-1.09
70.8	30	-1.07	-1.09
74.3	35	-0.97	-1.10
77.7	36	-0.96	-1.09
81.0	39	-0.93	-1.11
84.3	34	-1.02	-1.10
88.4	29	-1.23	-1.10
92.0	27	-1.37	-1.06
95.0	22	-1.76	-1.09
97.9	20	-2.00	-1.11
100.9	21	-1.91	-1.23
103.9	22	-1.63	-1.41
106.9	23	-1.40	-1.71
Means	20	-1.6	-1.1
95% Conf. limits	9	0.3	0.1

p = 107 years

t = 75 years

λ = 241 meters

Most of the other holes in southern Ontario did not have near surface water columns and so presented no opportunity to repeat the work. However examination of inversions from several holes in Eastern Ontario has revealed similar relationships in which the supposed change in surface temperature is consistantly larger than recorded changes in air temperature.

This analysis is not meant to rule out the possible effects of ground-water on shallow temperature measurements, because there is evidence in the three holes in Holland, Michigan that the inversions are modified by, if not controlled by ground-water movement. If does, however, suggest that many shallow inversions may be caused by slight increases in surface temperatures over the past 100 years.

XII-6 HISTORICAL TEMPERATURE CHANGES

Periodic cycles similar to one determined in the previous section will have little amplitude effect below 200 m, however a relatively small air temperature increase may cause a larger increase in the soil temperature. Thus geothermal results may be highly dependent on climatic changes in the past. The earliest systematic meteorological records were begun in Great Britain in 1680 and have been summarized by Manley (1961). Prior to 1900 the majority of the increases or decreases in temperature were not greater than 1°C on this record and the cold or warm spells generally did not last for longer than 10-20 years. Considering a step pulse of 1°C to have begun 100years ago and lasted for 10 years, the maximum effect in the temperature curve occurs at a depth of less than 91 m, but its maximum disturbance to the temperature is only .02°C. This effect would not be noticed in the normal noise level of the temperature analysis. Short-term effects of such a magnitude and period as those suggested by Manley's record would not be detectable in borehole temperature logs; they would merely contribute to the general noise level. Thus, only major, long-lasting changes in surface temperature need be considered in corrections for the period between 1700 and 1900.

A great deal of semi-quantitative data exists in historical records for surface temperature changes in the period prior to physical measurements. The evidence has been accumulated by such authors as Lamb (1965), Stefansson (1943), and Bergthorson(1962) from historical observations of sea-ice, glacier behaviour, upper limits of forest cover, highest levels of farming, positions of vineyards and many more factors. It suggests a climatic warm period existed between 950 A.D. to 1250 A.D. in the Northern Hemisphere with temperatures between 2 and 4°C warmer than the average over the past 100 years. This increase in temperature appears to have been larger in regions marginal to the Arctic Ocean, such as Greenland. In the sixteenth century the so-called 'Little Ice Age' occurred with temperatures probably reaching their minima around 1550. This period probably saw the lowest temperatures, about 2°C lower than at present, and the greatest extensions of the ice-sheets since the end of the last ice-age.

Air temperature changes with a period of 300 years about 1000 years ago would have given rise to maximum temperature effects of about .1°C at a depth of 183m. It is possible that this order of effect might be observed in a uniform medium, ie., a borehole through granitic rocks, but it certainly would not be detectable in the sedimentary sequences of southern Ontario. Recently a very good verification of some of the early historical data has come from an examination of the oxygen isotopes in the ice cores from Camp Century (Dansgaard and Johnson, 1969).

Major climatic changes beforethis time move into the realms of legend in the records of man. However, many techniques are available at present of making physical measurements on the 'remains' from those periods.

XII-7 PREHISTORICAL QUATERNARY CLIMATES

Throughout recorded history various hardships or sudden benefits have been attached to fairly rapid changes of climate in a region, eg., the worsening of the climate in Greenland in the fifteenth century and its resultant effect on the Norse communities, or the ability to grow grapes in Britain in the eleventh century. Recent changes and their association with temperature inversions at shallow depths in boreholes have been correlated with meteorological records. Before the advent of man the recorder, climatic changes were occurring. Since, if these disturbances were large, they may affect the equilibrium heat flow and it is important to deduce the variations. Several methods of determining this information exist, such as the O^{18} content of foraminifera and the coiling characteristics of globigerina to determine ocean palaeotemperatures (Ericson, 1959). Mollusc species in the Champlain Sea have helped determine immediate post-glacial conditions in the St. Lawrence region (Elson, 1960). Plant communities change in composition and migrations occur in the direction of more favourable environments. Linked with this is palynology studying the distribution of pollens and spores in sediments (Terasmae, 1961, Bryson, 1967). Many other methods exist which lack of space precludes discussing here.

Using these various methods and sources of data, a model of air temperature variations has been determined. An important feature is a very sudden increase in temperature at the time of deglaciation, since there is evidence of a boreal forest soon after the retreat of the ice sheet (Bryson, 1967). Temperatures in Southern Ontario were probably similar to those in northern Ontario today. In the last 10,000 years the proportion of hardwoods in the forests has changed. The hardwoods were most extensive 4,000 and 7,000 years ago when the air temperature was probably 3.0°C above that at present. After this a deterioration of climate occurred and the fluctuations since that time have been only 1.0 or 1.5°C (Terasmae, 1961).

Dreimanis (1964) has constructed a diagram of the advances and retreats of the Wisconsin ice-sheet in southern Ontario. The ice sheet did not cover the region for the whole Wisconsin period but advanced and re treated several times in the period 10,000 to 100,000 years ago, reflecting climatic oscillations superimposed on a general lowering of the mean annual air temperature.

Terasmae (1961) has presented evidence for a boreal climate in the Plum Point and Port Talbot interstadials so the air temperatures were similar to those of northern Ontario today. Using this data and Dreimanis' (1964), a model for the Wisconsin period in the Great Lakes region has been constructed.

The southern Ontario region was probably ice-free between the times of 75,000 to 68,000 in the Scarborough period, 47,000 to 64,000 the Port Talbot interstadial, 28,000 to 42,000 years ago the Plum Point interstadial. With the deposition of the Port Stanley Drift 14,000 years ago, the region was finally freed of ice. At its maximum extent the ice covered all of Ontario and extended 150 miles into Ohio. By about 16,000 years ago most of the Michigan peninsula was free of ice (Model I). The final retreat, however was fairly complex in southern Ontario and has been described in detail by Hough (1963), Chapman and Putnam (1966) and Prest (1969). The features concerning us here are to reconstruct the period of ice-cover, of water-cover and of exposed land. Around the southern and northern edges of the peninsula the ice was replaced by lakes, of which the old shorelines are still visible. However, much of the higher ground to the east and north of London formed the 'Ontario Island' and was not part of the lakes system. London itself was dry land except for the Lake Maunee period, when it was under a shallow sea. It is convenient then to assume a temperature model in which the peninsula was exposed after being freed of ice cover (Model II). A model III assumes water-cover until 10,000 years ago with a lake bottom temperature of $4^{o}C$ applicable to a lake about 60 m deep, with extensive winter ice-cover. The models are shown in FIG XII-2.

Prior to the first onset of the Wisconsin ice sheets in North America, several other advances and retreats of ice sheets occurred. In southern Ontario the evidence for these is sparse, since their deposits were removed by the advancing Wisconsin glaciers. The evidence in other parts of the world suggests that pre-Wisconsin ice sheets were not continuous but had long interglacial periods between their onsets. A classic account of the possible ice-age models is given by Birch (1948), based on the accounts of Zeuner, Antevs and De Geer. The total length of the Pleistocene is still not accurately known; Emiliani (1961), using oxygen isotope studies on oceanic cores suggested 600,000 years. Using evidence from magnetic stratigraphy and radiochemistry Ericson (1968) has modified these results. However the results given for the Wisconsin periods are inconsistent with those derived from glacial deposits and by Daansgaard & Johnson (1969). Thus a combined model is adopted below:

Glacial Stage	Interglacial Stage	Time of Onset in Years		
		Modified Ericson	Ericson	Birch
Wisconsin II		variable	20,000	10,000
	Port Talbot	45,000	100,000	45,000
I		60,000	120,000	60,000
	Sangamon	100,000	160,000	100,000
Illinoian		375,000	375,000	225,000
	Yarmouth	550,000	550,000	325,000
Kansan		910,000	910,000	525,000
	Aftonian	1,400,000	1,400,000	625,000
Nebraskan		1,720,000	1,720,000	925,000
		2,000,000	2,000,000	1,025,000

Table XII-8 PLEISTOCENE ICE ADVANCES AND RETREATS

XII-8 TEMPERATURES AT THE BASE OF ICE SHEETS

For hundreds of thousands of years during the Quaternary era when the area was covered by the ice sheets, the ground temperatures over a large part of Canada were depressed to those at the base of the ice sheet. Much speculation has occurred in the past as to what these ice-base temperatures might have been, eg., Birch (1948), Crain (1967, 1968), Jessop (1968). The usual assumption is that the ice-base temperature is close to the pressure melting point. Recent drilling through the Greenland and Antarctic ice-caps have given basal temperatures of -13 and -1.6°C beneath 1372 and 2164 meters respectively (Hanson et al, 1966, Gow et al, 1968). Wexler (1961) concluded his paper on the growth of an ice-sheet with the statement that an ice-sheet of thickness greater than 3.8 km might contain several hundred metres close to the pressure melting point at the base of the sheet. Lliboutry (1966) has examined the occurrence of low velocity zones at the base of the Antarctic ice-sheet and shown that these may be due to zones of temperate ice occuring in zones of high horizontal velocity. Ahlman (1948) has shown that the melting point temperatures for confined water in a glacier will be given by the product of the pressure of the ice at a depth z given in kg/cm^2 multiplied by .0073, the temperature change of melting point with pressure. Thus melting point temperatures range from -3.3°C beneath 4.9 km of ice to -1.0°C beneath 1.5 km of ice.

The base temperature of an ice-sheet will be a function of the past surface temperature ablation history, present rate of movement and the geothermal flux, as well as the thickness. Weertman (1966) has suggested several relations for the theoretical profile of an ice-cap which agree reasonably with known profiles of the Barnes and Antarctic ice-caps. Knowing the maximum extent of the Wisconsin ice-sheet and using Weertman's relations, the maximum ice thickness would lie between 3.4 and 4.9 km. A rough check on the minimum thickness may be made by considering the ice-sheet to have reached isostatic equilibrium and then to have recovered putting raised beaches 360 m above sea-level around Hudson Bay. Walcott (1970) has argued that the uplift has some 30% to go to reach equilibrium. Thus the minimum ice load was 1.5 km. Since isostatic equilibrium was not achieved the ice-thickness was greater. Maximum thickness of present ice-caps are 3.4 km for Greenland and 4.3 km for the Antarctic. Using the theoretical profiles, the maximum ice-thickness over southern Ontario, 1300 km from the centre of the ice-cap was probably 1.5 to 2.2 km. Present lapse rates on both the Antarctic and Greenland ice-caps are about $1^{o}C/100$ m of altitude. Assuming a thickness of 3.4 km for the ice-sheet and a depression of the land of 400 m, the air temperature difference between the margins and centre of the cap would have been a maximum of $30^{o}C$, and that between southern Ohio and southern Ontario $15^{o}C$. Present marginal air temperatures around the Antarctic ice-sheet are $-17^{o}C$. The surface temperatures on the Laurentide ice-cap were probably not as low as those of the Antarctica since it was centred at a lower latitude with shorter winters and a higher solar radiation flux.

Trees within 30 km of the ice-front with tall stands of pine only 160 km from the ice-front provide evidence for higher marginal temperatures. Both Bryson and Lamb suggest that the growth of the ice-sheet was very rapid, ie., in 5, to 10,000 years, leaving little time for extensive perma-frost to form beneath the ice-sheet excepting where it already existed and exists today. Robin (1955) has shown that for a 'cold ice-sheet' of thickness greater than 1.5 km in which there is no melting at the base, the temperature differences between surface and base of the ice-sheets vary between 12 and 23°C for accumulation rates of 16 to 64 cm/yr. For an accumulation rate equivalent to that in Greenland this is 17°C. On the basis of this model then the surface temperatures need to be less than -19°C to prevent basal melting temperatures from occurring. This model assumes no horizontal ice movement which generates heat through associated frictional and strain energies and would raise the basal temperature. Since Lamb (1970) in his reconstruction of the Wisconsin climate of North America deduces minimum ice-cap temperatures of -15°C, it is probable that for long periods of time ice-base temperatures over much of the ice-sheet were close to the pressure melting-point or above. This may only apply to the southern portions of the sheet. No attempt has been made to look at the ice-sheet spreading across the Arctic Islands.

XII-9 THEORY OF GLACIAL CORRECTION

The basic theory of ground temperature response to the rapid onset and retreat of an ice-sheet, or indeed to any rapid changes of ground temperature, has been set out by Birch (1948). The temperature T at a depth z is given by:-

$$T = T_o - gz - V_1R_1 - V_2R_2 - V_3R_3 - \ldots\ldots - V_nR_n \quad \ldots\ldots \quad x)$$

where T_o is the present surface temperature, g the geothermal gradient, and

$$R_n = \mathrm{erf}(z/2\sqrt{\kappa t_n}) - \mathrm{erf}(z/2\sqrt{\kappa t_{n-1}}), \quad \ldots\ldots\ldots \quad xi)$$

κ is thermal diffusivity in $cm^2\ sec^{-1}$ and $V_1, V_2 \ldots V_n$ are the respective surface temperature changes for periods from t_1 to t_2, t_2 to t_3, etc.

The quantities that are known from borehole temperature measurements are T, the sum effects of all the temperature disturbances, and z, the depth of this measurement. The surface temperature T_0 may be determined by extrapolating the gradient in the upper portion of the borehole but below the inversion zone. Values of V and t for shallow holes are for some reasonable model of past surface temperature changes. To correctly determine the validity of a complete Quaternary model is very difficult since very rarely does a borehole penetrate to the 3,000 m or so required to measure the undisturbed gradient. However, there are points at shallower depths at which the gradient observed is free of some changes. This may by shown as follows:

Differentiating ii) for a 2-step function gives

$$\frac{\partial T}{\partial z} = g - \frac{V_0}{\sqrt{\pi\kappa t_1}} e^{-z^2/4\kappa t_1} + \frac{V_0}{\sqrt{\pi\kappa t_2}} e^{-z^2/4\kappa t_2} \quad \ldots\ldots \text{xii)}$$

If $\frac{\partial T}{\partial z} = g$ then,

$$\frac{V_0}{\sqrt{\pi\kappa t_1}} e^{-z^2/4\kappa t_1} = \frac{V_0}{\sqrt{\pi\kappa t_2}} e^{-z^2/4\kappa t_2}$$

or

$$z^2 = 2\kappa\left(\frac{t_1 t_2}{t_2 - t_1}\right) \ln\frac{t_2}{t_1}$$

Thus if the Wisconsin Ice Age persisted from 10,000 to 100,000 years ago and the rock diffusivity is .01 cm^2/sec, then z = 1220 m.

Many areas do have boreholes this deep, so this relationship may be useful. For changes in surface temperature which have occurred in more recent times, this does provide a method of isolating various effects, one from the other, eg., for the change in surface temperature which occurred 4,000 to 6,000 years ago, the depth at which it would have little or no effect on the temperature gradient is 550 m. This depth depends only on the times of the ends of the pulse and the diffusivity of the medium.

Variation of glacial corrections is important on a regional and continental basis. For this the curvature with depth is not of specific interest so the equation is reduced to a simpler form for near surface values, ie., $z \to 0$, then

$$\frac{\partial T}{\partial z} = g - V_0 \left\{ \frac{e^{-z^2/4\kappa t_1}}{\sqrt{\pi \kappa t_1}} - \frac{e^{-z^2/4\kappa t_2}}{\sqrt{\pi \kappa t_2}} \right\}$$

Rewriting as a series solution with $z^2/4\kappa t_1$ as x_1 and $z^2/4\kappa t_2$ as x_2

$$\frac{\partial T}{\partial z} = g - V_0 \ \frac{1}{\sqrt{\pi \kappa t_1}} \left\{ (1 - x_1 + \frac{x_1^2}{2}) - \frac{1}{\sqrt{\pi \kappa t_2}} (1 - x_2 + \frac{x_2^2}{2} - \cdots\cdots) \right\}$$

For $z \leq 150$ m and $t_1 \gg 10{,}000$ years, this reduces to

$$\frac{\partial T}{\partial z} = g - V_0 \left\{ \frac{1}{\sqrt{\pi \kappa t_1}} - \frac{1}{\sqrt{\pi \kappa t_2}} \right\} \quad \ldots\ldots\ldots\ldots \text{xiii)}$$

or in terms of heat flow

$$Q = Q_0 - KV_0 \left\{ \frac{1}{\sqrt{\pi \kappa t_1}} - \frac{1}{\sqrt{\pi \kappa t_2}} \right\} \quad \ldots\ldots\ldots\ldots \text{xiv)}$$

Crain (1968) has pointed out that this approximation is only reliable to depths of 150 m.

XII-10 CORRECTIONS FOR LONG-TERM CLIMATIC CHANGES

The assumption that climatic changes follow a step-function requires some discussion. If in fact the last glacial retreat were rather more complex with several re-advances and a period of tundra-like conditions, a closer representation might be a ramp-like change. The effect on the temperature gradient between a step-like retreat 14,000 years B.P. and a ramp-like retreat between 11,000 and 17,000 years B.P. is only about 1% of the correction. However Bryson (1967) has suggested that climatic changes follow some kind of step-function with changes occuring in a rather abrupt manner.

Using equations x) and xi) and the models shown in FIG XII-2, the borehole temperature effects of climatic changes have been calculated. The colder surface temperatures which persisted from 500 to 1500 years ago have their major effects at depths between 240 metres and 400 metres, with the correction falling off at lesser or greater depths. The temperature disturbance due to this period reaches a maximum of -0.28^{o}C and varies from -0.05^{o}C at 39 metres to -0.08^{o}C at 750 metres. Before this time, the Boreal period, characterized by higher surface temperatures, occurred and lasted from 3000 to 6500 years. Maximum temperature disturbances caused by this effect reach a maximum of -0.03^{o}C at 30 metres to $+0.11^{o}$C at 1200 metres. It is useful to take note of the sum effect of all of the climatic changes of terms greater than 100 years which have occurred since the retreat of the Wisconsin glacier. This total effect is shown in FIG XII-3 and, as can be seen, seems most fortuitous, since between 390 metres and 1200 metres the temperatures vary by only 0.05^{o}C, or about 10%, and the correction to the temperature gradient over these intervals is less than

$0.02^{o}C/km$. Even on some of the very low gradient sections of the boreholes, this effect would place heat flows in error by a negligible amount. Another section of very uniform temperature correction exists between 240 and 360 metres, with a sharper gradient in the lower 30 metres connecting the two sections. However, above 240 metres the temperatures drop very rapidly to a zero correction at the surface, causing temperature gradients of $2^{o}C/km$. Temperature gradient errors of up to 30% in the low gradient, shallow sections of boreholes may result. So far, the effect back only to the retreat of the Wisconsin glaciers has been considered. The effect of Lake Maunee existing over a part of the region is to decrease the disturbance to the temperature gradient by about 5% in the section to a depth of 390 m. Below this depth the gradient disturbance is negligible.

The effect of the glacial periods must now be considered. A step in temperature of $10^{o}C$ is used, ie., the difference between present surface temperatures at London and Guelph of $8^{o}C$ and an ice-base temperature of $-2^{o}C$. Further to the south and west surface temperatures are higher but ice-base temperatures were probably higher also. Several models are considered for the Wisconsin period; a period of ice-cover from 14,00 to 100,000 years ago with one ice-free instadial, the Port Talbot, during which temperatures may have been as at present or as suggested by Terasmae (1961) similar to Northern Ontario today, ie., $3^{o}C$. These models cause disturbances to the temperature gradient varying from about $5^{o}C/km$ at the surface to $1.2^{o}C/km$ at a depth of 1000m. Using Dreimanis' model as discussed previously, the largest contribution is due to the final Wisconsin advance which lasted some 14,000 years. The two other advances of the Wisconsin ice-sheet, lasting 4,00 - 5,000 years, contribute only 14% and 7% respectively, at depths of about 1000 metres. The gradient corrections for the model are lower than the simpler ones but still range from 3.8°C/km at the surface to 0.8°C/km at 1000 metres. Results are shown in FIG XII-4. A very fortunate occurrence is apparent from this model, and this is that the change of gradient correction with depth varies by only 15% down to a depth of 450 metres, making Wisconsin corrections to 500-metre boreholes quite easy. The average correcting gradient is 3.5°C/km. In the section above 200 metres, the correction is about the same as that due to post-Wisconsin climatic changes; below 450 metres the correction becomes progressively less. The temperature disturbance is 2.8°C at a depth of 1200 metres. Between the depths of 1000 and 1500 metres, the gradient is essentially undisturbed by the Wisconsin period as was proved in an earlier section by determining the minimum in the curve. Since the post-Wisconsin changes also did not greatly affect the gradient at depths of 600 to 1200 metres, the ideal section of a borehole in which to make heat flow determinations would be at 1000 to 1200 metres (approximately 3000 to 4000 ft).

The effect of the rest of the pre-Wisconsin ice-ages is to introduce a constant gradient in boreholes to a depth of 1500 m. Thus it's presence must be assumed and can only be validated in deeper boreholes. Gradients caused by the pre-Wisconsin ice-ages range from .63°C/km to .53°C/km at the surface and from .48 to .52°C/km at 1000 m depending on whose chronology is used. These corrections are shown in FIG XII-5.

Each of these calculations has been made assuming a diffusivity of rocks of 0.010 cm^2/sec. For a higher diffusivity the maximum temperature disturbance at any depth and the surface temperature gradient disturbance will be less. However the former will be at a greater depth.

The results of this analysis are subject to errors both in the dates of onset and retreat of the various ice-sheets, in the surface temperatures used for various periods and in the assumptions of homogeneity of the medium with depth. However it does illustrate that large gradient disturbances, of the same order as some of those measured in southern Ontario, can occur and such corrections should not be neglected. The model is used in Chapter XI to correct the measured heat flow for the U.W.O. borehole.

It is useful to look at other parts of Canada and calculated maximum corrections required. Recently Prest (1969) has produced a map of the times of final retreat of the Wisconsin ice-sheet. His data plus the crude approximation that all earlier advances and reatreats occurred at the same time across Canada are used in the following calculations. Differences in advance and retreat times of several thousand years are unimportant at times 100,000 years ago. As shown in the previous section the most recent period of the ice-age contributes by far the greatest portion of the correction, if the ice-base temperature is considered to be the same for each of the ice-sheets. Illustrating this using the Models for southern Ontario, the total Pleistocene correction prior to Wisconsin I is 0.23 V_o ^{o}C/km whereas the single period of Wisconsin from 14,000 to 28,000 years B.P. contributes o.25 V_o ^{o}C/km to the surface gradient correction. Using extrapolated surface temperatures from the uppermost linear sections of the borehole, an ice-base temperature of -2^{o}C and the time of most recent retreat of the ice-sheets, approximate maximum corrections can be calculated using equation xiii) for selected areas as shown in the table following.

Location	Surface Temp. -Ice-Base ^{o}C	Time last Wisconsin Retreat Yrs x 10^{-3}	Correction to Gradient $^{o}C/km$	% Correction
Melville Is., NWT	-16	11	-11.1	-23
Inuvik, NWT	- 4.5	13	- 2.8	-16
Fort Providence, NWT	+ 0.8	10.6	+ 0.6	1
Thompson, Man.	+ 2.0	9.2	+ 1.6	11
James Bay Lowlands	+ 3.0	8.0	+ 2.6	17
Kapuskasing	+ 5.0	9.5	+ 3.8	25
Ottawa	+ 9.5	12.0	+ 6.2	41
Toronto	+10.2	13.0	+ 6.3	37
London	+ 8.6	13.2	+ 6.1	40
Windsor	+10.3	14.0	+ 6.0	40

Table XII-9 DISTURBANCES TO THE GEOTHERMAL GRADIENT CAUSED BY THE PLEISTOCENE ICE-SHEETS IN SEVERAL PARTS OF CANADA

The variation is considerable: -11 to $+6^{o}C/km$ correction to the measured near surface temperature gradients. North of the $-2^{o}C$ surface temperature line of 100 years ago the correction reverses gradient as it was considerably warmer beneath the ice-sheets. This would not be true to the far north-west in the Yukon, which was not glaciated. These corrections may be slightly high since an assumption of continuous ice-cover has been made between the last ice retreat and a time 65,000 years B.P. (ie., no Port Talbot). Certainly the ice-cover was not that persistent in southern Ontario.

The corrections between Windsor and Ottawa are very similar. However variations across Canada are considerable. Birch's corrections for middle latitudes remain reasonable but his conclusions of its relative unimportance do not, since it is not uncommon to have temperature gradients of 7 to 15^{o}C/km in shield areas and also, as is apparent from Chapter VI, in many horizons in a sedimentary basin. It is also apparent that there is a band across Canada, governed by surface temperature, in which the glacial corrections are very small. This presents an ideal zone to drill for heat flow purposes and obtain values with variations dependent only on tectonic differences. The corrections given for Arctic Canada may be too large in view of the expected lower surface temperatures on the ice-caps and Paterson's (1968) observation of ice-bottom temperatures of -16^{o}C for the stagnant Meighan ice-cap and Hansen & Langway's (1966) -13^{o}C for the Greenland ice-cap at Camp Century. Again the important determining factor will be ice thickness, as compared with the ice surface temperature.

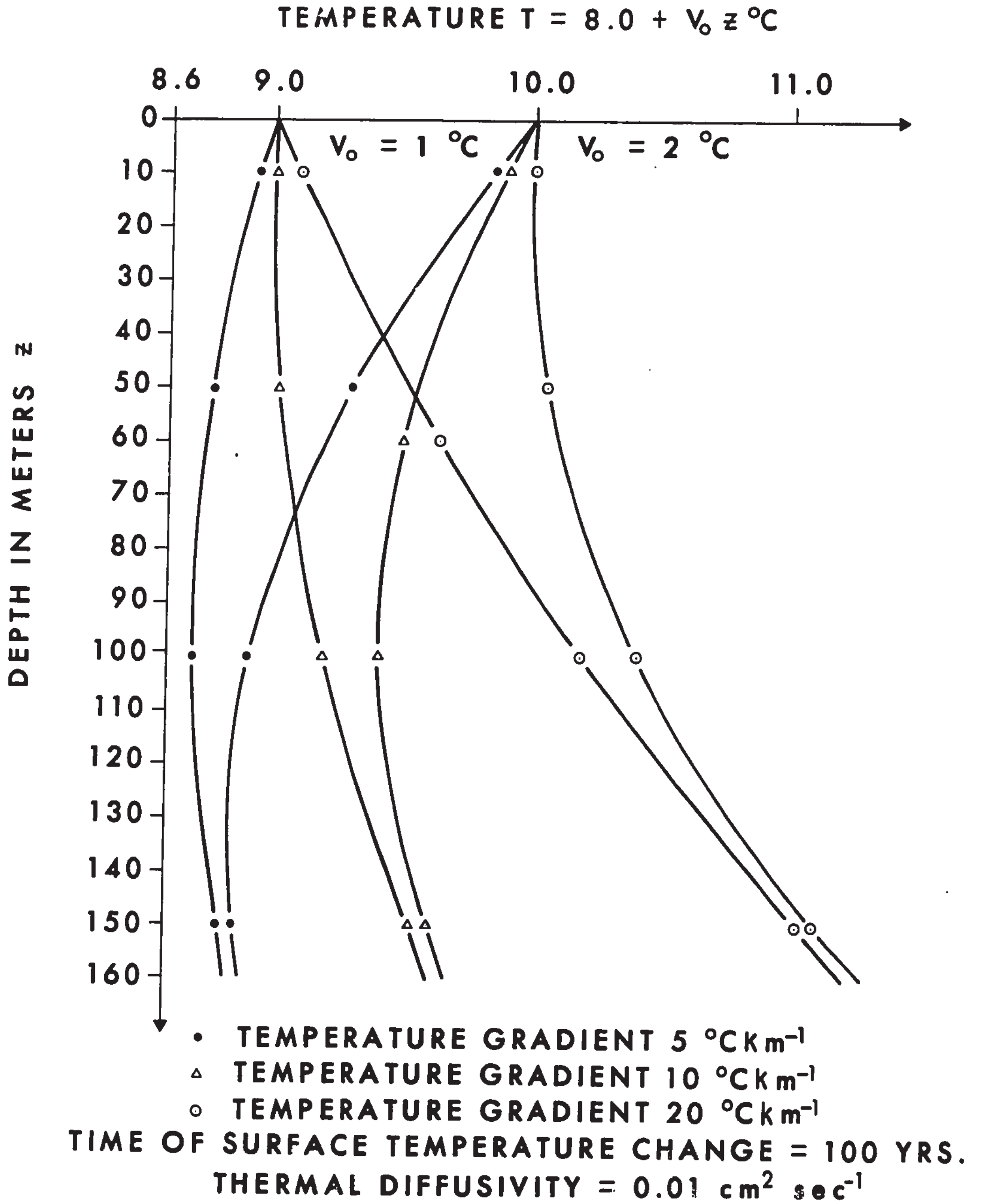


FIG. XII-1

DEPTH OF TEMPERATURE INVERSIONS DUE TO SURFACE TEMPERATURE INCREASE 100 YEARS AGO FOR A RANGE OF GEOTHERMAL GRADIENTS

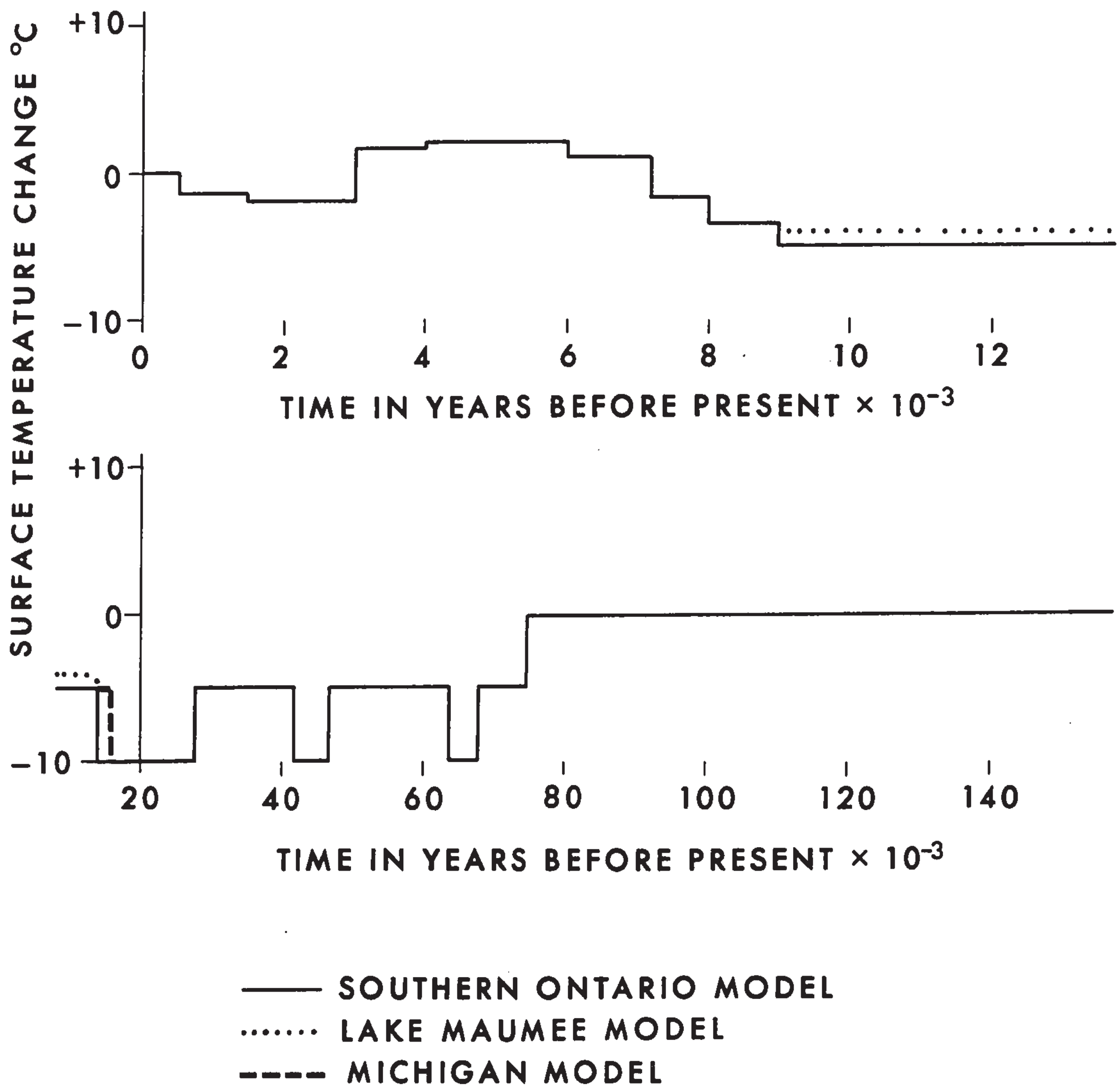


FIG. XII-2

MODELS OF SURFACE TEMPERATURE VARIATIONS DURING THE QUATERNARY IN THE GREAT LAKES

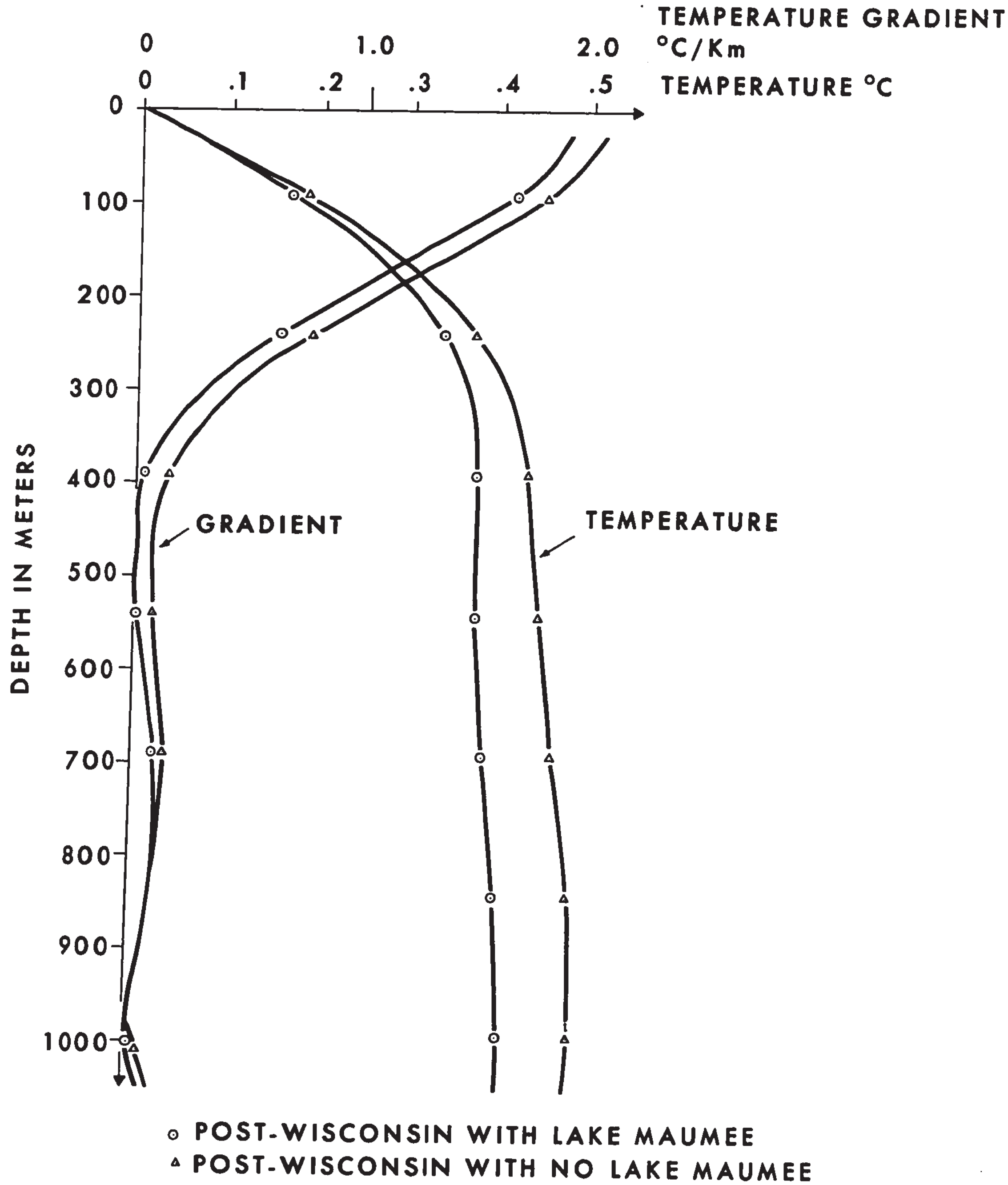
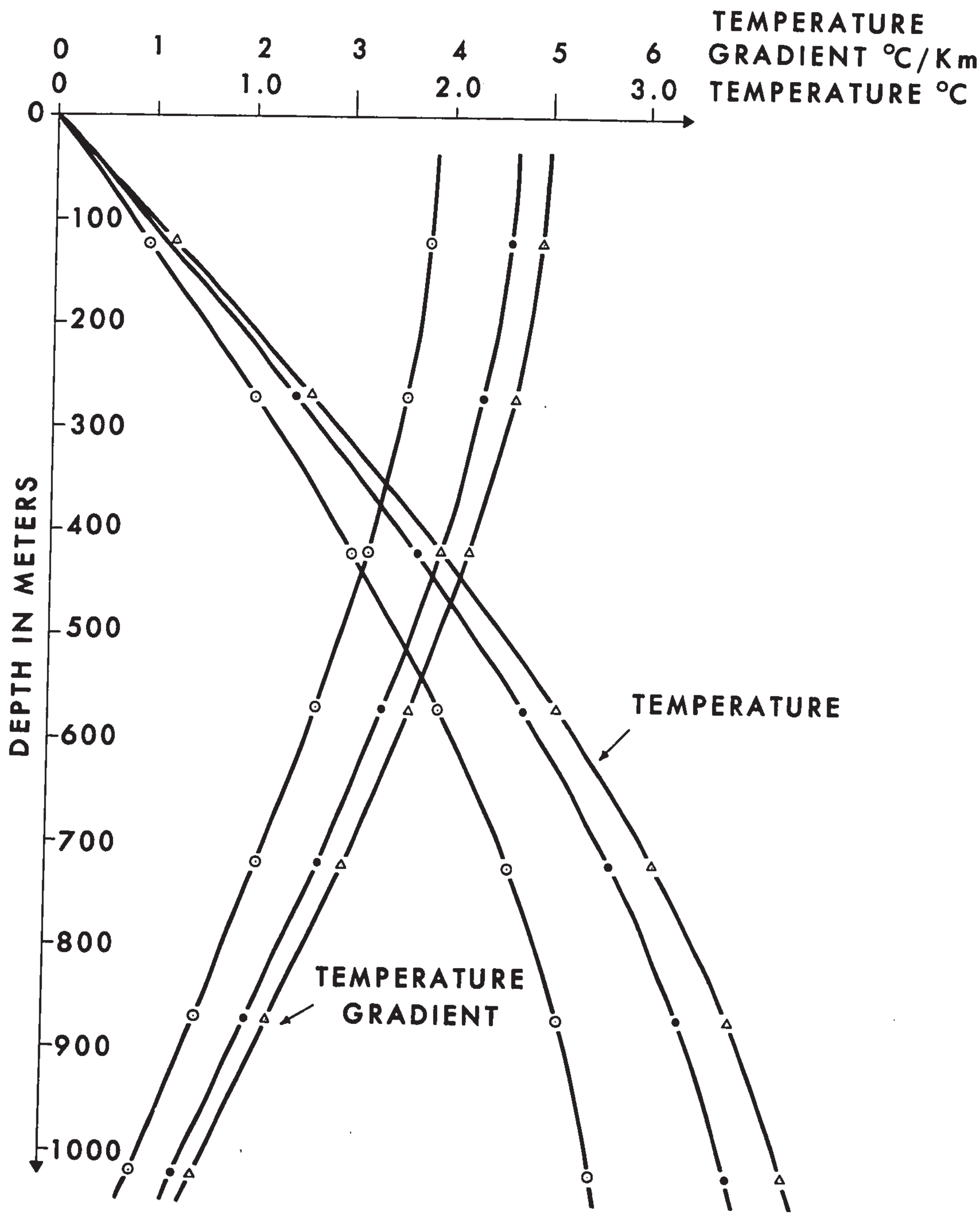
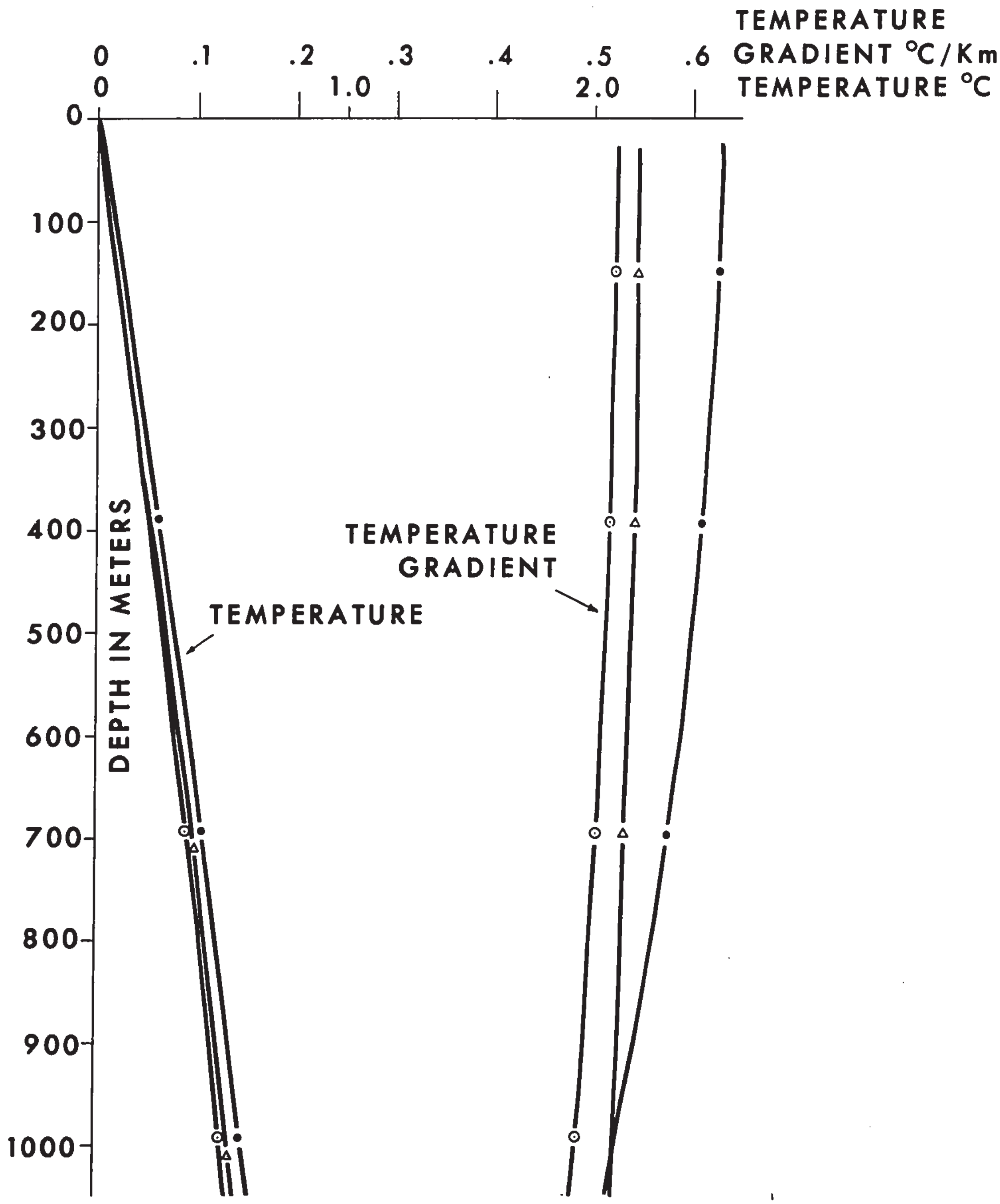


FIG. XII-3
POST-WISCONSIN CLIMATE CORRECTIONS
TO BOREHOLE TEMPERATURES



○ DREIMANIS MODEL WITH COOL INTERSTADIAL
• SIMPLE WISCONSIN-PORT TALBOT CLIMATE AS TODAY
△ SIMPLE WISCONSIN-PORT TALBOT CLIMATE AS NORTHERN ONTARIO TODAY

FIG. XII-4
BOREHOLE TEMPERATURE CORRECTIONS FOR WISCONSIN ICE-SHEETS



• EMILIANI'S TIMESCALE 10°C κ = .01

▵ ERICSON'S TIMESCALE

⊙ BIRCH'S TIMESCALE

FIG. XII-5
PRE-WISCONSIN PLEISTOCENE CORRECTIONS

CHAPTER XIII

CONCLUSIONS AND RECOMMENDATIONS

XIII-1 Summarised Measurements in a Single Borehole

1-1 Return to Equilibrium

i Bottom hole temperatures in diamond drill-holes and cable-tool holes are within 0.1°C and 0.01°C of true rock temperatures after periods of six and sixty hours respectively. This is probably also true of rotary holes if a heavy tool could be used to make contact with the bottom of the hole.

ii In regions of the borehole with low permeabilities, temperature gradients were within a few per cent of the equilibrium gradients after a period of sixty hours since fluid circulation ceased.

iii On completion of the hole the lower sections returned to within 0.05°C of equilibrium temperatures within one month of completion of the hole and the complete hole within three months. Final bottom hole temperatures were within this value after 5 days.

iv In general equilibrium temperature gradients were restored more rapidly than equilibrium temperatures. This has also been observed in many suspended oil-wells drilled by the rotary method where near equilibrium temperature gradients were restored to within a few per cent over a period of several years whereas restoration of equilibrium temperatures to 0.05°C takes approximately twice the interval.

1-2 Single Borehole Heat Flows

i Detailed temperature and thermal conductivity measurements in a single borehole have shown that heat flow may vary with depth by as much as 20% about a mean.

ii Reducing the number of sample points uniformly over the whole hole only alters the mean heat flow by 5%. However real variations with depth become masked with a decrease in the number of samples.

iii This suggests that many of the borehole results listed in Lee and Uyeda (1965) although reasonable averages, may contain unsuspected variations with depth. These variations are masked within the error limits.

iv Heat flow measurements over a single short interval in a borehole are better than no results but should not be considered more reliable than 20%. If this technique were used in South Western Ontario the most often cored sections are the reservoir rocks and hence the Guelph-Salina A1 would be a logical choice. The resulting heat flow in this borehole would have been unusually low. At least three intervals should be selected.

v A similar danger exists in 'in-situ' measurements. Garland and Wright (1968) obtained an 'in-situ' thermal conductivity at a single depth in a hole on Prince Edward Island. Combined with the temperature gradient over the interval it led to a heat flow of 0.98. However a more detailed analysis of the complete borehole by the author (Jessop and Judge, in press 1971) has suggested that the zone of the earlier measurements is disturbed by a small water-flow in the porous sandstones and the equilibrium heat flow for the complete hole is 0.88. Comparison of in-situ and divided-bar results permitted an estimate of the water-flow rate.

XIII-2 Regional Surveys

i Much care must be taken in selecting geological horizons such that they are continuous and similar in composition across the region. More than a single horizon must be studied. Although down-hole variations in thermal conductivity in a given formation may be a few hundred per cent, the standard deviation on the compared means between different boreholes may be of the order of ten per cent. Table XIII-1 shows the mean formation conductivities.

ii The standard deviations on the mean formation temperature gradients, shown in Table XIII-2 following, are larger than the conductivity deviations possibly indicating real variations in heat flow.

iii A mean regional heat flow for all the sites is 1.0 ± .2 μcals cm^{-2} sec^{-1}. The mean heat flow for all of the sites in southern Ontario and Michigan between the Frontenac Axis and the Grenville Boundary is 1.00 ± .12 with a range of 0.82 to 1.22 in 21 sites. Taking all available measurements in sediments on Grenville Basement, the mean is 0.97 ± .13 ranging from 0.74 to 1.28 over 25 sites compared with the exposed Grenville for which the mean heat flow is 0.89 ± .16 in 8 sites with a range of values of 0.79 to 1.22. The difference is not significant.

TABLE XIII-1 THERMAL CONDUCTIVITIES IN SOUTHWESTERN ONTARIO - INTERBOREHOLE COMPARISONS

Formation	Rock Type	Conductivity				No of Holes	Extent (km)
		Max	Min	Mean	Std Dev		
		mcals$^{o}C^{-1}cm^{-1}sec^{-1}$					
Hamilton	Shaley limestones	6.2	5.1	5.7	-	2	80
Delaware	Limestone	7.4	6.4	7.2	0.3	9	200
Detroit River	Limestone	9.4	7.2	9.0	-	3	100
Bois Blanc	Cherty limestone	9.1	7.8	8.5	-	3	100
Total undivided	Limestones, dolomite chert	9.4	7.0	8.1	0.7	14	200
Bass Island	Dolomite	13.3	8.7	10.7	1.4	9	200
Salina total	Mixed	10.8	8.8	9.6	0.8	4	200
Guelph	Dolomites	11.8	8.1	10.7	1.0	8	200
Clinton-Cataract	Mixed	9.3	4.9	7.3	-	3	80
Queenston	Shales	-	-	5.3	-	1	-
Meaford	Siltstone and shaley limestone	-	-	6.3	-	1	-
Collingwood	Shale	-	-	4.4	-	1	-
Trenton-Black River	Limestones, dolomites in west, shalier in east	6.7	5.0	6.1	0.6	6	600
Cambrian	Sandstones	12.1	11.0	11.3	-	3	300
Precambrain	Gneisses	7.5	4.4	7.1 (unweathered)	0.1	11	200

TABLE XIII-2

TEMPERATURE GRADIENTS ($^{o}C/km$) IN SOUTHERN ONTARIO AND MICHIGAN

Formation Name	Range	Mean	Stnd Dev.	No of Holes	Extent (km)
Precambrian	12.8 - 15.4	14.1	1.1	3	400
Cambrian	-	8.5	-	1	-
Trenton-Black River	14.8 - 24.6	18.4	2.6	12	670
Collingwood	30.0 - 45.0	36.7	4.3	5	600
Meaford-Dundas	19.7 - 31.5	26.1	3.6	7	300
Queenston	15.7 - 20.7	18.0	1.9	5	320
Clinton-Cataract	17.0 - 17.7	17.3	-	2	60
Guelph-Lockport	4.9 - 7.5	6.6	1.1	4	220
Salina complete	7.6 - 12.5	9.3	1.4	10	500
Salina upper	7.9 - 13.1	9.0	1.4	12	500
Salina Evap.	6.2 - 9.2	7.5	1.2	5	130
Bass Island	5.9 - 12.8	8.9	2.2	10	400
Bois Blanc	8.5 - 10.5	9.7	0.9	3	60
Sylvania	8.1 - 9.9	9.3	0.8	3	60
Detroit River	7.5 - 11.5	8.9	1.6	7	300
Dundee	-	-	-	-	-
Traverse	10.2 - 21.0	14.8	3.9	5	290
Antrim	26.9 - 49.2	34.4	8.1	5	290
Coldwater	20.3 - 38.0	25.9	7.2	4	200
Michigan	18.4 - 28.9	24.4	3.2	6	80
Saginaw	18.7 - 25.6	22.3	3.0	6	80
Red beds	18.7 - 21.7	20.2	1.5	2	40

FIG XIII-1 shows the variation of heat flow in the Great Lakes region, and Table in chapter VIII-4 compares the means with those in other parts of North America.

iv Since the heat flows across the region are fairly constant the temperatures at depth are governed chiefly by formation conductivities and mean annual ground temperatures. This simple picture is confused however, by shallow inversions in the temperature curves with depth.

v The existence of temperature inversions with minima as deep as 100m makes the use of boreholes less than 200m deep for heat flow determinations pointless.

vi In zones of high thermal conductivity and very low temperature gradient the corrections for past surface temperature changes using reasonable models may be as high as 40% of the measured gradient. However, real effects seem to be smaller than the predicted ones.

vii The regional heat flow across southern Ontario is typical of shield values but there are indications of increased heat flows in the Michigan Basin in which the basement is largely in a different metamorphic province.

XIII-3 Recommendations for Future Work

i It is imperative that since a liason has been established with ODERM that the work of logging boreholes about to be plugged be continued. This ought to be done whether or not the hole is of particular value to a heat flow programme.

ii The work should be extended to the bottoms of Lake Ontario and Lake Huron to increase the regional coverage and to tie in with work being done in Northern Ontario in particular.

iii Priority should be given to determining the radioactive heat production of basement rocks throughout the region, since it covers several different geological provinces. Comparison with other geophysical methods to reveal crustal thicknesses and composition would also be of value.

iv South Western Ontario is a fortunate sedimentary area in that many stratigraphic tests have resulted in continuous cores. In the province of Alberta where there is a great deal of fairly reliable temperature data (Anglin & Beck, 1965; Alberta Oil & Gas Conservation Board, 1967), there are two such continuous cores. Thus a great deal of detailed work on thermal conductivity versus density, porosity, pore structure and rock composition remains to be done. Following on from this, work is required on the estimation of thermal conductivity from well-logs taking advantage of the refined logs available today which Joyner (1960) did not have for his work.

v It is doubtful if one borehole in which both temperatures and conductivities have been measured is sufficient for a regional study in which partial measurements are being used. Tubing can be purchased for less than $ .60 per foot and thus more effort should be made to know of stratigraphic test-holes in advance by closer cooperation with provincial authorities and them to keep them open for temperature measurements.

Costs per 2000 ft hole would not amount to more than $2000.

vi Long term studies of ground and air temperatures are needed in conjunction with permanent cables for temperature measurements to depths of 600 ft. in order to correlate more closely the temperature inversions with changes of climate.

vii Several 10,000 ft boreholes are needed in various parts of Canada in order to determine the true effect of the Pleistocene ice-sheets on crustal temperatures. They should be chosen so as to have a simple and uniform stratigraphy and to be well away from any large bodies of water and should be distributed with one in the Arctic, one in southern Canada and one close to the present permafrost boundary. One such hole exists but unfortunately was inadequately cored.

viii A major uncertainty in the glacial corrections is the ice-base temperature, a problem which will be helped by more deep holes in ice-sheets and more theoretical work on the temperature distribution in ice-sheets.

ix Detailed studies of the hydrological situation in a sedimentary environment are required. These would involve analysis of formation waters for C^{14} and Tritium, the construction of sensitive flow-meters and perhaps a greater use of in-situ conductivity measurements to complement the divided bar ones.

x A great deal of work still needs to be done on the factors that disturb equilibrium heat flows before geothermal measurements reach the stage of mere collection of data for gross tectonic studies as Roy et al (1968) suggest that they have already reached.

xi In general, temperature and thermal conductivity measuring equipment is sufficiently accurate to warrant a great deal of emphasis, now be given to detailed borehole measurements and more careful analysis of the obtained results than has often been attempted in the past.

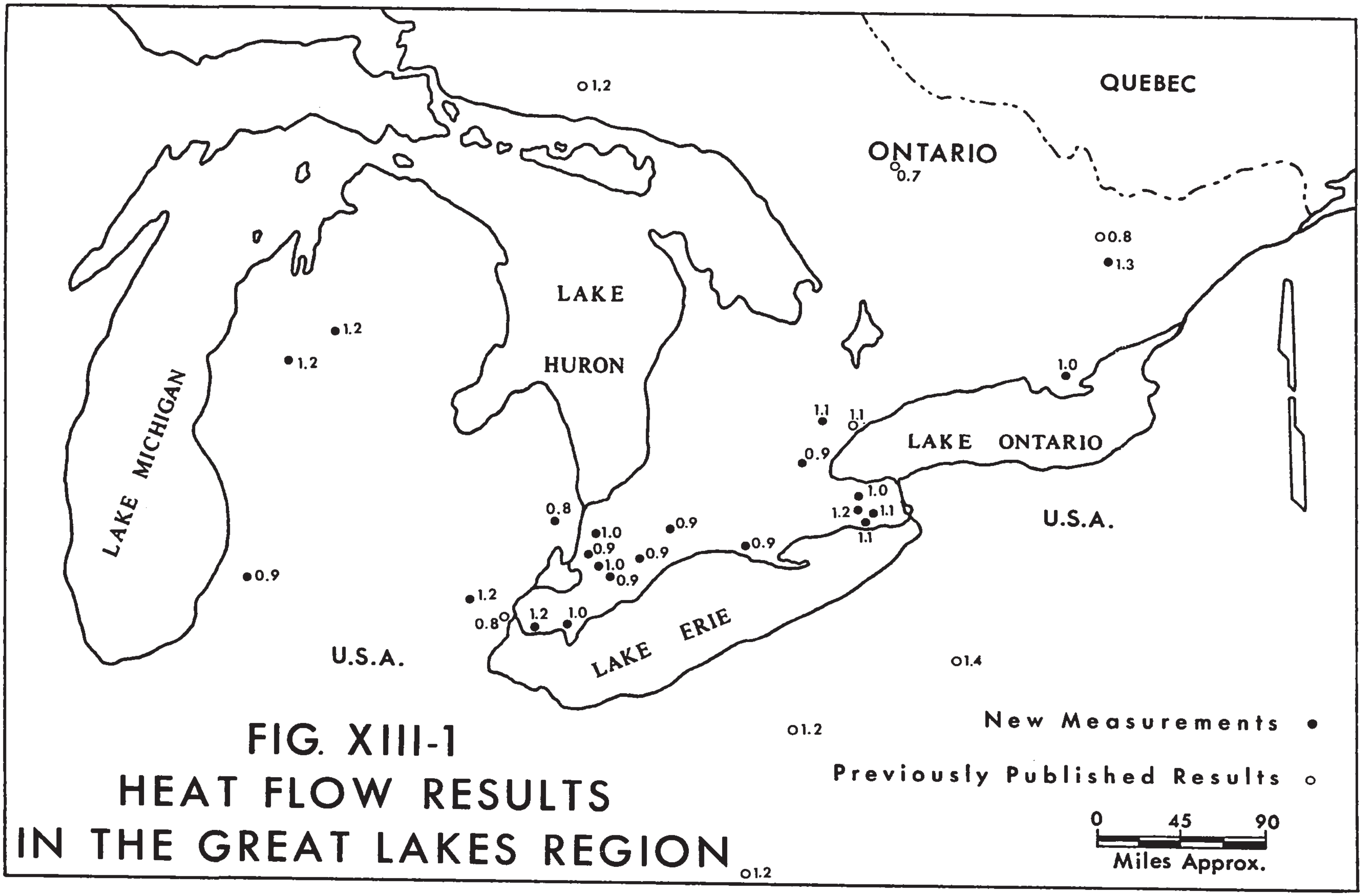


FIG. XIII-1
HEAT FLOW RESULTS
IN THE GREAT LAKES REGION

APPENDIX I

DESCRIPTIVE LOG

OF THE LITHOLOGY OF THE REGION

PREFACE

This description is not meant to be complete but rather to give a very general, brief introduction to the lithology for a reader who is unaware of the rock types associated with particular names mentioned in the text.

TABLE A-I-1 TABLE OF FORMATIONS

Period	Formation	Lithology	Maximum Thickness (m)
In Michigan Peninsula Only			In Michigan
Jurassic	'Red Beds'	Red-coloured shales and sandstones; poorly consolidated	65
Pennsylvanian	Grand River	Red and white alternating coarse quartzose and sandstones with quartz or ferruginous cement	}230
	Saginaw	Micaceous sandstones, shales and coal beds, very white quartzose sandstone at base	
Mississipian	Grand Rapids Group	Quartzose sandstone, greenish grey shale with limestone interbeds, some gypsum and anhydrite, blueish limestone and dolomite at top	230
	Marshall	Quartzose porous sandstone	100
	Coldwater	Grey, greenish-grey shale with limestone lenses grading to siltstone & fine grained sandstone	395
Mississipian -Devonian	Ellsworth	Greenish-grey silty shale only found in west	200
	Antrim	Grey to black calcareous shale with interbeds of brown argillaceous limestone in west to black calcareous shale in east	200

Table A-I-1 continued

Period	Formation	Lithology	Maximum Thickness (m)
Mississipian -Devonian	Traverse Group	Dolomites, cherts, light grey limestones, crystalline fragmental limestones, some quartz sandstones	255
In Southern Ontario and Michigan (Michigan equivalent not named where different)			In Southern Ontario
Devonian -Mississipian	Port Lambton	Grey fissile shale and dolomitic limestone, black fissile shale on top	60
	Kettle Point	Black biuminous fissile shale with interbeds of green shale	100
Devonian	Hamilton	Grey calcareous shale and shaly limestone	90
	Dundee	Lower buff, crinoidal and arenaceous limestone with chert, above buff crystalline limestone with chert and interbedded black bituminous shale	60
	Columbus	Calcareous samdstone, sandy limestone, dolomite	10
	Detroit River	Brown, finely crystalline, to granular dolomite, interbedded limestone	90
	Bois Blanc	Limestone, sandy limestone dolomite, chert	80
	Oriskany	Light grey, glauconitic sandstone	5
Silurian	Bass Island	Grey-buff finely crystalline dolomite - oolitic zones	10-120
	Salina	Dolomite, limestone, dolostone, shaly dolomite, shale, anhydrite, gypsum and salt	90-455
	Guelph-Lockport	Guelph - buff crystalline dolomite Lockport - grey crystalline limestone Amabel - brown, bluish-grey dolomite	20-40 50-65 20-35

Table A-I-1 continued

Period	Formation	Lithology	Maximum Thickness. (m)
Silurian (cont'd)	Clinton	Grey quartz sandstone, greenish black shale, grey crystalline dolomite, black calcareous shale, grey argillaceous dolomite	8-35
	Cataract	Sandstone, brown crystalline dolomite, red and green shale, arenaceous shales with sandstone interbeds	20
Ordovician	Queenston	Red shale	295
	Meaford-Dundas	Grey shales, thin limestone beds, interbedded grey shale, silty limestone, dolomite Shale	175 35
	Gloucester Collingwood	Soft, grey and brown shales Black fissile, calcareous and bituminous shale	25 5-20
	Trenton-Black River	Carbonate sequence fine crystalline - sub-lithographic limestone, overlain by finely crystalline limestone, with chert and bentonite, shale limestone	265
Cambrian	Eau Claire St. Simon Jacobsville	Arkose, grey, green, purplish samdstone, buff dolomite	135
Precambrian		Metamorphosed rocks	

References:

Jodry (1957)

Kelly (1936)

McGregor (1954)

Pearson (1963)

Sanford & Brady (1955)

Winder (1961)

APPENDIX II

PROPERTIES AND CALIBRATION OF THERMISTORS

A II-1 PROPERTIES OF THERMISTORS

The thermistor is a small, mechanically rugged, semiconductor element with a negative temperature coefficient of 4-5%. To a good approximation, a relationship may be written expressing its temperature-to-resistance dependance of the form

$$R_T = A \exp\left\{B(T + C)^{-1}\right\}$$

where A, B and C are constants of the thermistor and R_T is the resistance in ohms at the absolute temperature T. It is available in a large range of resistance and voltage ranges, and is unaffected by quite large changes in its chemical and physical environment. Certain problems exist with thermistors; for example, Beck (1956) and Schwarz (1961) have found that thermal shock may change the characteristics, thus suggesting that a heat sink should always be used during lead soldering. General stability in thermistors is a very important topic since it limits their use in absolute temperature measurement. Drifts with time occur in the resistance-temperature characteristics, usually in the direction of decreasing apparent temperature. For temperature measurements in this thesis, the calibrations were checked at two or three temperatures before and after each field trip. It is important that readings be checked at several temperatures, since all constants may vary. The drift which occurs is probably due to lattice diffusion of impurity ions. Wendon (1957) suggested such an effect for the change in resistivity of crystalline quartz. Beck (1956) discusses the size of drift, noting that it decreases with time and is slower in rarely used sensors. This seems to add weight to the idea of diffusive processes. Rates of drift in Beck's paper are very large, -30 to -210 × 10^{-3} °C/month, when compared to modern values. Annealing has been introduced recently as an aging process (Nielson, 1959), probably accounting for this discrepancy. In the present work, drift rates were noted on twelve thermistors for periods of up to three years. Most of these were of the glass probe or bead types produced by Fenwell. Drift rates varied from 0 to -60×10^{-3} ^{o}C/month over the first nine months of use but then decreased with increasing time to 0 to -5 ^{o}C/month after three years. Swartz (1966) quotes some unpublished results of Diment on glass probe mountings, the drifts being of the same order as our own. Amongst our results thermistors that had been initially used in heat dissipation tests up to 15 volts showed lower drifts, 0 to -5×10^{-3} ^{o}C/month. Several silicon sensors were also examined for drift, but, although Lister (1963) has used them successfully for oceanographic work, the U.W.O. sensors showed average drift rates of 20×10^{-3} ^{o}C/month at 5^{o}C, to 80×10^{-3} ^{o}C/month at 25^{o}C after six months. Without such large drifts their much lower dissipation constant would have been an advantage, permitting the use of higher bridge voltages and thus decreasing the necessary detector sensitivity.

Recently Swartz (1966) has published ice-point drift rates over a ten-year period of a 645 Western Electric 17A disc thermistor. Average drift rates were 5×10^{-3} ^{o}C/month which, when compared to Beck (1956) for similar types, illustrates the reduction of drift rate with time.

A II-2 SHOCK EFFECTS

As previously mentioned, several authors have described the effect of sudden mechanical or thermal shock on thermistor stability. However, no such effects were observed during this work even though the probes were occasionally used to pound past obstacles blocking boreholes.

A II-3 PRESSURE EFFECTS

Unprotected thermistors in fluid-filled boreholes are subject to pressures substantially higher than those at the surface. If the thermistor is responsive to pressure as well as temperature changes, serious errors in temperature gradients might result in the lower portion of a hole. Several workers have examined these effects, namely Tavernier and Prache (1952) who found coefficients of 47×10^{-7} bar^{-1} for Compagnie Generale de Telegraphie San Fils disc types under pressures up to 4400 bar, and Misener and Thompson (1952) who found coefficients of 0.9 to 1.4×10^{-7} bar^{-1} for Western Electric 12A rod thermistors under pressures up to 130 bars. Since this time, small glass beads have been introduced and these might be expected to be more seriously affected by high pressures. The pressures involved for heat flow work in this area are, for the most part, up to 100 bars. With a borehole on campus, it was decided that the easiest way to determine such effects would be to lower a probe containing two thermistors, one protected in a pressure-tight housing and the other open to borehole fluid. The results obtained indicated that the pressure effect caused temperature differences of less than .005°C, suggesting a pressure coefficient less than 3×10^{-7} bar^{-1}. (Accuracy of these temperatures is not as great as more recent results because of the very leaky cable which had to be used.) This compares favourably with Diment and Robertson (1963) who tested glass probes up to 700 bars and found a pressure coefficient of 1×10^{-7} bar^{-1}. These values indicate that the effect is negligible except in very deep boreholes and the small effect is reversible up to 700 bars (Diment & Robertson, 1963). This is equivalent to a vertical water-filled borehole 7000 m long or a mud-filled hole of 5000 m.

A II-4 DISSIPATION AND TIME CONSTANTS

The dissipation constant of thermistors which expresses the ability to dissipate the resistive heat generated by an applied voltage, is an important factor in their selection and in the design of probes. Disc and rod thermistors have much higher dissipation constants than beads or probes, but also have longer time constants. As a result of laboratory tests on various types of thermistors the bead types were selected for general use as it was found that bridge voltages of 0.5 volts could be used and the self-heating effect kept below 0.01°C, and at 0.3 volts below .003°C. This self-heating effect is further reduced by improving the conduction of heat away from the sensing element. For this reason it was sprung into contact with the probe metal and a drop of oil added to reduce the contact resistance.

Of equal importance is the time constant of the system which expresses the time for the system to reach a new thermal equilibrium when the temperature of the surroundings is changed which is, of course reduced by using a smaller sensing element of smaller thermal mass. In this way the dissipation and the time constant require opposite conditions. However, the time constant of the thermistor itself is generally swamped by that of the containing probe. In the sense that a high probe conductivity lowers the time constant, the two factors act together. Since logging was done at intervals rather than continuously, a 2-3 minute time constant was considered as an upper limit. With this order of equilibrium period, the probe was at constant temperature as soon as the operator had managed to null the bridge.

AII-5 CALIBRATION PROCEDURES

The thermistors were either encased in small brass tubes filled with oil and sealed with paraffin wax where the wires emerged, or, if already mounted, were calibrated in their probes. Each one was placed in one of the twelve holes drilled in a circular aluminium block about eight inches in diameter. This block sat on a polystyrene insulator in a Colora 20S constant temperature bath with a range of -20°C to 40°C. The short-term fluctuations in the bath over several minutes were of the order of 0.015°C, which was reduced to 0.004°C in the aluminium block. A platinum resistance thermometer (Leeds & Northrup 126619) was placed in the centre of the block. An ice-point determination indicated a drift from the original claibration by the National Bureau of Standards of about 1%, and a later triple-point calibration at U.W.O. confirmed this variation. The platinum resistance values were read on a Mueller Bridge using an Astrodata TDA-121 Nanovolter as a null detector. Platinum readings were checked after every two thermistor readings at the same temperature. Thermistor resistances were measured on a Pye Wheatstone Bridge 7384, using an Alinco ND1 null detector as the galvanometer and a bridge voltage of 0.3 volts. Each thermistor was calibrated at 2°C intervals over the range 0 to 30°C. The platinum resistances were converted to temperatures using a programme based on Werner and Frazer (1952). Thermistor resistance and platinum resistance temperatures were fitted to a computer programme to print out temperature values for 10 ohm resistance intervals. For the purposes of this work the standard thermistor equation of Bosson, Gutman & Simmons (1950) was modified to the approximate form:

$$\ln R_T = A - BT - CT^2$$

and data fits made over 6^{o}C intervals. Worst deviations at the centre of each range amounted to several thousandths of a degree. Recently Steinhart and Hart (1968) have suggested a logarithmic expansion in log R_T as giving a better fit; however no comparisons have been made using both equations on the author's data. An estimated accuracy of 0.005^{o}C is given to the thermistor readings, which probably is improved to 0.003^{o}C by the smoothing effect of the programme. Platinum resistance temperatures are probably accurate to 0.002^{o}C. Once the thermistor has been calibrated absolutely, interchanges between cables can be accomplished without recalibration.

APPENDIX III

TEMPERATURE LOGGING EQUIPMENT

AIII-1 PROPERTIES OF CABLES AND BRIDGES

An ideal cable would be one possessing infinite open-circuit resistance, a zero short-circuit resistance, zero temperature coefficient, perfect rigidity, no weight and complete inertness to the corrosive solutions such as brine. In practice the various types of cable used are compromises of these factors. Cables are manufactured with very low series resistance and thus low temperature coefficients but they are heavy and need to be truck-mounted, an acceptable situation in deserts or populated farmlands. In many parts of Canada, such as the Shield, the Arctic or even in shrub-covered regions in areas of greater population density, the equipment must be light for transport by air or for back-packing on foot. In these circumstances, some of the advantageous properties must be sacrificed for portability.

Three types of cable were used in the course of this project, all designed by Northern Electric. The first used was a very leaky 1800 m cable with a laboratory shunt resistance of 1 megohm and a series resistance in each lead of approximately 400 ohms. A three-lead stranded copper cable, each lead coated with nylon and cotton with an outer coating of rubber, it was well able to support its own weight in a 1500 m borehole with an extension of only a few meters. The second was a light-weight cable with a series resistance of 320 ohms over 2350 m and a shunt resistance over 500 megohms between each lead. Each of its three leads was 26-gauge solid copper with a polyethylene coating and all three leads enclosed in an outer sheath of polyvinylchloride. The weight was approximately

1 kg/100 m, with a cable extension maximum of about 1% in 1000 m. After use in a 1200 m borehole, permanent extension was about 0.3 m in the top 30 m, where most of the weight was taken up. This suggests that, for very accurate work, the cable should be pre-stretched in a borehole before marking and should not be wound on reels under tension. Breaking strain was approximately 18 kg, so that 1200 m lengths may safely be used in water-filled boreholes. A very similar but heavier-duty version of the above was also used; in this, 22-gauge solid copper leads with thicker insulations were used, bringing the weight to 5.4 kg/300 m and the breaking strain to 27 kg. This cable had a shunt impedance of over 1000 megohms and a series impedance of 55 ohms in 1040 m. The latter two were used very extensively in the later work, but after a period of several years holes developed at spots in the light-weight cable where the outer sheath was thinner than average. Shunt resistance had dropped to 50 megohms. To obtain accurate temperatures as these cables aged, it was important to know the changes in values of shunt resistance. The heavier cable still showed values of over 500 megohms after frequent use; it was decided that it could safely be used in a bridge with a three-lead configuration to compensate for the series resistance of the cable. To what extent such generalized circuits might be used was examined.

The properties of the leaky cable were examined in two boreholes to find out what requirements might be necessary in a bridge designed for use with such cheap cables, and which could be given to borehole drilling crews to make measurements during and immediately after the drilling, and then thrown away after a few holes. In the U.W.O. borehole, the series resistance dropped by some 4 ohms in the upper part and then was constant from 183 m to the bottom. Similarly, in the Morriston hole the series resistance changed by only 2 ohms between 183 m and 914 m. These two results tend to suggest that there may be a drop of 3 to 4 ohms in the top section, but very little after that. This change occurred in each pair of leads tested and thus seemed to verify the idea that one set of leads could be used to compensate the thermistor leads. Any initial difference in the resistance of the two sets of leads, usually one or two ohms, would persist down the borehole. It would seem that a check on the variation of series resistance in the borehole is not necessary if 10 kilohm thermistors are used and an accuracy of about 0.01^{o}C is required, and hence compensation is valid.

Rather different situations occurred with the open-circuit resistance where variations of the order of 20% were observed on three occasions in measurements to 1000 m depths. On one occasion in the U.W.O. borehole with the probe stationary at 300 m, a shunt variation of some 20% was noticed as a thunder and hail storm passed over. Tests on the dependence of the bridge resistance on ambient temperature were too small to explain the differences, and thus it was concluded that a telluric pickup was being received in the high resistance circuitry. Thus with leaky cables the variation of shunt resistance is large enough to warrant its occasional measurement. Such are the points from which the bridges of Beck (1963) and Jessop (1965) must be considered.

Beck (1963) describes a modified Wheatstone Bridge circuit in a 4-lead cable which is used to eliminate series and shunt resistances by compensation in opposite arms of the bridge. Disadvantages of the system are the increased number of leads, the assumption of constant temperature coefficients in the cable, the need for small corrections to the thermistor resistance to correct for assymmetry of the cable lead resistances and lack of knowledge as to what is happening to the shunt impedance. In response to this Jessop (1965) described a relay system. Although it only uses three conductors it requires three independant measurements of series, shunt and thermistor resistances which increases on-site time and requires tedious calculations to be made. Measurement of the shunt impedance can be very time-consuming as the relay current charges up the cable which acts as a capacitor with a long time constant.

AIII-2 SHUNT DIALING BRIDGE

Taking into account the bridges which had already been constructed and the following requirements:

a) simple operation,

b) a value of the shunt resistance be obtainable,

c) the bridge decade to read the thermistor resistance directly,

it was decided to construct a slightly different bridge, utilizing some aspects of both bridges.

Initially, shunt compensation was considered with a single fixed shunting resistance placed in parallel with the decade box, but this resulted in errors of $1^{\circ}C$, illustrated in FIG AIII-1, where R_m is the measured thermistor resistance and R_T the true thermistor resistance. In designing a bridge for the leaky cable, it was found that placing a 5-megohm composition potentiometer in parallel with the bridge decade and adjusting it as near as possible to the shunt resistance of the cable gave good compensation over the range 0 to 40°C. Since errors of reading the resistance of an uncompensated 10-kilohm thermistor persist up to cable open-circuit resistances of up to 60 megohms, switches were necessary to compensate resistances up to 50 megohms. This allows reading accuracies of .005°C with a light-weight cable, and yet leaky cables of low open-circuit resistance can still be used to accuracies of .02°C, since the 5-megohm potentiometer could be resolved to .2 megohms.

FIG AIII-2 shows the complete circuit diagram for the bridge. The bridge is designed for use with three lead cables; two leads are used for the thermistor and the open-circuit reading and the third lead for the series lead resistance compensation and for the power lead to the coil of the relay in the probe, which switches between open-circuit and the thermistor. A function switch, consisting of a 2-deck rotary in its first position, switches the relays and connects 12 volts on the bridge to balance the shunt resistance against the open-circuit cable resistance. In its next positions, it switches the relays again and connects the thermistor to the cable. Now in this position, unless switch S is open, the shunt compensating potentiometer is also switched in across the thermistor balancing decade, effectively compensating the cable shunt resistance. This is valid only if the shunt resistance is not fluctuating rapidly. As can be seen, the bridge can also be used without this compensator in circuit, reverting it to a simple 3-lead series compensated system for high shunt impedance cables. If the switch S is closed, the resistance reading on the output is the thermistor resistance and the calibration charts may be used directly to derive temperatures in the field without any intermediate calculations. If the shunt compensating circuit has been calibrated, the shunt resistance may also be recorded as a permanent record of cable performance.

The factor of cost must be considered if this is to be a reasonable expendable piece of equipment. Decade resistance boxes with .01% accuracy and a temperature coefficient of a few parts per million are very expensive. Therefore, instead of using one of these, a Spectral series 800 1-kilohm wire-wound potentiometer was used, together with a Model 25 Multidial to count the number of turns. This combination gave an accuracy of 1 ohm between the smallest division on the dial.

Usually thermistor resistances are greater than 1kilohm, so that an extra decade is required. Suitable for this purpose is the Mallory decade switch requiring only four resistors, 1, 2, 3 and 4 kilohms, to produce a 1 kilohm increment up to 10 kilohms. For a 5-kilohm thermistor, this would produce a useful temperature range of 4 to 40°C. It is fairly easy to arrange to switch in further steps of 10-kilohms should they be desired.

The bridge was nulled using a Fluke Electronic Galvanometer Type 840. The relay used for switching between cable open-circuit and the thermistor was a SPDT 12V latching type built into a TO5 transistor case by Teledyne (Type 1053). Selection of the relay coil is dependent on the length of cable and its series resistance. This type of relay, however, permits a large reduction in the size of probe over that used initially with Branson latching relays (Type LSA - 20 - 24A).

Bridge procedure was to plug in the cable connection from the winch and switch on the galvanometer. With the rotary switch in the shunt position, the bridge was balanced using potentiometer P1, a null position on the galvanometer being an indication of this. The rotary dial was twisted to thermistor position. Again the null was found, this time using the Multidial P2. For the light-weight cable it was sufficient to check P1 at top and bottom of hole, whereas for leaky cables, it was necessary to check at each position. The reading was the true reading of the thermistor, and when converted was the true temperature of the sensor. Bridge voltage applied to the thermistor was checked on the galvanometer dial by means of push-button switch SV.

AIII-3 OSCILLATOR PROBE

In most electronic communications systems the best way to transmit information that suffers a serious loss in signal strength, in some manner which cannot be described in an exact fashion, is to frequency modulate it at the source and demodulate after passage through the medium. Thus, with leaky cables, it was considered that the best method of determining temperatures in a borehole might be to use a temperature-sensitive oscillator. Such methods have been used extensively in oceanography and biophysics (Campanella, 1961, Kurtenbach, 1965). Various methods have been tried for logging boreholes, from the vibrating wire of Guelke et al (1949) to the quartz thermometer of Hammond et al (1964). The cheapest and most simple approach is to use an R-C oscillator in which the resistances are thermistors, and thus modulate the frequency of oscillation. Doig et al (1961) described a phase shift oscillator that worked on this principle; however, they found very serious drifts in frequency with time whilst the probe was held at a constant temperature. The reason for this was probably to be found in the lack of any voltage stabilization in the circuit and in the use of germanium transistors.

Several types of R-C oscillator exist which might be pressed into use for this type of work, but to make any decision they have to be compared on the basis of number of components, power needed and operating stability. The two major types of R-C oscillator are the so-called 0^o and the 180^o phase shift varieties. The commonest forms of each one are the Wien Bridge and the Phase Shift types, respectively. The main advantages of the former are that it requires less gain to go into oscillation and it has good stability at higher frequencies. With the high gain preamplifier transistors available today at relatively low prices, the first advantage is not so great and high frequency operation is not too advantageous. Its disadvantage is the greater number of components required in comparison to the phase shift type, hence its greater power consumption and greater chance of failure. For these reasons, it was decided that a phase shift oscillator would prove the best. A basic type is shown in FIG AIII-3 and, as can be seen, some kind of a ladder network of resistances and capacitances is used to produce the phase shift required in the feedback loop. In order to achieve temperature dependence, the resistances are replaced by thermistors and thus the oscillation frequency varies as the resistance of the thermistor changes. Since the output of such a circuit is very sensitive to load changes, a buffer amplifier with a transistor in the common-collector configuration is used to present a low impedance source to the load. Silicon transistors were use, as were low temperature coefficient wire-wound resistors and low temperature coefficient capacitors to reduce the temperature dependence of the circuit outside of the thermistors.

A thermistor was used in the voltage divider circuit supplying the base bias to complete the temperature stabilization of the oscillator. Finally, stabilization of the voltage supply is necessary, which may be done in one of two ways; Keojion (1956) describes a common collector configuration utilizing a capacitance stabilization in the base circuit, but a further method is to use a breakdown diode across the collector to base of an emmiter follower, with a resistance across the base to emmiter to provide breakdown current to the diode and the base bias to the transistor. In the latter case, the value of the resistance is chosen by experimenting. The use of mercury batteries and the circuit above reduces the supply dependence to one hertz for a two-volt drop in the D.C. voltage. The frequency drift over a period of three days in a constant temperature bath was equivalent to 0.02^{o}C, once the results had been corrected for bath drift. Finally, to provide a check on whether any drift was due to thermistor drift or transistor drift, a set of wire-wound resistors was added to the circuit in such a way that they could be interchanged with the thermistors by means of a miniature DPDT relay. This last check was performed infrequently, as it disturbs the stability of the oscillator for a period of fifteen minutes. The final circuit with all of the ramifications is shown in FIG AIII-3. Comparison of temperatures in the U.W.O. hole led to differences of no more than .01^{o}C. Since a 3-lead cable was available, the power supply was kept on the surface for much of the time. However, the batteries can quite easily by placed in the probe, in which case a single-strand cable with a ground return is all that is necessary. Frequencies were recorded on a Racal SA 535 1.2 Mc/s counter-timer. Two 12V car batteries were used to supply the 13 watts power consumption of the instrument. Maximum sensitivity requires an input of at least 70 mv. rms power or ±100 mv excursion about a mean level. Accuracy is 1 count or 1 part in 10^6, whichever is larger. Frequency output is displayed on a 6-digit illuminated display set in vertical scales, which are switched off during counting. Counting periods of 0.1, 1 and 10 seconds are available in the frequency mode and a measuring time of 1, 10, 1000 cycles of the input on the period mode. Further provision exists for control by external clock pulses giving count times of 1 msec to 10^4 seconds. Resetting of display is accomplished either manually or automatically with further provision for using external pulses.

In field use the oscillator is switched on in the probe while the rest of the equipment is being set up. Initially the standards are switched in and the rate of drift checked for 10 minutes, then the transistors are switched in and the probe lowered in the borehole. At each chosen depth the display is noted when several consecutive readings agree. When the bottom of the borehole is reached, the standards are switched back in and the drift checked. In the event of serious drift, corrections may be made to the thermistor readings from a drift curve. Freequencies were converted to temperatures by means of a calibration curve for the probe.

AIII-3 PROBES

AIII-3-1 Probe Description and Design

The design of non-leaky probes proved initially to be something of a hit-and-miss operation. Probably the major defects were due to over-sophistication and lack of sufficient surface preparation practice.

To describe each of the probe designs attempted would take many pages, so only the final designs are discussed here. These are one each for the 3-lead series compensated, relay shunt compensation and the oscillator probe. A very simple design was selected for the first system, which consists of a 12 cm length of brass tube with a 0.5 cm bore. An eyelet is drilled out at the bottom and a threaded cap drilled to cable size at the top. The thermistor is merely soldered to the cable and epoxied into the brass tube. Several of these probes have been used down to 1200 m and have had fairly continuous use over a 3-year period. A weight is hung beneath the probe, since the probe is very light. It was found by testing the the U.W.O. borehole that a 1 cm diameter weighting probe slung 12 cm beneath the thermistor probe had no noticeable effect on the recorded temperatures. Since a brass weight of several pounds would have to be unduly long, the weight was constructed of a copper cylinder filled with lead. Copper cylinders up to 30 cm in length were used so that holes filled with heavy oil could be tackled.

The relay probe had to be more sophisticated, since the relays are too expensive to epoxy into expendable probes and need better pressure protection to operate reliably. Partial protection against moisture movement down the cable wires is provided by a 6-inch brass tube filled with epoxy. In addition to this, a turnbuckle arrangement at the top enables a R.T.V. rubber cylinder to be put under pressure and clamped tightly against the cable as the turnbuckle is screwed down. The bottom thermistor protection was machined out to reduce the time constant of the probe and it was thermally isolated from the main body by a 1 cm fibreglass piece. Again a turnbuckle arrangement was used to avoid twisting the wire during probe assembly. The relay sits in the well at the top of the thermistor housing. A little oil is placed in the housing to improve the time constant. All screw threads are protected by O-ring seals.

The 15 cm epoxy protection had to be dispensed with in the oscillator probe and the probe increased in diameter to 3 cm to accommodate the increased number of components. This probe was 45 cm long with the component and thermistor housing made of aluminium and the end pieces of stainless steel. An O-ring bearing on the cable and a 2.5 cm epoxy seal were the only protections against leakage. The rest of the design was similar to the relay probe.

AIII-3-2 Probe Production

Certain very important factors must be remembered during the manufacture of probes if leaks are to be avoided. Any oxide on machined parts must be removed with a fine emery cloth. Wire, joints, probes and housing must all be thoroughly cleaned with acetone and methyl ethyl ketone (the latter seems to be very good at removing grease). All traces of flux must be removed from leads (the acetone does this well). During lead soldering, a heat sink should be used to avoid thermal shock on the thermistor.

AIII-4 CABLE PREPARATION

Cables are colour-coded with Armaco Plastic tape at 30 m (100 ft) intervals whilst being wound from the wooden drum onto a Sharpe drum. The 100 ft points were laid out with a steel tape. To find the position for the colour marker, the cable was stretched with a 10-15 lb pull.

After use in several deep boreholes, the light-weight cable was checked for permanent stretch. This amounted to less than 1% in the top 100 ft and occurred during the first use of the cable, indicating a value in stretching cables in a deep hole before marking, particularly before their use in attempting very accurate lithological work.

AIII-5 CONTINUOUS LOGGING

Using very narrow probe elements and fast-response thermistors, it is easy to build a system with a time constant of 10-20 seconds. This suggests that a fairly accurate continuous logging system might be constructed. Accuracies of .01^{o}C are readily achievable at speeds of 6m per minute in low gradient sections such as the upper 30 m of the U.W.O. borehole. Use of a 3-lead jack mounted in a brass cylinder and a plug held in a small clamp gave a noise level of 1 ohm, or .002^{o}C. The system was utimately abandoned mainly because of the difficulty of maintaining constant lowering speed by hand and the resulting depth uncertainty between 100 ft markers. However, the system could very readily be adapted to a motorized winch with a magnetic paint cable-marking system and an inductive pickup to the marker channel on the recorder. With faster response probes, this obviously would provide close to an optimum amount of data on each borehole.

APPENDIX IV

THE DIVIDED BAR:

CONSTRUCTION AND CALIBRATION

AIV-1 DETAILS ON BAR CONSTRUCTION & CALIBRATION

The first bar constructed consisted of two brass cylinders through which the circulated water was passed. The end of one of these was machined flat and polished to 1/30th thou. At the other end the stack consisted of a 3.2 mm. fused silica disc and a 1/8" machined brass disc, glued to the cylinder with silver epoxy. Thermocouples of 26-gauge copper-constantin were fixed in grooves in the discs initially, but this was considered to be too close to a disturbed interface; radial holes were then drilled through to the centre ¼" from each face. Thermocouples were calibrated against a platinum resistance standard and it was determined that using the e.m.f. voltage developed across the discs as opposed to the actual temperature resulted in maximum errors of about 0.5% around room temperature. The thermocouple e.m.f.'s across the fused silica disc and across the unknown were initially measured on a Cambridge Vernier Potentiometer. However, for later measurements the signals were amplified by two D.C. amplifiers and the output fed into a ratiometer. The system was calibrated using a set of crystalline quartz X-cut discs and fused silica discs; the amplifier gain was adjustable on one instrument so that the ratio could be set to unity with a similar fused silica disc in place to the one fixed into the bar. The system worked well except for the fact that the rock contact resistance were high. The method assumes contact resistances to be the same for the rocks and the fused quartz. It was decided that more axial pressure than the original 20 lbs on the bar was necessary. The system was redesigned to fit into a hydraulic jack and the stack made symmetrical with a quartz disc at top and bottom; thus two ratios were found across each of the standards and the unknown. The latter addition, as well as compensating heat losses better, was found to be very informative, as bad positioning of the unknown led to very disimilar ratios. Each of the stacks is cemented to the reservoir with heat-sink compound. This arrangement has a fairly high axial strength but parts easily under shear. In this way it is easy to change the stack for one of a different diameter. FIG 4-1 on page of the text shows the general bar construction. Increasing axial pressures greater than 400 psi made little difference to the results. Since the use of the ratiometer seemed to be successful, it was boxed up with two Fluke Electronic Galvanometers and with various switches for checking gains, zeros etc, by F. Anglin. The amplifiers used have a disadvantage in that they have a very low input impedance, this requiring close matching of thermocouple impedances. The reservoirs were held at constant temperatures by circulating water through them from constant temperature baths set at approximately 20 and 30^{o}C. Initially two Townsend amd Mercer baths were used but replaced by Colora type NB baths. The latter when cycling correctly varied by approximately .03^{o}C with a 15 sec. cycle.

AIV-2 EFFECT OF ENVIRONMENT

Room temperature in the vicinity of the bar usually remained constant to better than 1^{o}C during a day's measurements and with the reservoir temperatures straddling the mean room temperature the repeatability of measurements was usually better than 1%. Room temperature varied by no more than 4^{o}C during the year.

A few measurements were made on the effect of the mid-point temperature between the two reservoirs not being equal to the ambient temperature. FIG AIV-1 shows the apparent change in ratio for a) a 0.25 inch thick disc of fused silica, b) the change in the summed voltages down the bar reflecting the change in apparent conductivity of the bar and c) the change in slope and intercept on a set of fused silica discs. As is apparent the effect is linear over a range of 1.5°C but the variation of ratio is of the order of 10% per°C. Thus in order to maintain an accuracy of 2% on the bar the mid-point bar temperature must be within 0.25°C of the ambient room temperatures. This is easily done by checking the ambient temperature with a thermometer or thermistor in the vicinity of the sample in the divided-bar, and adjusting the upper and lower reservoir temperatures to keep the mid-point of the bar at the ambient temperature.

Conductivity variation with axial pressure is less than 1% variation above 200 lb/sq. in., but up to 5% deviations at very low pressures. These variations were the major reason for abandoning the initial divided-bar arrangement with axial pressures of only 20 lbs; all the discs measured on this bar were repeated on the new bar.

Increasing pressure showed increased conductivity, indicating that the contact resistance was being decreased effectively. Between axial pressures of 400 to 1500 lb/sq. in, no major trends could be seen, particularly none of the reversals noted by Birch (1954). The U.W.O. bar could probably detect conductivity differences of .3% with careful use.

AIV-3 CALIBRATION OF BAR

Calibration standards consisted of a set of X-cut quartz discs of thicknesses 0.4, 0.6, 0.8 and 1.2 cm for the 7/8" and the 1 3/8" diameter bars, and sets of fused silica discs of $^{1}16$, $^{1}8$, $^{1}4$" thicknesses for the 1", 1 1/8", 1 1/4" bars. The large diversity of sizes of bar was found to be necessary to accomodate the various sizes of diamond-drill core. Larger cores were, if possible, drilled out to the 1 3/8" size; this was the most frequently used bar. Since the U.W.O. core averages 1 1/8" diameter a bar was made of this size.

Procedures for calibration were as follows. Each calibration disc was placed in the bar and an axial pressure of 600 psi placed on it. A glycerine and water mix was used to improve the bar contact with the specimen. After a period of ten minutes the ratio was recorded every two minutes until three consecutive readings were the same. This was repeated several times for each disc until three similar sets of results could be obtained. Next an average thickness for each of the calibration discs is calculated from micrometer measurements between the two faces. A plot can then be made of voltage ratios against the thickness of the sample. Using the absolute value of the conductivities of the two types of quartz given by Ratcliff(1959, 1963) a graph may be constructed of ratio against thermal resistance. Thus if the ratio and thickness of any other sample are known, its conductivity may be calculated. In practice since the plot of ratio versus thermal resistance was generally linear, a least squares line through the points was calculated and this was used to determine the unknown conducitivities.

Although the plot gave the same slope for the two types of silica, the intercepts were offset with the X-cuts having much lower contact resistances than the fused. This is discussed again in section A-IV-6.

A-IV-4 CONDUCTIVITY STANDARDS

As discussed above the divided-bar is only a relative method and absolute standards must be used as calibrations. Provided the same standards are used by all workers so that interlaboratory measurements are comparable, the absolute conductivities of the standards to 1 or 2% are not too important. For this reason the standards used for this work were those of optical quality fused silica and crystalline quartz with the discs cut perpendicularly to the optic axis. The literature contains a large number of determinations for those standards. Results for fused silica show a great deal of dispersion of values even in the most recent ones. Recent comprehensive summaries have been given by Ratcliffe (1963) and Devyatkova et al (1960) which show an agreement to within 2% of the mean curve given by Ratcliffe (1959) for temperatures between -150°C and +50°C in which:

$$k = 10^{-7}\ (31600 + 46t - 0.16t^2)$$

where k is the thermal conductivity in cal/cm.sec.°C and t is the temperature in °C. Several authors have obtained results substantially off the above curve (eg. Kanamori et al 1968). Whether the uncertainty is due to quality variations is still largely unanswered. Although determinations for crystalline quartz show less variation there are inconsistancies of several per cent. The mean curve given by Ratcliffe (1959) for a temperature range of 0°C to 100°C fits the relationship:

$$k = (60.7 - 0.242t)^{-1}$$

These materials have been adopted as standards for this work using Ratcliffe's curves. The range of thermal conductivities of common rocks is bracketed by these standards. If calibration curves are linear for both of these substances, it hopefully is safe to assume a similar relationship for rocks with intermediate conductivities. Calibration curves for the U.W.O. bars are discussed in section A-IV-5.

A table below gives values for thermal conductivity of the standards at the temperatures encountered:

Temperature (^{o}C)	Conductivity (mcal/cm.sec.^{o}C)	
	Fused Silica	Quartz normal to c-axis
10	3.20	15.8
20	3.25	15.3
30	3.28	14.7

TABLE AIV-4-1 THERMAL CONDUCTIVITY OF CRYSTALLINE QUARTZ AND FUSED SILICA

As more reliable determinations of absolute conductivity become available corrections can be made as necessary.

A-IV-5 CALIBRATION RESULTS

Calibration methods were as described earlier. Frequent recalibrations of the various bars were carried out. The most frequently used bar, the 1⅜" diameter one, was recalibrated on six different occasions over a period of a year and by three different operators. Thus it is a useful indicator of the human element present. In these repeats the ambient temperatures varied from 24.5 to 25.5^{o}C and axial pressures from 300 to 700 psi. Calibration differences in the slope of the least squares fit amount to less than ±1% while those in the contact resistance amounted to 20%. Thus for samples with thermal resistances of 50 which would correspond to a disc of 1 cm thickness with a thermal conductivity of 20 mcl/cm.sec.^{o}C which is probably higher than most values to be encountered, the variations in contact resistance cause approximately the same error as the variations in slope. Thus the divided-bar is probably accurate to better than ±2% for a range of thermal resistances of 50 to 400.

A-IV-6 THERMAL CONTACT RESISTANCE

During the calibration of the bar it was observed that the thermal contact resistance of fused quartz was always greater than that of the crystalline, even when the surfaces of both were ground to the same finish.

Usually the thermal resistance of a specimen in the divided-bar is considered as

$$T.R. = \frac{a}{k_c} + \frac{t}{k_s}$$

where a is the summed thickness of the contacts, t is that of the unknown and k_c and k_s are the conductivities of the contact and the unknown respectively.

The simplest way to consider the contact is as shown in FIG A-IV-2 where the contact is a fluid film, usually a water and glycerine mixed so that

$$T.R. = \frac{a}{k_w} + \frac{t}{k_s}$$

However as mentioned the contact resistance varied between standards of different conductivities. The magnitude of this variation is insufficient to change sample conductivities by more than a per cent however for precise work it needs to be accounted for. It also does not seem to have been mentioned in the literature.

Reconsider the contact as shown in FIG A-IV-4 with a series of 'parallel ' paths of sample abutting the much flatter metal surface; the above can then be rewritten as

$$T.R. = \frac{a}{k_c} + \frac{t-a}{k_s}$$

where $k_c = \alpha k_s + (1-\alpha)k_w$, and α is the proportion of 'parallel' paths.

Selection of intermediate values for α, which is really dependent on the type of material and the degree of grinding gives variations of the thermal contact resistance of the correct order.

APPENDIX V

PREPARATION OF DISCS

AND MEASUREMENT ON DIVIDED BAR

A-V-1 CUTTING OF DISCS

The core obtained came in a variety of sizes from $1\frac{1}{8}$-inch diameter drill up to a 4-inch rotary drill. Some of it had been split or sliced by the oil companies for geological examination. Where the core was already of bar diameter, it was used as obtained, but the rest was cored to one of the bar sizes, either $1\frac{3}{8}$, $1\frac{1}{4}$, $1\frac{1}{8}$, 1 or $\frac{7}{8}$ inches, depending on the diameter available. If enough core was available, composite discs were made up so the largest diameter discs possible could be realized. Coring was with thin walled diamond corers, made by Habit, mounted in a Canadian Buffalo drill press with water at 40 lb/sq inch as cutting fluid. Cores of salt or with salt infilling were drilled out using a light oil as cutting fluid. The thin walled corers produce a core of more uniform diameter and of smoother finish than the thicker walled types although their lifetime is shorter. For coring the raw core was either mounted in plaster of paris blocks or held in lead weighted G-clamps, again depending on core size. Where possible, core lengths of 4 inches or more were drilled out, thus allowing extra core in case it should split during discing. Larger cores were trimmed off to $2\frac{1}{2}$ inches, washed and allowed to dry out thoroughly. Once dry, a piece of 'Scotch' tape was pressed on along the length of the core, followed by a wedge of Lepages epoxy smeared on top of the tape. Very soft shales were completely encased in epoxy before the wedge was added. Complete encasement in epoxy was a very successful means of ensuring that cores did not split during discing and the subsequent soaking. An epoxy wedge ensured that, at the end of the cut, the break was in the wedge and the rock was cut through smoothly. Generally twenty cores were prepared at one time and then left to set for a full day before discing. Discing was performed with a modified Linley drill press, the main spindle being geared up to rotate at 4600 rpm, and fitted with two Golden Rimlock diamond impregnated saw blades spaced .85 cm apart, producing discs of .825 cm thickness. Later the spacing of the saw blades was increased to give discs with a thickness of .95 cm. In initial experiments the core, held vertically in a chuck, was rotated at various speeds between 2 and 20 rpm as suggested by Misener and Beck (1960). However, with the vertical setup one side of the disc broke off with a pip in its centre which required an additional amount of careful grinding. In the final system the core was held steady in the chuck and the carriage automatically fed the core into the saw at 0.15 inches/min, driven by a geared-down Bodine speed reducer motor with a speed of 7.2 rpm. Hand feeding was found to be too irregular, producing scores on the rock surface.

The cutting fluid used was Esso Cutwell EP 65, diluted 1 part of oil to 4 of water. The carriage was stopped by a microswitch, which is usually set so that the saw cuts right through the rock and half way through the epoxy wedge. Once the saw has come to a halt, the drive mechanism on the carriage is disconnected and the carriage redrawn manually. This prevents small scratches on the disc surface from the slowly rotating saw blade. The core is then removed from the chuck, the disc broken carefully away from the epoxy to avoid breaking the disc edges, and then washed in warm water to remove oil. The cutting fluid does not have time in 8 minutes to penetrate far into the rock, so is easily removed by washing. Any slight roughness on the disc is removed by grinding with a mixture of a fine corundum powder (#305) and water on a glass plate.

The effect of various surface finishes on the thermal conductivity of three disc sets was investigated to determine which might be most suitable for taking out small imperfections left after sawing. At axial pressures of 1000 lb/sq inch, a 160 micron surface (80 grade corundum powder) gave conductivities low by about 3%. However, any grade better than a 40 micron surface (240 grade) gave differences of less than .5%. To give a smooth finish with few imperfections and without exessive grinding being necessary, the #305 grinding powder was selected. Results for one set of discs are shown in diagram (FIG A-V-1), for which the particle sizes were found in A.B. Metal Digest, Vol. II, 1965. Discs were washed and dried and the thickness estimated on a dial gauge reading to .0005 cm. In general, the discs showed a wedge shape of .004 cm with the Ordinary Rimlocks which, with the thicker Golden Rimlock blades, was reduced to .002 cm. In both cases the surface finish was flat to better than .001 cm, the latter blades actually being about twice as good. The recommended finish for discs is given as .003 cm by Misener & Beck (1960), and these discs were well within this range. Disc diameters were measured to .01 cm by means of Vernier calipers.

A-V-2 SOAKING OF DISCS

Discs prepared from more permeable materials showed a tendency to dry out in the divided bar. At first, this was combatted by coating such discs with a thin layer of quick drying varnish around the curved surface. Later it was found that a strip of Saran-wrap achieved better results with little variation in divided bar ratios being observed over periods of several hours. Before soaking the discs were weighed dry and then one of three methods employed to soak them. Initially, discs were merely left to soak in distilled water with a little Photoflo added as a wetting agent. Some of these specimens took up to a week to attain the final weight. In this time a fairly large amount of leaching of salts from Salina specimens occurred, which was evidenced by soaked weights lower than the dry weights and salty-tasting water. In order to minimize the soaking time, a vacuum soaking system was built. A large brass cylinder was constructed with two inlets through a brass stopper with an 'O'-ring seal. One inlet led to a rotary pump via a vacuum gauge and the other to a container of distilled water and wetting agent. The discs were stacked in the cylinder with porous discs separating each of the rock discs. The clip on the distilled water inlet was closed off and the rotary pump turned on. After a period the pump clip was closed, the distilled water end opened and the discs soaked for a further period of time. FIG A-V-2 shows the effect of various times of vacuuming followed by 45 minutes of soaking. Longer periods of soaking had little further effect. The optimum period of vacuuming can be seen to be about 3 hours if one is also to take care of tight impermeable sections, followed by 45 minutes soaking. This method was generally used where samples might leach considerably or the material was very shaley and might swell and split on the addition of water. Other specimens were placed in a beaker of 'wetted' distilled water and sat in vacuum at room temperature overnight. About 6-7 hours soaking under these conditions had the same effect as the vacuum system. All samples containing salt were soaked in a saturated brine solution.

A-V-3 <u>MEASUREMENT OF THERMAL CONDUCTIVITY</u>

After a sufficient soaking period, the samples were removed from the water, the excess wiped off the surface and the disc weighed to determine the wet weight. Before placing in the divided bar, a few drops of the (Bullard) mixture, 1 part glycerine to 4 parts water, were placed on the surfaces. The sample was carefully positioned in the bar described in Section IV-2-4 and the press cranked up to give an axial pressure of 1000 psi as recommended by Diment (1962) and Birch (1950). Ratios were noted every two minutes until three consecutive readings were alike. However, if the ratios across top and bottom were wildly different, the sample was removed and re-prepared after wiping the surface to remove any grit. Widely different ratios were found to be good indicators of one of four things: a) grit on surfaces, b) drying contact at one surface, c) incomplete soaking and hence probably water migration and d) bad grinding of one surface. Occasionally a sample was re-weighed after measurement on the bar; the difference was always less than 2% of the total water content. Finally the discs were dried out thoroughly at 50°C and re-weighed.

Diameters were measured on the soaked discs since many of them were not exactly the same size as the divided bar, eg., the core diameters varied between 1$\frac{1}{8}$ and 1$\frac{3}{16}$ inches along the U.W.O. core. Thus a correction was necessary to correct for these size differences. For an undersize disc the correction is merely the ratio of the areas. However that for an oversize one is slightly more complex. Jaeger & Beck (1954) have suggested a suitable correction and tested it. Their solution may be further simplified as suggested by Doig (1961). However since the computer was available for analysis the more rigorous solution was used.

Now the thermal resistance of the unknown is given by

$$T.R. = A + B.Q$$

where A and B are the calibration constants for that particular divided bar. In this case single discs were being used so that A is given a value believed to be typical of the types of rocks measured as discussed in Chapter IV-3, and B is the effective average thermal resistance of the sandwiches. Q is the ratio of the e.m.f. across the unknown to the average across the samdwiched standards above and below it. The thermal conductivity of the unknown is given by the ratio of its thickness to its thermal resistance.

Dry density is given by

$$\rho = \frac{4 \times (\text{Dry Weight})}{\pi \times D^2 \times t}$$

where D and t are the diameter and thickness of the disc respectively.

Disc porosity is given by

$$\sigma = \frac{4 \times (\text{Wet Weight - Dry Weight})}{\pi \times D^2 \times t}$$

where the solution soaking the discs is water with a density of 1.0.

The density, porosity and thermal conductivity were obtained from a programme written for the IBM 7040 and the results plotted against depth for individual holes.

A-V-4 ESTIMATION OF COMPOSITION

The disc composition was estimated with the aid of a microscope. Limestones and dolostones were distinguished using 10% hydrochloric acid. In cases where it was of interest to know the composition with greater certainty, the individual discs were stained with alizarin red S, which provides a useful method of distinguishing between limestone and dolomite. Calcite becomes stained deep red, ferrous dolomite a purple colour and dolomite is unstained. In the latter case an image contrasting microscope, the Quantimet, could be used to estimate composition percentages.

MATRIX I Inorganic Low Melting-Point Compounds

A search of the literature led to the following list of compounds which might have a low melting point and a thermal conductivity similar to that of rocks:

	M.P. °C	B.P. °C	
Aluminium nitrate $Al(NO_3)_3.9H_2O$	73.5	134	(decomposes)
Chromium nitrate $Cr_2(NO_3)_6.18H_2O$	60	100	(decomposes)
Ferric nitrate $Fe(NO_3)_3.9H_2O$	47.2	125	(decomposes)
Manganese nitrate $Mn(NO_3)_2.6H_2O$	26	129	(decomposes)
Nickel nitrate $Ni(NO_3)_2.6H_2O$	56.7	136.7	(decomposes)
Chrom alum $KCr(SO_4)_2 12H_2O$	89	100	(loses $10H_2O$)
Zinc nitrate $Zn(NO_3)_2.6H_2O$	36.4	103-130	(loses $6H_2O$)

TABLE AVI-1 COMPOSITION AND BEHAVIOUR OF LOW MELTING POINT SALTS

Aluminium nitrate was easily available and so tests were carried out using it. The salt proved very difficult to handle as when heated too rapidly, it decomposed producing nitrogen peroxide. Heated more slowly it melted into a slush at 80°C and then became fairly liquid although the liquid contained large amounts of gas. De-gassing is difficult even under vacuum. On solidifying in the mould the solid phase expands and without care stresses the mould so much that removal is very difficult. The salt was made into cylinders which had fairly firm edges but were very porous. Addition of water caused the substance to break up so that discs must be cut off on the saw-blade without coolant. The substance is also deliquescent so that discs become quite damp after a few days in the labor-atory. Conductivities measured on discs of aluminium nitrate showed var-iation of twenty per cent, the variation seemingly being due to size of crystals, dampness of specimen and gas content. In view of this variation

APPENDIX VI

THERMAL CONDUCTIVITY OF CHIPS

these substances were considered unsuitable as a matrix.

MATRIX II Use of Plaster of Paris and Epoxies

For the application of simple theories it is useful to have matrix conductivities similar to those of rock. Possible choices along these lines are plaster of paris and an epoxy with a high thermal conductivity such as Stycast. (Emerson & Cumming Stycast 2850). In either of these cases it is not possible to measure exactly the proportions of each material. However, by knowing the total volume of the disc and the densities of both chips and matrix, the proportions can be arrived at as follows:

$$\alpha = \frac{V_m}{V_t} = \frac{\rho_T - \rho_e}{\rho_m - \rho_e} \quad \ldots\ldots\ldots\ldots \quad i).$$

where V_m is the chip volume, V_t is total disc colume, ρ_e is density of matrix, ρ_m that of the chips and ρ_T the density of composite disc. V_m/V_t is the volume ratio of chips.

The use of plaster was not very successful since the thermal conductivity of the material varied by 5 to 10% between batches. Much of this variation was probably due to varying moisture content. However the Stycast epoxy conductivity varied by 2% at the most between batches and was thus considered a suitable matrix material. Several pieces of rock on which the conductivity had already been measured were broken up and mixed into an epoxy matrix in a long cylinder. Once the material had cured discs were prepared and measured as described in Appendix V.

As with most materials composed to two or more phases, uncertainties arise as to how to treat the results to obtain the thermal conductivity of the chips. Below is a table showing several conductivities calculated by various models:

	Disc Conductivity	Disc Density	Conductivity Simple	Of Chips Models	Calculated by	
Composition	k_{total}	ρ	k_{series}	$k_{parallel}$	k_{geom}	k_{true}
epoxy	2.7	2.23	-	-	-	2.7
epoxy - shale	3.1	2.30	4.9	4.1	4.4	4.4
epoxy - sandstone	5.5	2.20	9.9	6.7	7.5	7.0

TABLE AVI-2 THERMAL CONDUCTIVITIES OF EPOXY MATRIXES

where k

$$k_{series} = \frac{k_{epoxy} \cdot k_{total}}{k_{epoxy} - (1-\alpha)\, k_{total}} \qquad \text{ii)}$$

$$k_{parallel} = \frac{k_{total} - (1-\alpha)\, k_{epoxy}}{\alpha} \qquad \text{iii)}$$

$$k_{geom} = \frac{k_{total}}{k_{epoxy}^{1-\alpha}}^{1/\alpha} \qquad \text{iv)}$$

α is given by i) and k_{series}, $k_{parallel}$ and k_{geom} are the matrix conductivities calculated by the respective models and k_{true} is the thermal conductivity of a disc of the material comprising the matrix.

The methods of arriving at the chip thermal conductivity are dependent on the configuration of the chips in the matrix but it is apparent that reasonable accuracies can be obtained by this method. Once the chip conductivity is determined it still remains, however, to try and calculate an 'in-situ' value allowing for porosity, water saturation and salt infilling of the rock.

MATRIX III Water Matrix

As mentioned in the first section water is not an ideal matrix since it has a low thermal conductivity. However, it is a simple one to use either with the chips mixed in a small cell for use in the divided bar or in a larger container and using a needle-probe. Horai & Simmons (1969) have used the method with a needle-probe to determine the thermal conductivity of the chips from the measured bulk value they use a mathematical solution suggested by Hashin & Shtrikman (1962). However very high temperature gradients can result close to the needle probe leading to water movement problems and so a system using small cells in the divided bar is to be preferred. Again a similar mathematical solution could be used but the geometric model of Woodside and Mesmer (1961) is just as accurate, if the chip conductivity is less than 10x conductivity of water.

$$k_T = k_W^{\sigma} \cdot k_g^{1-\sigma} \qquad \text{v)}$$

where k_w, & k_g are the thermal conductivity of the water and the chips respectively, σ is the effective cell porosity.

The results can be reduced to a simple plot of $\ln(k_T/k_w)$ against $\ln(k_g/k_w)$ which is a series of straight lines of slope $(1-\sigma)$. Since k_T and σ can easily be measured, the chip conductivity k_g is determined. This latter method is being used regularly at the Dominion Observatory for oil-wells in which no core is available. However it cannot be expected to give the accuracy of heat flow measurement achieved in cored drill-holes. As well as the problem of contamination of the chips with those from other horizons, a variety of well-logs must also be used to reconstruct the 'in-situ' rock. The latter points apply whatever the matrix.

APPENDIX VII

LIST OF BOREHOLES USED FOR

TEMPERATURE AND THERMAL CONDUCTIVITY MEASUREMENTS

Boreholes used for temperature determinations:-

	Co.	Twn.	Lot	Conc.	Elev. ft
Geol. Surv. Canada #1	PCE	ATL	27	1	256
Geol. Surv. Canada #2	RSL	RSL	24	2	251
M&M - C. Lyons #1	PEL	CGC	14	6W	819
Birchfield Oil & Gas #1 - C. Telfer #1	WLG	PLC	30	7	1031
Consumers Gas #14	WCD	BRT	6	15NR	587
Consumers Gas #466	WLD	BRT	32	BF	574
Consumers Gas #648	WLD	CRD	11	5	611
Consumers Gas #1158	LCL	LTH	4	3	280
Consumers Gas #1511	HLT	EQG	17	5	936
J. Brown	NFK	SWG	5	B	605
Dom.Obs.-U.W.O. #4	MDX	LND	16	3	815
Bluewater IOE. Dunwich 7-17-1	ELG	DNC	17	1	701
Riddell-Hanna-Ewing #4A	KNT	CTM	11	13	582
Imperial East Becker #2	LMB	SBR	20	6	599
Imperial Bickford #16	LMB	SBR	5	11	608
Imperial Kimball #5	LMB	MRE	15	4	626
Dow Farm #13	LMB	SRN	DOW REF		615
Dow Farm #14	LMB	SRN	DOW REF		615
IHS Colchester S #1	ESX	CCS	79	1	604
O.K. West - Gosfield N.#2	ESX	GFN	13	9	627
Imp. Calvan Malden #75	ESX	MLD	25	3	610
Imp. Calvan Malden #25	ESX	MLD	25	3	591
Northville #106	WSH	SLM	1-15-7E		984
Muttonville - C. Fawver #2	MCB	LNX	13-4N-14E		668
Overisel #150	ALL	OVR	7-4N-14W		661
Overisel #157	ALL	OVR	13-4N-14W		718
Overisel #162	ALL	OVR	28-4N-14W		667
Billingsly #1	NWG	GDW	8-14N-11W		1072
Austin-Marek #1	MEC	AUS	23-14N-9W		1028

E Bregg #2	OSC	RCH	5-17N-10W	1093
Marion #192	MSK	RCL	24-21N-8W	1252
Marion #829	MSK	CLM	23-21N-6W	1146
Marion #965	OSC	MAR	28-20N-7W	1165
Brazos-Sun State Foster #1	OGE	RSE	28-24N-2E	1457

Boreholes with complet . cores used for conductivity determinations:-

	Co.	Twn.	Lot	Conc.	Elev. ft.
Geol. Surv. Canada #1	PCE	ATL	27	1	256
Geol. Surv. Canada #2	RSL	RSL	24	2	251
U. S. Steel	NFK	CTL	21	1	704
Dom. Obs. - U.W.O. #4	MDX	LDN	16	3	815
Midrim #3A	MDX	ADD	9	2NER	762
Midrim #4	MDX	ADD	10	3NER	755
Imp. MacGillivray	MDX	MGL	5	19	711
Imp. Plympton	LMB	PLM	1	3	643
Imp. Enniskillen	LMB	ENN	20	5	664
Argor #1	LMB	MRE	28	2	598

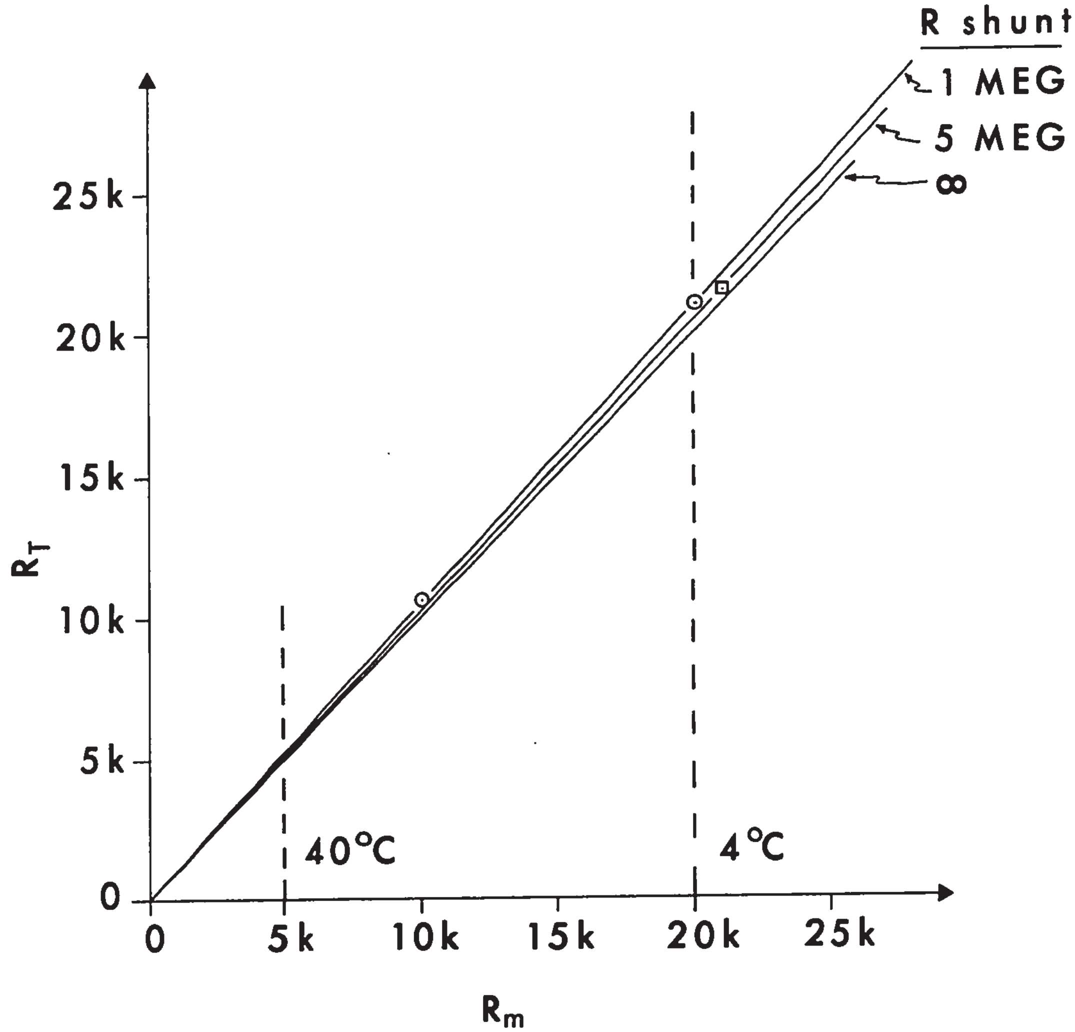


FIG.A3-1

EFFECT OF FIXED RESISTANCE ON BRIDGE READINGS

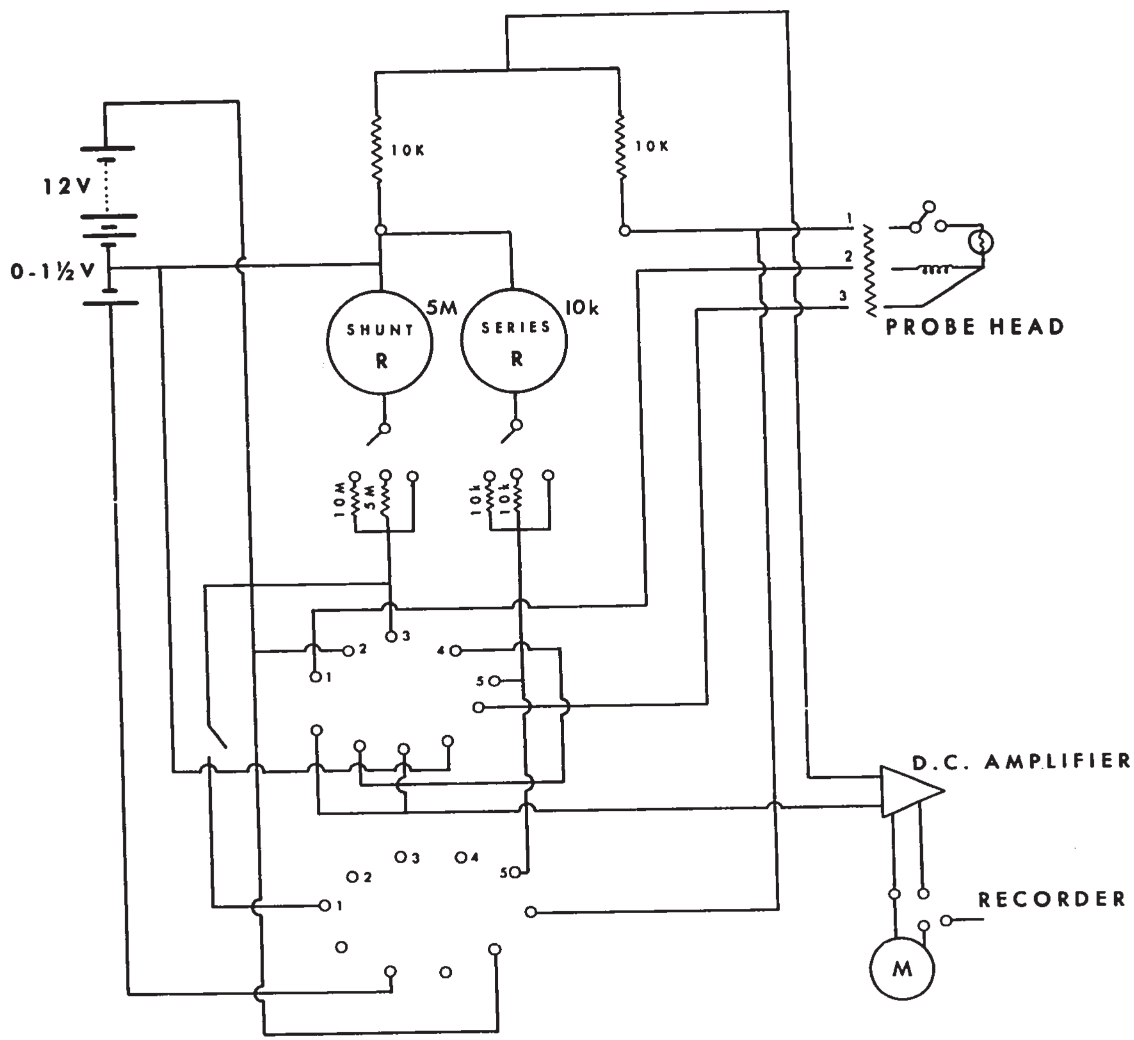


FIG. A3-2 MODIFIED WHEATSTONE BRIDGE DESIGN

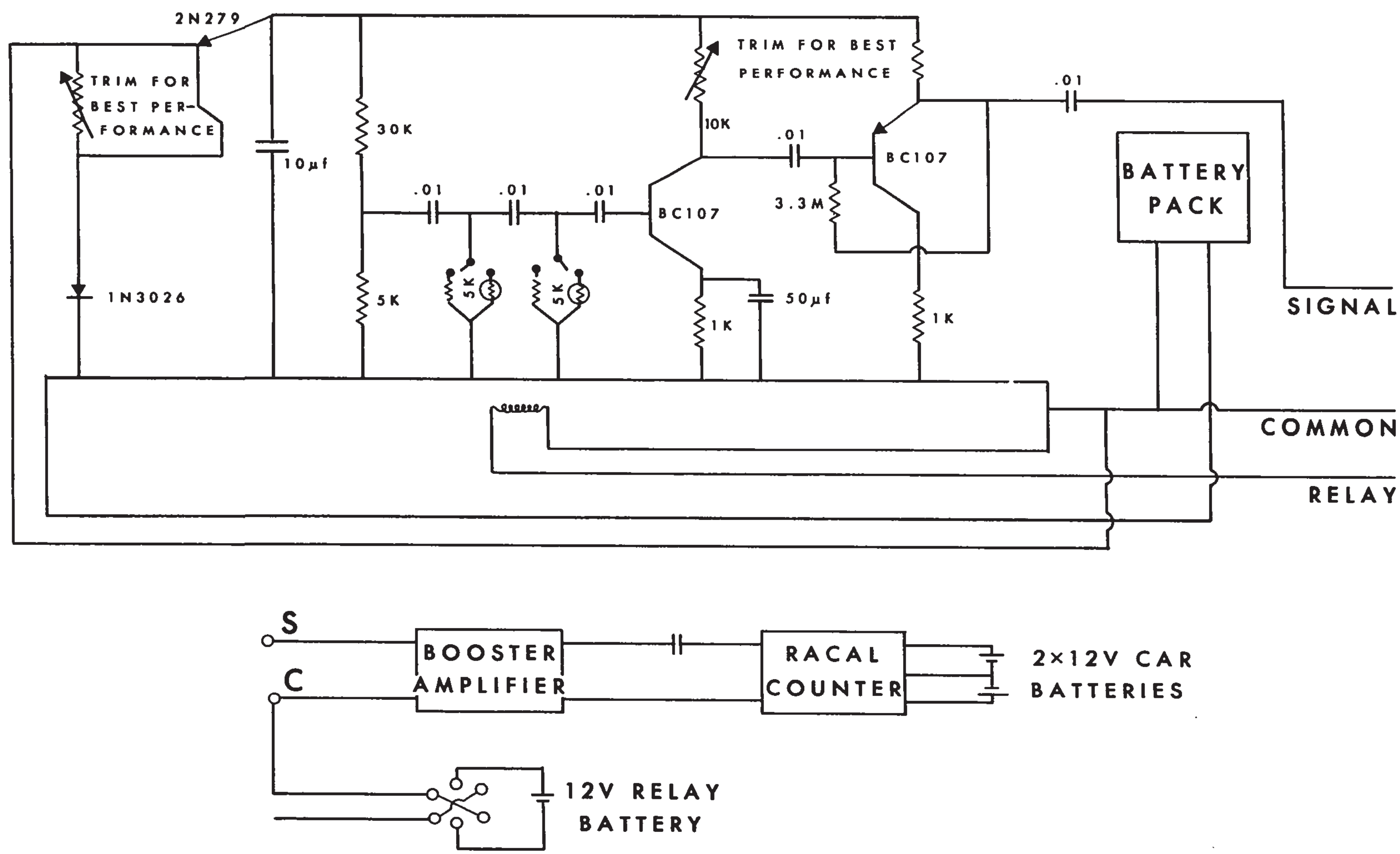


FIG. A3-3 OSCILLATOR PROBE DESIGN

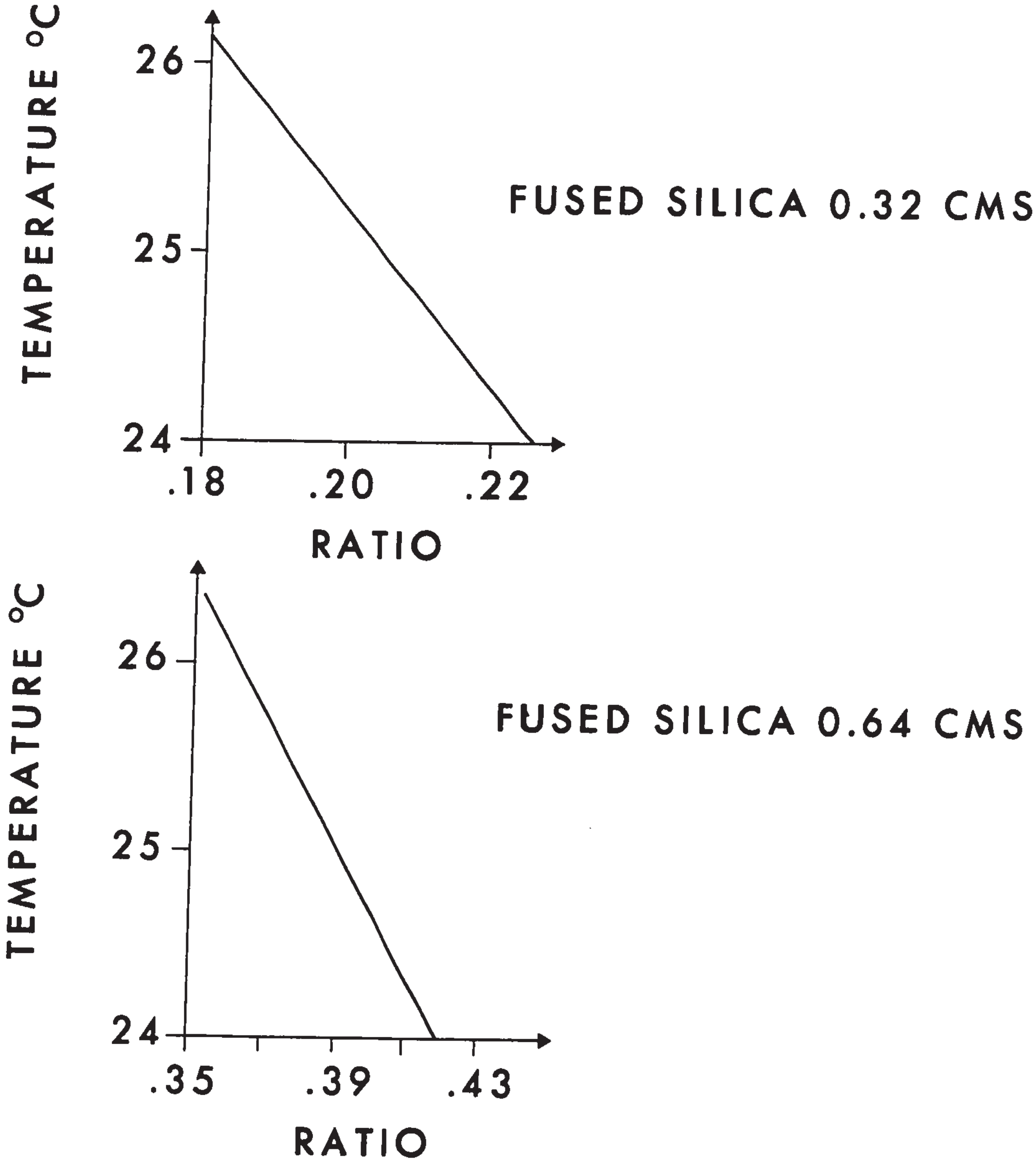


FIG. A4-1
EFFECT OF AMBIENT TEMPERATURE
ON DIVIDED-BAR RESULTS

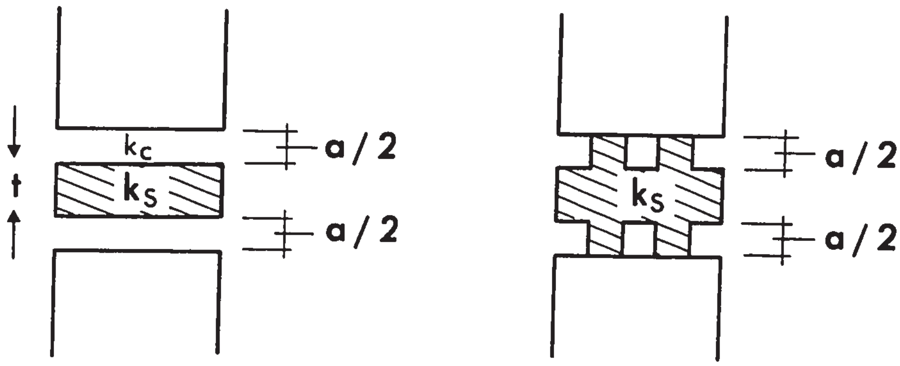


FIG. A-4-2
MODELS OF THERMAL CONTACT RESISTANCE

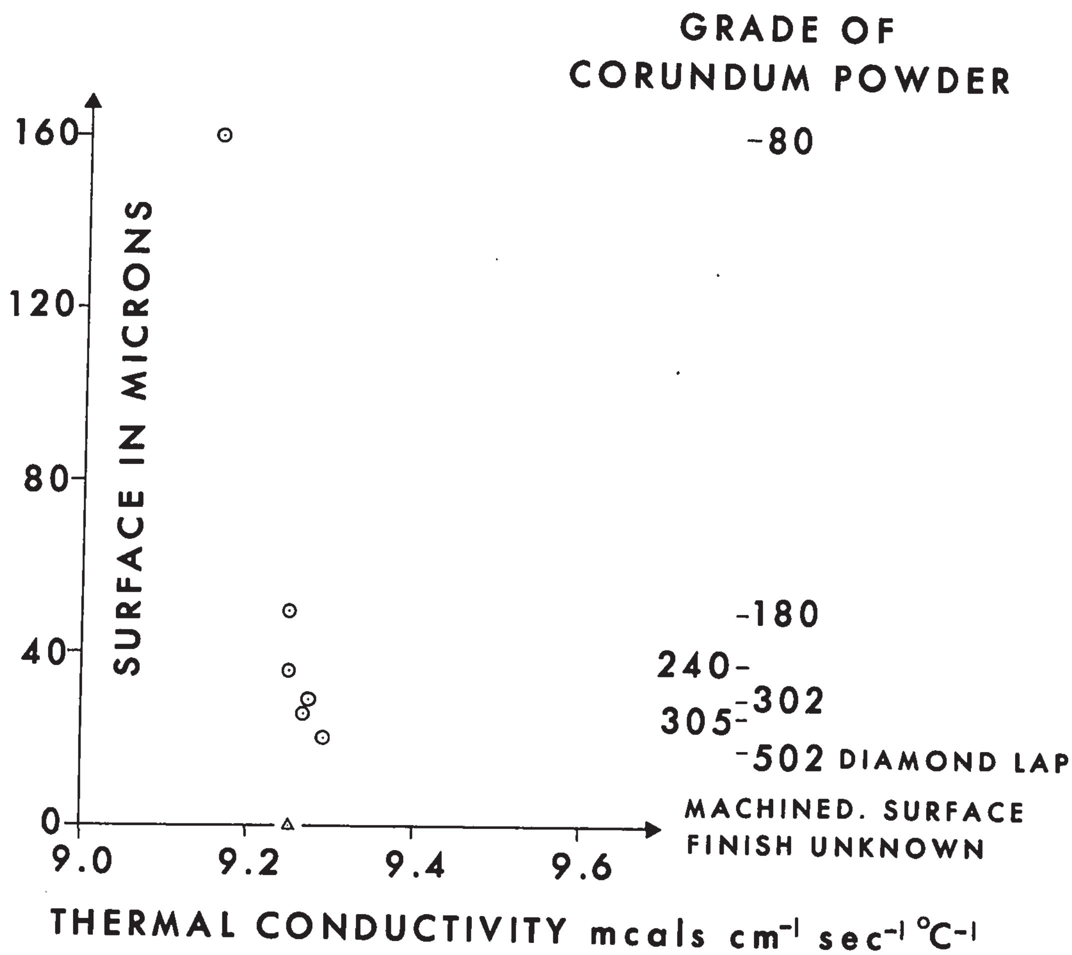


FIG. A-5-1
EFFECT OF SURFACE FINISH ON THERMAL CONDUCTIVITY

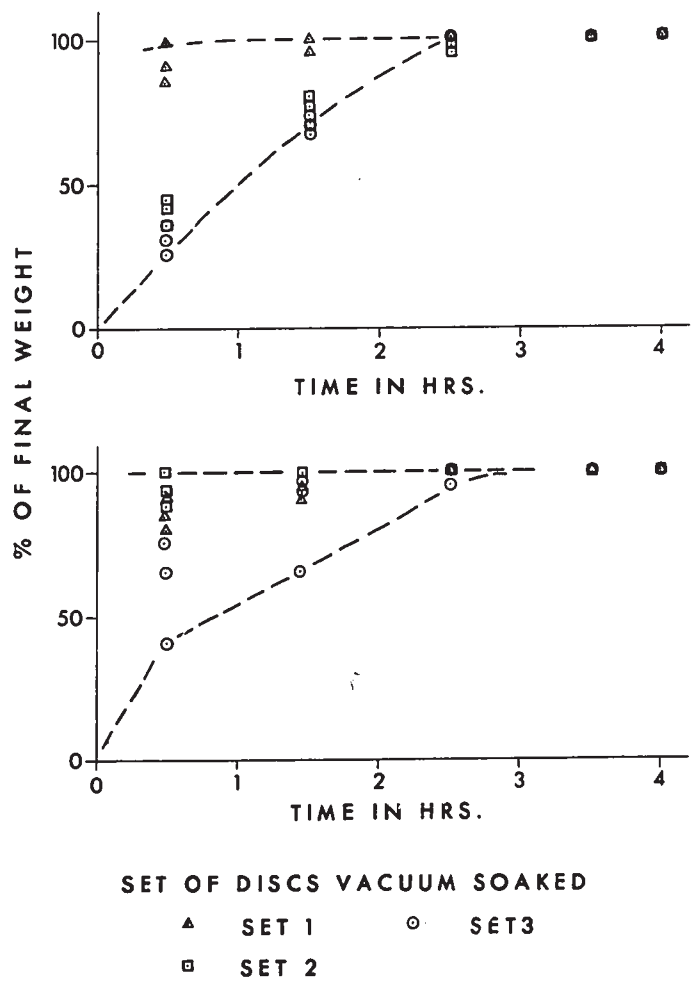


FIG. A-5-2 TIME TAKEN FOR ROCKS TO REACH SATURATION WITH VACUUM SOAKING :- LIMESTONES & DOLOMITES

BIBLIOGRAPHY

BIBLIOGRAPHY

Ahlman, H.W., 1948. Glacialogical research on the North Atlantic coasts, Royal Geogr. Soc. Res. Series #1.

Alberta Oil and Gas Conservation Board, 1967. Presure-depth and temperature-depth relationships in Alberta crude oil pools, Oil and Gas Consv. Brd., 67-22.

Anglin, F.M., 1964. Temperature measurements in western Canada, M.Sc. thesis, University of Western Ontario.

Anglin, F.M., and A.E. Beck, 1965. Regional heat flow pattern in western Canada, Can. J. Earth Sci., 2, 176-182.

Aston, D., 1969. Average soil temperature based on period 1960 to 1968, Canada, Dept. of Transport, Meteorological Branch, CDS #5-69, pp. 1-8.

Bacon, L.O., 1966. Geologic structure east and south of Keweenaw Fault on basis of geophysical evidence, Earth Beneath the Continents, A.G.U. Mono. 42-55.

Ball, M.W., T.J. Wenver, H.D. Crider, and D.S. Ball, 1941. Shoestring Gas Fields of Michigan, Amer. Assoc. Petrol Geol. Spec. Publ. "Stratigraphic Type Oil Fields", Levorsen (ed.), pp. 237-266.

Beck, A.E., 1956. Stability of thermistors, Jour Sci. Instrum, 33, 16.

Beck, A.E., 1957. Steady state method for rapid measurement of thermal conductivity of rocks, J. Sci. Instrum, 34, pp. 186-189.

Beck, A.E., 1963. Light-weight borehole temperature measuring equipment for resistence thermistors, J. Sci. Instrum., 40, pp. 452-454.

Beck, A.E., 1965. Measuring heat flow on land, In: Terrestrial Heat Flow, W.H.K. Lee (ed.), Amer. Geophys. Union Monograph 8.

Beck, A.E., 1970. Non-equivalence of oceanic and continental heat flow and other geothermal problems, Comments on Earth Sciences: Geophysics, pp. 29-35.

Beck, A.E., F.M. Anglin, and J.H. Sass, 1971. Analysis of heat flow data--'in-situ' thermal conductivity measurements, Can J. Earth Sci., 8, pp. 1-19.

Beck, A.E., and J.M. Beck, 1958. On the measurement of thermal conductivities of rocks by observations on the divided-bar apparatus, Trans. A.G.U., 30, pp. 1111-1123.

Beck, A.E., and A.S. Judge, 1969. Analysis of heat flow data - 1. Detailed observations in a single borehole, Geophys. J.R. astr. Soc., 18, pp. 145-158.

Benfield, A.E., 1939. Terrestrial heat flow in Great Britain, Proc. Roy. Soc., London, A 173, pp. 428-450.

Bergthorson, P., 1962. In: Proc. Conf. Climates of 11th and 16th centuries, Aspen, Colo., Natl. Center Atmos. Res. Tech. Notes 63.

Best, E.W., 1953. Pre-Hamiltonian Devonian stratigraphy of southwestern Ontario, Ph.D. thesis, University of Wisconsin.

Birch, F., 1947. Temperature and heat flow in a well near Colorado Springs, Am. J. Sci., 245, pp. 733-753.

Birch, F., 1948. The effect of Pleistocene climatic variations upon geothermal gradients, Am. J. Sci., 246, pp. 729-760.

Birch, F., 1954. The present state of geothermal investigations, Geophysics, 19, pp. 645-659.

Birch, F., 1954. Thermal conductivity, climatic variation and heat flow near Calumet, Michigan, Am. J. Sci., 252, pp. 1-25.

Birch, F., and H. Clark, 1940. The thermal conductivity of rocks and its dependence upon temperature and composition, pts. 1 and 2, Am. J. Sci., 238, pp. 529-558 and pp. 613-635.

Birch, F., and H. Clark, 1945. Estimate of surface heat flow in west Texas Permian Basin, Am. J. Sci., 243A, pp. 69-75.

Boldizar, T., 1958. Temperature drop of incompressible fluids rising in a borehole, Acta Techn. Acad. Scient. Hung., 19, pp. 371-378.

Boldizar, T., 1958. Geothermic investigations in the Hungarian plains, Acta Geol., 5, pp. 245-254.

Boldizar, T., 1964. Heat flow in the Hungarian Basin, Nature, 202, pp. 1278-1280.

Bolton, T.E., 1957. Silurian stratigraphy and Palaeontology of the Niagara Escarpment, Geol. Surv. Can, Memoir 289.

Bosson, G., F. Gutman, and L.M. Simmons, 1950. A relationship between resistance and temperature for thermistors, J. Appl. Phys., 21, pp. 1267-1268.

Bradbury, J.C. and E. Atherton, 1965. Precambrian basement of Illinois, Illinois State Geol. Sur. Circ. #382, p. 12.

Bradley, E.A., and T.J. Bennett, 1965. Earthquake history of Ohio, Bull. S.S.A., 55, pp. 745-752.

Brant, A.A., 1943. Gravimetric and magnetic geophysical surveys in the gas fields of southwestern Ontario, Ont. Dept. Mines Rept., 52, pp. 73-88.

Brigham, R.J., and C.G. Winder, 1966. Structural geol. of Palaeozoics in southwest Ontario, Ont. Petrol. Instit. Ann. Meeting.

British Association for the Advancement of Science, 1882, Report 74.

Bryson, R.A., and W.M. Wendland, 1967. Tentative climatic patterns for some late glacial and post glacial episodes in central North America, In: Life, Land and Water, U. of Manitoba Press, pp. 271-298.

Budyko, M.I., and K.Y. Kondratiev, 1964. The heat balance of the earth, In: Research in Geophysics, 2, Solid Earth and Interface Phenomena, Odishaw (ed.), M.I.T. Press.

Bullard, E.C., 1939. Heat flow in South Africa, Proc. Roy. Soc. London, A, 173, pp. 474-502.

Bullard, E.C., 1954. Flow of heat through the floor of the Atlantic Ocean. Proc. Roy. Soc. London A, 222, pp. 408-429.

Bullard E.C., and E.R. Niblett, 1951. Terrestrial heat flow in England, M.N.R.A.S. Geophy. Suppl., 6, pp. 222-258.

Caley, J.F., 1943. Palaeozoic geology of the London area, Geol. Surv. Can., Memoir 237.

Caley, J.F., 1940. Palaeozoic geology of Toronto-Hamilton area, Geol. Surv. Can., Memoir 224.

Callender, G.S., 1961. Temperature fluctuations and trends over the earth, Quart. J. Roy. Meteorol. Soc., 87, pp. 1-12.

Campenella, A.J., 1962. A telemetering thermometer, Marine Sci.,Instrum. 1.

Carslaw, H.S. and J.C. Jaeger, 1959. Conduction of heat in solids, Oxford Press.

Chapman, L.F., and D.F. Putman, 1966. Physiography of southern Ontario, U. of Toronto Press, pp. 13-51.

Clark, S.P., 1966. Handbook of physical constants, Geol. Soc. Amer. Memoir #97, pp. 459-483.

Clark, S.P., 1957. Absorption spectra of some silicates in the visible and near infra-red, Am. Minera., 42, pp. 732-742.

Clark, S.P. and E.R. Niblett, 1956. Terrestrial heat flow in the Swiss Alps, M.N.R.A.S. Geophys. Suppl. 7, pp. 176-195.

Clark, S.P., and A.E. Ringwood, 1964. Density and constitution of mantle, Rev. of Geophysics 2, pp. 35-88.

Claude, A., 1952. Anomalies at Pechelbronn, Acad. Sci. Comptes. Rend. 234 (21), pp. 2097-2098.

Clément and Peclet, 1841, Ann. de chimie et de physique 3e ser. 2., p. 107.

Cohee, G.V., 1948. Cambrian and Ordovician rocks in the Michigan Basin and in adjoining areas, Amer. Assoc. Petrol. Geol. Bull., 32, pp 1417-1448.

Cohee, G.V., 1965. Geologic history of the Michigan Basin, J. Wash. Acad. Sci., pp. 211-222.

Cohee, G.V., 1947. Lithology and thickness of traverse group in Michigan basin, U.S.G.S. Oil and Gas Inven. Prel. Chart #28.

Cohen, T.J. and R.P. Meyer, 1966. Mid-continent gravity high, In: Earth Beneath the Continents, A.G.U. Mono., pp. 141-165.

Combs, J., and G. Simmons, 1969. Heat flow measurements in Iowa (abs.), Trans. Am. Geophys. Union, 50, p. 316.

Coster, H.P., 1947. Terrestrial heat flow in Persia, Monthly Notices, Roy. Astr. Soc. Geophys. Suppl. 5, pp. 131-145.

Crain, I.K., 1967. The influence of post-Wisconsin climatic changes on thermal gradients in the St. Lawrence lowlands., M.Sc. thesis, McGill University.

Crain, I.K., 1968. Glacial effect and significance of continental terrestrial heat flow measurements, Earth and Planetary Sci. Letters, 4, pp. 69-72.

Dansgaard, W., and S.J. Johnson, 1969. One thousand centuries of climatic record from Camp Century on the Greenland ice sheet, Science, 166, pp. 377-381.

Darton, N.H., 1920. Geothermal data of the United States, United States Geol. Surv. Bull. 701.

Dept. Energy, Mines and Resources, 1965. Aeromagnetic Series, Geophysics Paper 4578.

Devyatkova, E.D., A.V. Petrov, I.A. Smirnov, and B.Y. Moizhes, 1960. Fused quartz as a model material in thermal conductivity measurements, Fiz. Tverdogo Tela 2, p. 738.

Diment, W.H., 1967. Thermal regime of a large diameter borehole: instability of the water-column and comparison of air and water-filled conditions, Geophysics 32, pp. 720-726.

Diment, W.H., and E.C. Robertson, 1963. Temperature, thermal conductivity and heat flow in a drilled hole near Oak Ridge, Tennessee, Jour. Geophys. Res., 68, pp. 5035-5047.

Diment, W.H. and J.D. Weaver, 1964. Subsurface Temperatures and Heat Flow in the AMSOC Core Hole near Mayaguez, Puerto Rico, in A Study of Serpentenite, US -NAS-NRC Publ. 1188.

Doig, R., 1961. A further Study of Terrestrial Heat Flow in the St. Lawrence Lowlands of Quebec, M.Sc. Thesis, McGill U.

Doig, R., V.A.Saull and R.A. Butler, 1961. A New Borehole Thermometer, J.G.R. 66, 4263-4264.

Donaldson, I.G., 1961. Free Convection in a Vertical Tube with a Linear Wall Temperature Gradient, Austral. J. Physics 14, 529-539.

Dreimanis, A., 1964. The Pleistocene Time Scale in Canada, in Geochronology of Canada, U. of Toronto Press, 139-156.

Eade, K.E., W.F. Fahrig and J.A. Maxwell, 1966. Composition of Crystalline Shield Rocks and Fractionating Effects of Regional Metamorphism, Nature 211, 1245-1249.

Ells, G.D., 1967. Michigan's Silurian Oil and Gas Pools, Mich. Geol. Sur. Report of Investigation #2.

Elson, J.A., 1960. Littoral Mollusks of the Champlain Sea, Exc. Guide Dept. Sci., McGill U.

Emiliani, C., 1961. Cenozoic Climatic Changes as Indicated by Stratigraphy and Chronology of Deep-Sea Cores of Globigerina-Ooze Facies, Ann. N.Y. Acad. Sci. 95, 521-536.

Ericson D.B. and G. Wollin, 1968. Pleistocene Climates and Chronology in Deep-Sea Sediments, Science 162, 1227-1244.

Ericson, D.B., M. Ewing, G. Wollin and B.C. Heezen, 1961. Atlantic Deep-Sea Sediment Cores, Geol. Soc. Amer. Bull. 72, 193-286.

Evans, C.S., 1950. Underground Hunting of the Silurian in Southwestern Ontario, Proceed. Geol. Assoc. Can. 3, 55-85.

Fahrig, W.F., K.E. Eade and J.A.S. Adams, 1967. Abundance of Radioactive Elements in Crystalline Shield Rocks, Nature 214, 1002-1003.

#

Garland, G.D. and D.H. Lennox, 1962. Heat Flow in Western Canada, Geophys. J. Roy. Astron. Soc. 6, 245-262.

Geil, F.G. and J.H. Thompson, 1962. In-Situ Temperature Sensor - a Mechanical Resonator, Marine Sci. Instrum. 2.

Gough, D.I., 1963. Heat Flow in the Southern Karroo, Proc. Roy. Soc. London A 272, 207-230.

Gow, A.J., H.T. Ueda and D.E. Garfield, 1968. Antarctic Ice Sheet: Preliminary Results of First Core Hole to Bedrock, Science 161. 1011-1013.

Gretener, P.E., 1967. On the Thermal Instability of Large Diameter Wells - an Observational Report, Geophysics 32, 727-738.

Gretener, P.E., 1968. Temperature Anomalies in Wells due to Cementing of Casing, J. Petrol. Technol. Feb., 147-150.

Grieve, R.O., 1955. Leaching of Silurian Salt-Beds in Southwestern Ontario as evidenced in Holes drilled for Oil and Gas, C.I.M.M. Trans, 63, 10-16.

Guelke, R., J.C.R. Heydenrych and F. Anderson, 1949. Measurement of Radioactivity & Temperature in Narrow Boreholes and Development of Instruments for Purpose, Jour. Sci. Instrum. 26, 150.

Guyod, H., 1944. Temperature Well Logging, Oil Weekly - compendium of papers.

Hammond, D.L., C.A. Adams and P. Schmidt, 1964. Linear Quartz Crystal Temperature Sensing Element, 19th Annual I.S.A. Conf., p. 11.

Hamza, V.M. and R.K. Verma, 1969. The Relationship of Heat Flow with Age of Basement Rocks, Bull. Volcanol. 33, pp. 123-152.

Hancock, J., 1967. Wyllie Field, Ont, Petrol. Instit. Ann. Meet.

Hansen, B.L. and C.F. Langway, 1966. Deep Core Drilling in Ice and Core Analysis at Camp Century, Greenland, 1961-1966, Antarctic J. 1, 207-208.

Hashin, Z. and S. Shtrikman, 1962. Variational Approach to the Theory of Effective Magnetic Permeability of Multiphase Materials, J. Appl. Phys. 33, 3125.

Heiland, C.A., 1946. Geophysical Exploration, Prentice-Hall Chapt. XI.

Heiskanen, W.A. and U.A. Uotila, 1956. Gravity Survey of State of Ohio, Ohio Geol. Surv. Rept. Inv. #30, 34.

Herrin, E. and S.P. Clark, 1956. Heat Flow in West Texas and Eastern New Mexico, Geophysics 21, 1087-1099.

Hewitt, D.F., 1962. Salt in Ontario, Ont. Dept. Mines Indust. Min. Rept. #6.

Hinze, W.J., 1963. Regional Gravity and Magnetic Anomaly Maps of Southern Peninsula of Michigan, Mich. Geol. Surv. Rept. Inv. 1. 26.

Hobson, G.D., 1960. Seismic Refraction and Reflection Surveys in Southwest Ontario, Can. Min. J. 81, 83-88.

Hodgson, J.H., 1953. A Seismic Survey on the Canadian Shield, Dom. Obs. Publ. 16.

Hohlt, W., 1948. Nature and Origin of Limestone Porosity, Colorado Sch. of Mines Rept. 43.

Holmes, A., 1915. Radioactivity in Earth and Earth's Thermal History, Geol. Mag. 2, 60-71, 102-112.

Horai, K., 1969. Effect of Past Climatic Changes on the Thermal Field of the Earth, Earth and Planetary Science Letters 6, 39-42.

Horai, K. and G. Simmons, 1969. Thermal Conductivity of Rock-Forming Minerals, Earth and Planetary Science Letters 6, 359-368.

Hough, J.L., 1958. Geology of the Great Lakes, U. of Illinois Press.

Hough, J.L., 1963. Prehistoric Great Lakes of North America, Am. Sci. 51, 84-110.

Howard, L.E. and J.H. Sass, 1964. Terrestrial Heat Flow in Australia, J.G.R. 69, 1617-1626.

Hyndman, R.D., I.B. Lambert, K.S. Heier, J.C. Jaeger and A.E. Ringwood, 1968. Heat Flow and Surface Radioactivity Measurements in the Precambrian Shield of Western Australia, Phys. Earth Planet. Interiors 1, 129-135.

Ingerson, E., 1962. Problems of the Geochemistry of Sedimentary Carbonate Rocks, Geochemica et Cosmochemica Acta 26, 815-847.

Ingham, W.N. and N.B. Keevil, 1951. Radioactivity of the Bourlamaque, Elzevir and Cheddar Batholiths, Canada, Bull. Geol. Soc. Amer. 62, 131-147.

Jacobs, J.A. and D.W. Allan, 1954. Temperatures and Heat Flow within the Earth, Trans. Roy. Soc. Can. 48, 33-39.

Jaeger, J.C., 1956. Conduction of Heat in an Infinite Region Bounded Internally by a Circular Cylinder of a Perfect Conductor, Austral. J. Phy. 9, 167-179.

Jaeger, J.C., 1956. Numerical Values for Temperatures in Radial Heat Flow, J. Math. Physics 34, 316-321.

Jaeger, J.C., 1965. Applied Theory of Heat Conduction, pp. 7-24 in Terrestrial Heat Flow, A.G.U. Geophysical Monograph #8.

Jaeger, J.C. and A.E. Beck, 1955. Calculation of Heat Flow through Discs and its Application to Conductivity Measurements, Brit. J. Appl. Phys. 6, 15-16.

Jaeger, J.C. and J.H. Sass, 1964. A Line Source Method for Measuring of Thermal Conductivity and Diffusivity of Cylindrical Specimens of Rock and other Poor Conductors, Brit. J. Appl. Phys. 5, 1-8.

Jessop, A.M., 1964. Geothermal Studies in Canada, C.I.M.M. Bull. Feb., 1-4.

Jessop, A.M., 1965. A Lead-Compensated Thermistor Probe, J. Sci. Instrum. 41, 503.

Jessop, A.M., 1968. Three Heat Flow Measurements in Canada, Can. J. Earth Sciences, 5, 61-68

Jessop, A.M. and A.S. Judge, 1971. Five Heat Flow Measurements in Southern Canada, Can. Jour Earth. Sciences, in press.

Jodry, R.L., 1957. Reflections of Possible Deep Structures by Traverse Group Facies Changes in Western Michigan, A.A.P.G. Bull. 41, 2677-2694.

Jodry, R.L., 1969. Growth and Dolomitisation of Silurian Reefs, St. Clair Co., Michigan, Amer. Assoc. Petrol. Geol. Bull. 53, pp. 957-981.

Joyner, W.B., 1960. Heat Flow in Pennsylvania and West Virginia, Geophysics 25, 1229-1241.

Judge, A.S. and A.E. Beck, 1967. An Anomalous Heat Flow Layer at London, Ontario, Earth and Planetary Science Letters 3, 167-170.

Kanamori, H., N. Fujii and H. Mizutani, 1968. Thermal Diffusivity Measurements of Rock-Forming Minerals, J.G.R. 73, 595-605.

Katz, S., 1955. Seismic Study of Crustal Structure in Pennsylvania and New York, Bull. S.S.A. 45, 303-325.

Kelly, W.A., 1936. Pennsylvanian System in Michigan, Mich. Geol. Survey Publication #40.

Kelvin, Lord, 1864. Secular Cooling of the Earth, Trans. Roy. Soc. Edin. 23, 157.

Kidoo, G., 1962. Silurian Reefs of St. Claire and Macomb Counties, Michigan, Ont. Petrol. Instit. Ann. Meeting.

Koepke, W.E. and B.V. Sanford, 1965. Silurian Oil and Gas Fields of Southwestern Ontario, G.S.C. Paper, 65-30.

Kraskowski, S.A., 1961. On the Thermal Field of Old Shields, Izv. Akad. Nauk. S.S.R. Ser. Geofiz. 1961, pp. 247-250.

Krige, L.J., 1939. Borehole Temperatures in the Transvaal and Orange Free State, Proc. Roy. Soc. London A 173, pp. 450-474.

Lachenbruch, A.H., 1968. Preliminary Geothermal Model of the Sierra Nevada, J.G.R. 73, p. 6977.

Lachenbruch, A.H., 1957. Three-Dimensional Heat Conduction in Permafrost beneath Heated Buildings, United States Geol. Surv. Bull. 1052-B, 51-69.

Lachenbruch, A.H., 1957. Thermal Effects of the Ocean on Permafrost, Geol. Soc. Amer. Bull. 68, 1515-1530.

Lachenbruch, A.H., 1968. Rapid Estimation of the Topographic Disturbance to Superficial Thermal Gradients, Rev. of Geophys. 6, 365-400.

Lachenbruch, A.H. and M.C. Brewer, 1959. Dissipation of Temperature Effect in Drilling a Well in Arctic Alaska, United States Geol. Surv. Bull. 1083-C, 73-109.

Lamb, H.H., 1965. The Early Medieval Warm Epoch and its Sequel, Palaeog., Palaeoclim. & Palaeoecol. 1, 13-37.

Lamb, H.H. and A. Woodroffe, 1970. Atmospheric Circulation during the Last Ice Age, Quaternary Res. 1, 29-58.

Lambert, I.B. and K.S. Heier, 1967. Vertical Distribution of Uranium, Thorium and Potassium in the Continental Crust, Geochem. et Cosmo. Acta 31, 377-390.

Landes, K.K., 1945. Salina and Bass Island Rocks of the Michigan Basin, United States Geol. Surv. Prel. Map. Oil and Gas Invest. #40.

Landes, K.K., 1948. Structure of Typical American Oil Fields, A.A.P.G. Bull. 3, 299-304.

Landes, K.K., 1951. Detroit River in the Michigan Basin, United States Geol. Surv. Circular #133.

Landsberg, K., 1853. Pogg. Ann. B 139,1.

Lee, W.H.K. and S. Uyeda, 1965. Review of Heat Flow Data, Terrestrial Heat Flow, Geophy. Mono. 8, A.G.U., 87-190.

Lees, C.H., 1892. On the Thermal Conductivity of Crystal and Other Bad Conductors, Phil. Trans. Roy. Soc. London A 183, 481-509.

Leney, G.W., 1956. Preliminary Investigation of Rock Conductivity and Terrestrial Heat Flow in Southeastern Michigan, M.Sc. Thesis, U. of Michigan.

Lidiak, E.G., R.F. Marvin, H.H. Thomas and M.N. Bass, 1966. Geochronology of Mid-Continent Region, U.S., J.G.R. 71, 5427-5438.

Lister, C., 1963. Geothermal Gradient Measurement using a Deep Sea Corer, Geophys. J. 7, 571-583.

Lliboutry, L., 1966. Bottom Temperatures and Basal Low-Velocity Layer in an Ice-Sheet, J.G.R. 71, 2535-2543.

Lodge, O.J. 1878. Proc. Phys. Soc. London, 1, 201.

Longley, R.W., 1953. Temperature Trends in Canada, Roy. Meteorol. Soc. Proceed. Toronto Conf., 207-211.

Lovering, T.S. and J.G. Goode, 1963. Measuring Geothermal Gradients in Drill-Holes less than 60 feet Deep, East Tintic District, Utah, United States Geol. Surv. Bull. 1172.

MacDonald, G.J.F., 1959. Calculations on the Thermal History of the Earth, J.G.R. 64, 1967-2000.

Manley, G., 1961. A Preliminary Note on early Meteorological Observations in the London Region with Estimates of Monthly Mean Temperatures 1680-1706, Meteorol. Mag. 90, 303-310.

Marshall, G.S. and R.H. Henderson, 1963. Recording Temperatures in Deep Boreholes, Engineering, London 196, 540.

Maxwell, J.C., 1892. A Treatise on Electricity and Magnetism, Vol.1, p. 440, Clarendon Press, Oxford, 2 Vols.

McCormick, G.R., 1961. Petrology of Precambrian Rocks of Ohio, Ohio Geol. Survey Report, Inv. #41.

McGregor, D.J., 1954. Stratigraphic Analysis of Upper Devonian and Mississippian Rocks in Michigan Basin, Amer. Assoc. Petrol. Geol. Bull. 38, pp. 2324-2356.

McLauglin, D., 1954. Suggested Extension of the Grenville Orogenic Belt and Front, Science 120, 287-289.

Mereu, R.F. and J.A. Hunter, 1969. Crustal and Upper Mantle Structure under the Canadian Shield from Project Early Rise Data, Bull. S.S.A. 59, 147-165.

Michigan Dept. of Conservation, 1964. Stratigraphic Succession in Michigan, Chart 1, Michigan Geological Survey.

Misener, A.D. and A.E. Beck, 1960. Measurement of Heat Flow over Land, in Methods and Techniques of Geophysics, 10-61, Academic Press.

Misener, A.D., L.G.D. Thompson and R.J. Uffen, 1951. Terrestrial Heat Flow in Ontario and Quebec, Trans. A.G.U. 32, 729-738.

Misener, A.D. and L.G.D. Thompson, 1952. Pressure Coefficient of Resistance of Thermistors, Can. Jor. Technol. 30, 89.

Mitchell, J.M., 1961. The Thermal Climate of Cities, in Sympos: air over cities, U.S. Public Health Service Tech. Rept. A62-5, pp. 131-145.

Mustonen, E.D., 1967. Micro-geothermal Survey, Lake Dufault, Quebec, M. Sc. Thesis, University of Western Ontario.

Newstead, G.N. and J.C. Jaeger, 1956. The Determination of Underground Water Movements from Measurements in Drillholes, The Engineer, London 202, pp. 76-78.

Nichols, E.A., 1947. Geothermal Gradients in Mid-Continental and Gulf Coast Oil Fields, Trans, A.I.M.E. 170, 44-50.

Nielson, K.E.C., 1959. Temperature Measurements with Thermistors in Canada, Swedish Cement & Concrete Research Instit. Bull. #34.

Nwachukwu, S.O., A.E. Beck and J.B. Currie, 1965. Magnetic Provinces of Lake Huron and Adjacent Areas and their Geological Significance, Can. J. Earth Sci. 2, 227-236.

Oldham, 1954. Gravity and Magnetic Investigation along Alaska Highway and Southeast Ontario, Ph. D. Thesis, U. of Toronto.

Paterson, W.S.B., 1968. A Temperature Profile through the Meighen Ice Cap, Arctic Canada, I.U.G.G. Berne Ass. - Commission of Snow and Ice Reports and Discussions, pp. 440-449.

Pearson, W.J., 1963. Salt Deposits of Canada, North Ohio Geol. Soc. Symposium on Salt, 197-239.

Penrod, E.B., J.M. Elliot and W.K. Brown, 1960. Soil Temperature Measurements at Lexington, Kentucky, Kentucky Acad. Sci. Trans. V 21, 49-60.

Peter, G. and R.E. Wall, 1961. Total Magnetic Intensity Measurements on Lake Erie, Lamont Geol. Obs. Tech. Rept. #1.

Petterson, H., 1949. Exploring the Bed of the Ocean, Nature 164, 468-470.

Pounder, J.A., 1961. Guelph-Lockport Formation of Southwestern Ontario, Ontario Petrol. Instit. Ann. Meeting.

Powell, J.M., 1965. Annual and Seasonal Temperature and Precipitation Trends in British Columbia since 1890, Meteorol. Br. D.O.T. Circular 4296.

Prendergast, 1952. Gravity Correlations in Southwest Ontario, M.Sc. Thesis, U. of Toronto.

Prest, V.K., 1969. Retreat of Wisconsin and Recent Ice in North America, Geol. Surv. Canada Map 1257.

Ratcliffe, E.H., 1959. Thermal Conductivities of Fused and Crystalline Quartz, Brit. J. Appl. Phys. 10, 22.

Ratcliffe, E.H., 1963. Survey of Most Probable Values for Thermal Conductivities of Glasses, Glass Technology 4, 113-128.

Reilly, G.A. and D.M. Shaw, 1967. An Estimate of the Composition of Part of the Canadian Shield in Northwestern Ontario, C.J.E.S. 4, 725-741.

Riggs, C.H., 1938. Geology of Allegan Co., Mich. Geol. Survey Prog. Report #4, 29.

Robin de Q, 1955. Ice Movement and Temperature Distribution in Glaciers and Ice Sheets, J. Glaciol. 2, 523-543.

Rodgers, G.K. and D.V. Anderson, 1963. Thermal Structure of Lake Ontario, in Proc. Sixth Conf. on Grt. Lakes Res., U. of Mich. Publ. #10, 59-69.

Roliff, W.A., 1949. Salina-Guelph Fields of Southwestern Ontario, Amer. Assoc. Petrol. Geol. Bull. 33, 153-188.

Roller, J.C. and W.H. Jackson, 1966. Seismic-Wave Propagation in Upper Mantle: Lake Superior to Denver, Colorado, Earth Beneath the Continents, A.G.U. Mono., 270-275.

Rostoker, G., 1963. Low Frequency Variations in Earth's Magnetic Field and Relation to Conductivity of Upper Mantle, M.Sc. Thesis, U. of Toronto.

Roy, R.F., 1963. Heat Flow Measurements in the United States, Ph.D. Thesis, Harvard U.

Roy, R.F., D.D. Blackwell and F. Birch, 1968. Heat Generation of Plutonic Rocks and Continental Heat Flow Provinces, Earth and Planetary Science Letters 5, 1-12.

Roy, R.F., E.R. Decker, D.D. Blackwell and F. Birch, 1968. Heat Flow in the United States, J.G.R. 73, 5207-5221.

Rozychi, 1948. Note on Distribution of Geothermal Gradients in Poland, Warzawe Towarzystro Nankowe Spawozdania 47, 115-122.

Rudman, A.J., C.H. Summerson and W.J. Hinze, 1965. Geology of Basement in Midwestern States, Amer. Assoc. Petrol. Geol. 49, 894-904.

Sammel, E.A., 1968. Convective Flow and its Effect on Temperature Logging in Samll Diameter Holes, Geophysics 33, 1004-1012.

Sanford, B.V., 1961. Subsurface Stratigraphy of Ordovician Rocks in Southwest Ontario, Geol. Surv. Can. Paper 60-26.

Sanford, B.V., 1962. Sources and Occurrences of Oil and Gas in Sedimentary Basins of Ontario, Geol. Assoc. Can. Proceed., 59-89.

Sanford, B.V., 1964. Subsurface Stratigraphy of Silurian Rocks in South-Western Ontario, Geol. Surv. Can. 64-2, 14-19.

Sanford, B.V., 1965. Salina Salt-Beds - Southwestern Ontario, Geol. Surv. Can. Paper 65-9.

Sanford, B.V. and W.B. Brady, 1955. Palaeozoic Geology of the Windsor-Sarnia Area, Ontario, Supplement to Memoir 240 Geol. Surv. Can., Memoir 278.

Sanford, B.V. and P.G. Quillian, 1959. Subsurface Stratigraphy of Upper Cambrian Rocks in Southwest Ontario, Geol. Surv. Can. Paper 58-12.

Sass, J.H., 1964. Heat Flow Value from Precambrian Shield of Western Australia, J.G.R. 69, 299-308.

Sass, J.H., P.G. Killeen and E.D. Mustonen, 1968. Heat Flow and Surface Radioactivity in the Quirke Lake Syncline near Elliot Lake, Ontario, Canada, C.J.E.S. 5, 1417-1429.

Sass, J.H., R.J. Munroe and A.H. Lachenbruch, 1968. Measurement of Geothermal Flux through Poorly Consolidated Sediments, Earth and Planetary Science Letters 4, 293-298.

Saull, V.A., T.H. Clark, R.P. Doig and R.B. Butler, 1962. Terrestrial Heat Flow in the St. Lawrence Lowland of Quebec, C.I.M.M. Bull. 65, 63-66.

Schlichter, L.B., 1951. Crustal Structure in Wisconsin Area, O.N.R. 86200.

Schlumberger, M., H.G. Doll and A.A. Perebinossoff, 1937. J. Instit. Petrol. Tech. 23,1.

Schneider, R., 1964. Relation of Temperature Distribution to Ground-Water Movement in Carbonate Rocks of Central Israel, Geol. Soc. Amer. Bull. 75, 209-216.

Schwarz, J.H., 1966. Properties of Thermistors used in Geothermal Investigations, U.S.G.S. Bulletin 1203-B.

Schwarz, J.H. and R. Raspet, 1961. Thermal Shock and its Effect on Thermistor Drift, Nature 190, 875.

Shaw, D.M., 1967. Radioactive Elements in the Continental Crust, Nature 208, 479-480.

Simmons, G., 1965. Continuous Temperature-Logging Equipment, J.G.R. 70, 1349-1352.

Simmons, G. and K. Horai, 1968. Heat Flow Data 2, J.G.R. 73, 6608-6629.

Sobczak, L.W., J.R. Weber and E.F. Roots, 1970. Rock Densities in the Queen Elizabeth Islands, Northwest Territories, Geol. Assoc. Canada Proceed 21, 5-14.

Smith, W.E.T., 1962. Earthquakes of Eastern Canada and Adjacent Areas, 1534-1927, Dom. Obs. Publ. #26.

Smith, W.E.T., 1966. Earthquakes of Eastern Canada and Adjacent Areas, 1928-1959, Dom. Obs. Publ. #32.

Smith, T.J., J.S. Steinhart and L.T. Aldrich, 1966. Crustal Structure under Lake Superior, Earth Beneath the Continents A.G.U. Mono., 181-197.

Stefansson, V., 1943. Greenland, Harrap, London.

Steinhart, J.S., 1964. Lake Superior Seismic Experiment: Shots and Travel Times, J.G.R. 69, 5335-5352.

Steinhart, J.S. and S.R. Hart, 1968. Calibration Curves for Thermistors, Deep Sea Research 15, 497-503.

Steinhart, J.S. and R.P.Meyer, 1961. Explosion Studies of Continental Structure, Carnegie Instit. Wash. Publ. 622, 409.

Steinhart, J.S. and G.P.Woollard. 1961. Seismic Evidence concerning Continental Structure, Carnegie Instit. Wash. Publ. 622.

Stockwell, C.H., 1965. Structural Trends in the Canadian Shield, Amer. Assoc. Petrol. Geol. Bull. 49, 887-904.

Strong, N.W., 1934. Significance of Underground Temperatures, Jour. Instit. Petr. Technol. 20, 63.

Summers, P.W., 1967. An Urban Heat Model: its Role in Air Pollution with Application to Montreal, in Proc. First Canad. Conf. on Micrometeorol. Prt II, Meteorol. Br. D.O.T.

Tavernier, P. and P. Prache, 1952. Effect of Pressure on Resistance of a Thermistor, Jour. Physique et Radium 13, 423-426.

Terasmae, J., 1960. Notes on late Quaternary Climatic Changes in Canada, Ann. N.Y. Acad. Aci. 95, 658-675.

Thomas, M.K., 1955. Climatic Trends along the Atlantic Coast of Canada, Trans. Roy. Soc. Can. 49, 15-21.

Thomas, M.K., 1964. Climatic Trends on the Canadian Prairies, in Symposium on Water and Climate, Saskatoon.

Thompson, L.G.D. and A.H. Miller, 1958. Gravity Measurements in Southern Ontario, Dom. Obs. Publ. 19.

Tilton, G.R. and G.W. Reed, 1963. Radioactive Heat Production in Ecolgite and some Ultramafic Rocks, Earth Science and Meteorites, 31-44, North-Holland.

Trusheim, F., 1960. Mechanism of Salt Migration in Northern Germany, Amer. Assoc. Petrol. Geol. Bull. 44, 1519-1540.

U.S.S.R. Coll., 1957, 1958. Geothermal Gradients in the U.S.S.R., Compass 34 (2), 160-162, Compass 35 (2), 137-138.

Van Everdingen, R.D., 1968. Studies of Formation Waters in Western Canada: Geochemistry and Hydrodynamics, Canad. J. Earth Sciences 5, 523-545.

Van Orstrand, C.E., 1924. Apparatus for Measurement of Temperature in Deep Wells by Means of a Maximum Thermometer, Econ. Geol. 19, 229.

Van Orstrand, C.E., 1934. Application of Geothermics to Geology, Bull. Amer. Assoc. Petrol. Geol. 18, 13-38.

Van Orstrand, C.E., 1935. Normal Geothermal Gradients in the United States, Amer. Assoc. Petrol. Geol. Bull. 19, 78-115.

Van Orstrand, C.E., 1951. Observed Temperatures in the Earth's Crust, Internal Constit. of Earth, ed. B. Gutenberg, 107-149, Dover.

Vening-Meinesz, F.A., 1962. Thermal Convection in Earth's Mantle, Continental Drift, ed. K. Runcorn, Academic Press, 145-176.

Verhoogen, J., 1956. Temperatures within the Earth, Physic & Chemistry of the Earth 1, 17-43.

Walcott, R.I., 1970. Isostatic Response to Loading of the Crust in Canada, Canad. J. Earth Sciences 7, 716-727.

Walsh, J.B. and E.R. Decker, 1966. Effect of Pressure and Saturating Fluid on the Thermal Conductivity of Compact Rock, J.G.R. 71, pp. 3053-3062.

Weatherill, G.W., 1966. in Clark Handbook of Physical Constants, Geol. Soc. Amer. Memoir #97.

Weertman, J., 1966. Effect of a Basal Water Layer on the Dimensions of Ice Sheets, J. of Glaciol. 6, 191-209.

Weiss, O., 1938. Temperature Measurements with an Electro-Resistance Thermometer in a Deep Borehole on the East Rand, Chem., Met. & Min. Soc. South Africa J. 39, 149-166.

Wendon, H.E., 1957. Ionic Diffusion and Properties of Qüartz, Direct Current Resistivity, Am. Mineral 42, 859-888.

Werner, F.D. and A.C. Frazer, 1952. New Methods of Converting Platinum Resistance Values to Degrees Centigrade, Rev. Sci. Instrum. 23, 163-169.

Wexler, H., 1961. Growth and Thermal Structure of the Deep Ice in Byrd Land, Antarctica, J. Glaciol. 3, 1075-1087.

Whitham, K., 1965. Geomagnetic Variation Anomalies in Canada, J. of Geomag. & Geoelec. 17, 481-498.

Winder, C.G., 1961. Lexion of Palaeozoic Names in Southwestern Ontario, U. of Toronto Press.

Woollard, G.P., 1966. Regional Isostatic Relations in the United States, Earth Beneath the Continents A.G.U. Mono., 557-594.

Woodside, W. and J.H. Mesmer, 1961. Thermal Conductivity of Porous Media, 1. Unconsolidated Sands, 2. Consolidated Rocks, J. Appl. Phys. 32, pp. 1688-1706.

Wright, J.A. and G.D. Garland, 1968. In-Situ Measurement of Thermal Conductivity in Presence of Transverse Anisotropy, J.G.R. 73, 5477-5484.

Zietz, I., E.R. King, W. Geddes and E.G. Lidiak, 1966. Crustal Study of Continental Strip from Atlantic Ocean to Rocky Mountains, Geol. Soc. Amer. Bull. 77, 1427-1448.

Fairbridge R.W. Comm. # 107 Intern. Assoc. Seis. I.U.G.G. Toronto 1957

PAPERS PUBLISHED ON THESIS WORK

Judge and Beck, "An anomalous heat flow layer at London, Ontario" Earth and Planetary Science Letters 3, p. 167-170, 1967.

Beck and Judge, "Analysis of heat flow data 1 Detailed observations in a single borehole: Geophys. J. Roy. Astr. Soc. 18, p. 145-158, 1969.

See back of thesis for reprints.

Analysis of Heat Flow Data—I
Detailed Observations in a Single Borehole

A. E. Beck and A. S. Judge

DEPARTMENT OF GEOPHYSICS
UNIVERSITY OF WESTERN ONTARIO
LONDON — CANADA

REPRINTED FROM

Geophys. J. R. astr. Soc. (1969) **18,** 145–158

PUBLISHED BY

BLACKWELL SCIENTIFIC PUBLICATIONS
OXFORD AND EDINBURGH

PRINTED IN ENGLAND BY

C. F. HODGSON & SON LTD., LONDON

CORRECTION SHEET

Due to a ridiculous set of circumstances this paper was published without being proofread. Nearly all the errors are minor or self evident but nevertheless irritating and this sheet contains corrections to the errors so far spotted.

P. 145

It should have been noted that this paper is Canadian contribution No. 190 to the International Upper Mantle Project.

P. 150

equation 10 should read

$$\frac{1}{V}\frac{\partial^2 V}{\partial z^2} = \frac{-8\pi^2}{\lambda} \tan 2\pi\left(\frac{t}{P}-\frac{z}{\lambda}\right) \qquad (10)$$

---

equation 11 should read

$$\frac{t}{P} - \frac{z_1}{\lambda} = \frac{1}{8} \qquad (11)$$

---

equation 13 should read

$$k^2 + k\left(\frac{\partial V}{\partial z}/V\right) + \frac{1}{2}\left(\frac{\partial^2 V}{\partial z^2}/V\right) = 0 \qquad (13)$$

Sixth line from bottom - reference should read

Beck et al (1956)

P. 151 Table 1

Second heading "Step function (equation 2)" refers to second and third columns.

Third heading "Periodic function (equation 7)" refers to fourth and fifth column (P, t, λ)

OVER

P. 154

In sixth column, opposite the 258 to 325 meter range, the H value should be 0.80 (not 0.08).

P. 155

Section 7, Second paragraph, last line "with" should read "which"

P. 158

Reference: Jaeger, J. C., 1965, is in "Terrestrial Heat Flow", ed. W. H. K. Lee, American Geophysical Union Monograph.

Figure 1.

In heading, delete "s" from "contours"

Figure 3.

In heading, "+" should be "•"

Figure 4.

"mcal cm^{-2} s^{-1}" should read "μcal cm^{-2} s^{-1}" in body of figure.

Geophys. J. R. astr. Soc. (1969) **18**, 145–158.

Analysis of Heat Flow Data—I Detailed Observations in a Single Borehole

A. E. Beck and A. S. Judge*

(Received 1969 March 25)†

Summary

Heat flow data from a 600-m deep diamond drilled borehole has been used to estimate how short a section of borehole will give a valid heat flow value, to test for recent and ancient climatic changes, underground waterflows and the variation of terrestrial heat flow with depth. Temperatures were repeatedly measured at 3-m intervals; measurements of thermal conductivity. density and porosity were made on specimens sampled at approximately 4-m intervals along the length of the hole. The mean heat flow for the whole borehole before applying any corrections is 0·76 h.f.u. while after correcting for the Wisconsin glaciation the mean value is 1·17 h.f.u., but in both cases some 30 to 100-m sections of the borehole differ by ± 20 per cent from the mean values. The differences cannot be entirely explained as being due to structure, topography, climatic changes or underground waterflows.

1. Introduction

Although measurements of terrestrial heat flow have been made for many years by numerous workers there are a number of aspects which have not been adequately investigated. For instance, over how large an area can heat flow values determined from observations in a single borehole reflect the mean heat flow value for that area? Do underground waterflows exist at depth and how significant are they? How significant are topographic, climatic and structural effects? Do some igneous and sedimentary formations produce more heat than others because of higher radioactive content or exothermic reactions? How long a section of borehole is required to give a reliable value of heat flow representative of that for the whole borehole? For many years, a group of workers at this university has been investigating some of these problems and this paper is the first of a series dealing with the results of these investigations.

To increase the output of heat flow data over land means, basically, making greater use of holes that have already been drilled or are being drilled for commercial purposes. Unfortunately, such holes are frequently not completely cored so that although they may be available for temperature measurements, reliable conductivity data must be obtained by *in situ* methods (Beck 1965), computed methods (Beck & Beck 1965) or by some other method (Diment & Robertson 1963). However, in nearly all cases certain short sections of these boreholes are cored and it is conceivable that detailed

* Present address: Dominion Observatory, Ottawa, Canada.

† Received in original form 1968 November 11.

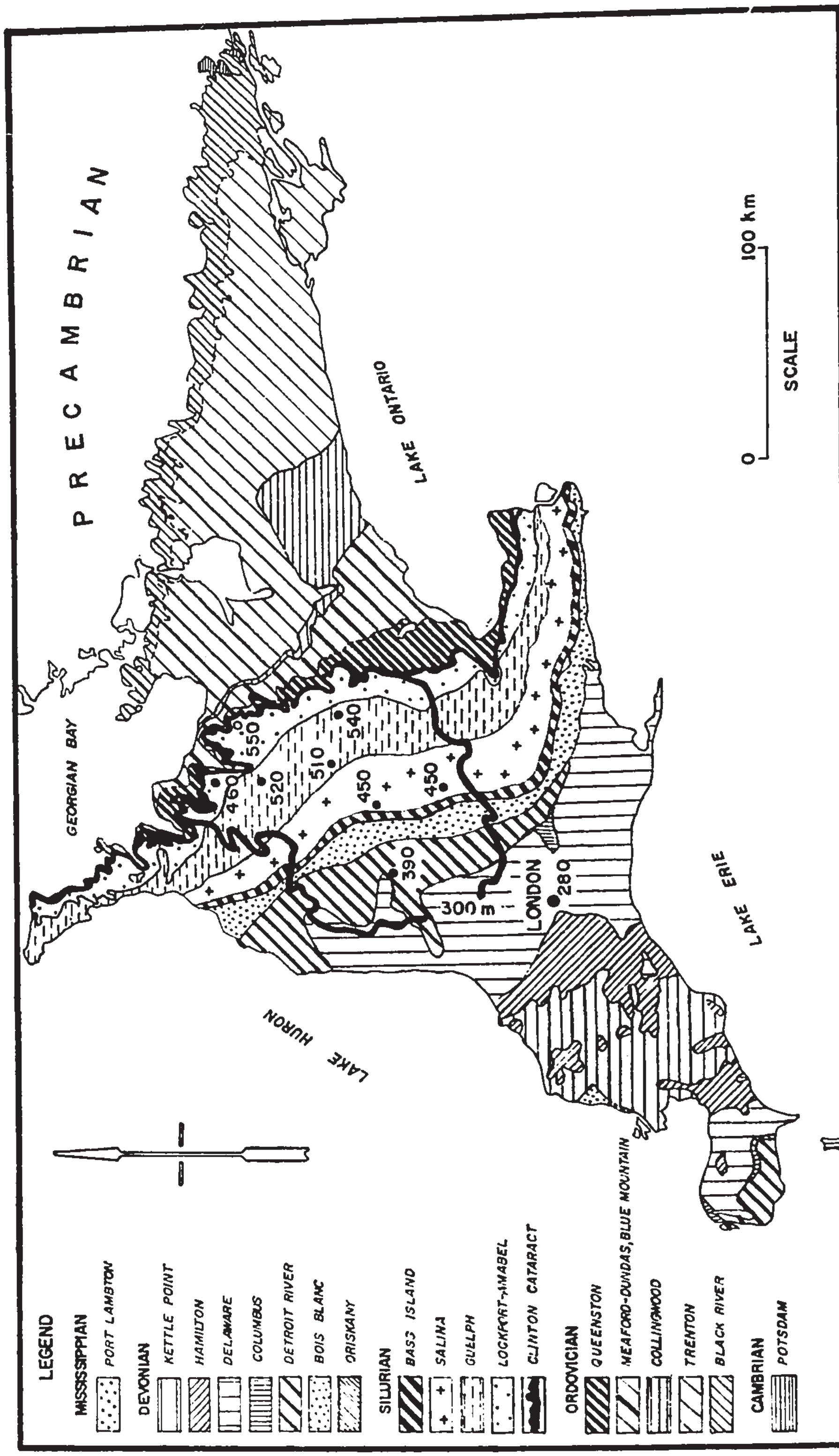


FIG. 1. Geology of the region with 300-m contours.

observation of temperature and thermal conductivity over these sections will yield valid heat flow values.

The principal objective of this paper is therefore to obtain a reliable plot of heat flow versus depth in a borehole and from this to estimate how short a randomly chosen section of borehole could be and still give a heat flow value that is representative of the whole borehole.

Some attempt has been made to correlate thermal conductivity with density and porosity; if such correlations can be demonstrated, and if only spot measurements of thermal conductivity can be made, it may be possible to use approximate weighting factors for conductivity in the unknown intervals using the geophysical logs that are normally obtained. The data were also used to deduce information about recent climatic changes.

2. The borehole

The borehole was commenced on 1963 September 18 and completed on October 23 after passing vertically ($\pm 1°$) through Palaeozoic sediments to a depth of 594 m. The collar elevation is 248 m above mean sea level with coordinates 43° 00·6′ N, 81° 16·3′ W. During drilling, records of input and output water temperatures and water flow rates were kept and frequent stoppages were arranged in order to make temperature measurements down the drill stem; core recovery in the consolidated rock was essentially complete, the only significant loss being 3 m between 67 and 70 m. Loss of circulation occurred at 55 and 96 m where cementing was required. The hole was cased to 441 m with BX drill rod inside BX casing to 34 m which in turn was inside NX casing to 29 m with H casing to 3 m; the lower section was uncased to determine end effects of a casing on temperatures and heat flow values, and to check the validity of the theoretical assumption that there is no difference in the heat flow values of cased and uncased sections of a borehole.

After 41 m of Pleistocene tills and clays, the hole passes through Palaeozoic sediments commencing in the Devonian Delaware formation and finishing at the top of the Ordovician Meaford–Dundas formation. An indication of the regional surface geology is given in Fig. 1. The regional dip is less than two degrees and the local topographic relief is within ± 20 m about 260 m above mean sea level. Since the beds are relatively flat lying there is no significant regional structural correction to be considered.

3. Temperature measurements

After allowing the hole to return to thermal equilibrium, temperature measurements were made at 3-m intervals, using equipment of the type described by Beck (1963), with a precision that was certainly better than 0·005 °C and may have been better than 0·003 °C. Since the gradient in some sections of the borehole was as low as 10 °C km^{-1}, to take the gradient over 3-m intervals, with individual measurements having errors possibly as high as 0·003 °C, might easily result in errors in the gradient as high as 20 per cent and therefore a noisy curve of temperature gradient versus depth. For this reason the gradient has been calculated over approximately 10 m intervals (i.e. over three consecutive temperature measurement intervals) and is shown in Fig. 2.

The zero gradient (inversion point) occurs at a depth of 80 m. This is too deep to be explained by the annual variation of surface temperature. The only two other possibilities are a longer period change in surface temperature of regular or irregular period, or underground water flows. For climatic variations it is usual to make an assumption about the form of the temperature function, its period, its amplitude and

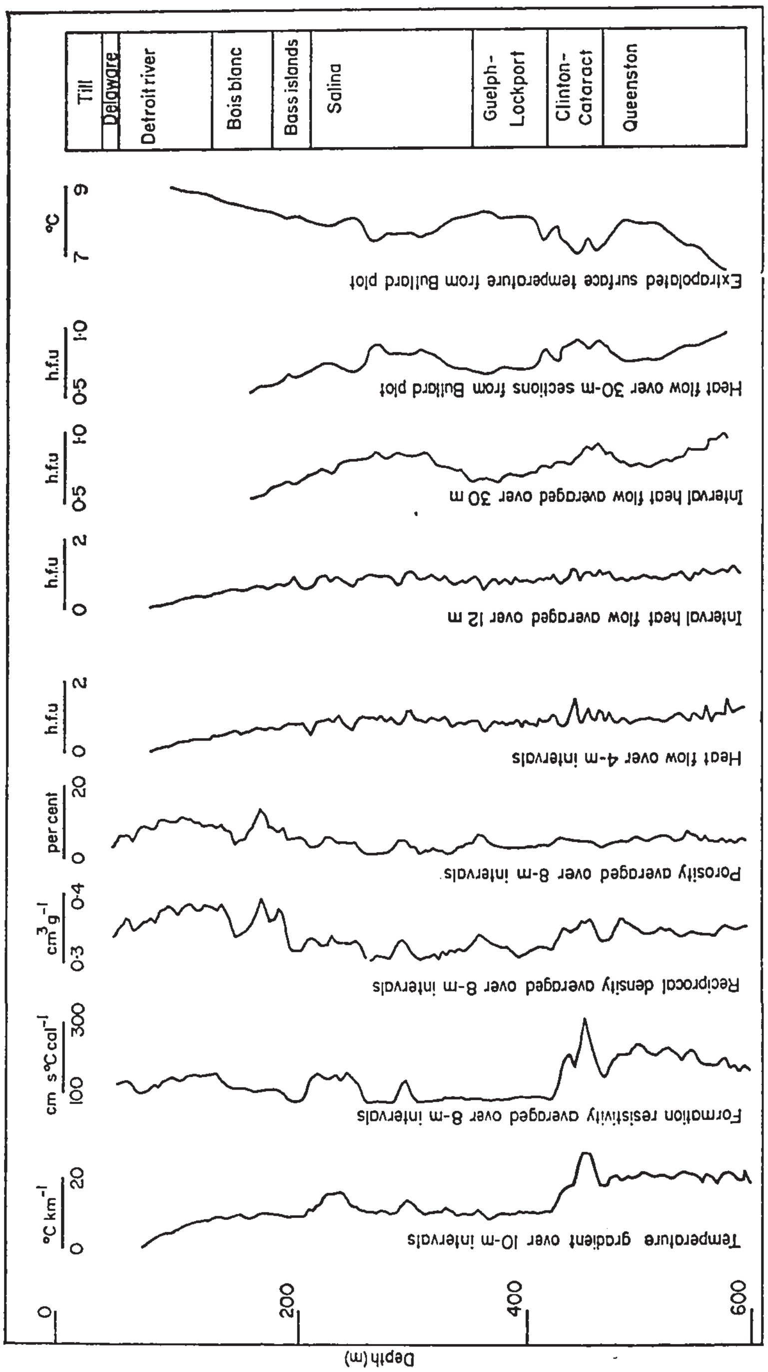


FIG. 2. Plot of various parameters, as labelled on the individual curves, versus depth. Major components of formations: Delaware, limestones; Detroit River, limestones; Bois Blanc, limestones; Bass Islands, dolomites; Salina, dolomites with shales and some anhydrite bands; Guelph–Lockport, dolomites; Clinton–Cataract, shales and limestones with some dolomites; Queenston, shales.

the diffusivity of the formations and make corrections to the temperature at various depths. However, if the temperature measurements are detailed enough it is really only necessary to assume the form of the surface temperature variation and the diffusivity of the formations. The observed data can then be used to test the validity of the assumptions and to obtain the amplitude and period of the variation if the assumptions are correct.

Let T be the observed temperature at depth z, then

$$T = T_0 + gz + V. \tag{1}$$

Where T_0 is the surface temperature before the onset of the climatic change, g is the undisturbed temperature gradient in rocks of diffusivity κ, and V is the temperature disturbance at depth z due to a surface temperature variation of any form. Two forms are frequently used; the step change and the periodic change.

For a step change

$$V = V_0 \operatorname{erf} \frac{z}{2\sqrt{(\kappa t)}}, \tag{2}$$

where V_0 is the step amplitude and t is the time since the change occurred (Jaeger 1965).

Thus from equation (1),

$$T = T_0 + gz + V_0 \operatorname{erf} \frac{z}{2\sqrt{(\kappa t)}} \tag{3}$$

and

$$\frac{\partial T}{\partial z} = g + \frac{V_0}{\sqrt{(\pi \kappa t)}} \exp\left(-\frac{z^2}{4\kappa t}\right) \tag{4}$$

$$\frac{\partial^2 T}{\partial z^2} = -\frac{z}{2\kappa t} \frac{V_0}{\sqrt{(\pi \kappa t)}} \exp\left(-\frac{z^2}{4\kappa t}\right) \tag{5}$$

from equations (4) and (5)

$$\frac{\partial^2 T}{\partial z^2} = \frac{z}{2\kappa t}\left(g - \frac{\partial T}{\partial z}\right). \tag{6}$$

When $\partial^2 T/\partial z^2 = 0$, $g = \partial T/\partial z$ provided $z/\kappa t$ is not very large, as would be the case in the upper levels of the hole. Since T and z are observed values, $\partial T/\partial z$, $\partial^2 T/\partial z^2$ and hence g can be found. g could also be found graphically since it is really the undisturbed gradient beneath the disturbed region. In either case, T_0 can be found by extrapolation back to the surface.

If κ is known, only t in equation (6) is unknown, thus a value for t can be computed for each depth point down to the value for z at $\partial^2 T/\partial z^2 = 0$. If the values are inconsistent, the assumption of the step change in temperature at the surface is incorrect. This conclusion is independent of the value of κ provided it does not vary significantly with depth. If the values of t are consistent then V_0 can be found from equations (3), (4) or (5), but this value of V_0 will depend on κ.

For a periodic function,

$$V = V_0 \exp(-2\pi z/\lambda) \cos 2\pi(t/P - z/\lambda), \tag{7}$$

where V_0 is the amplitude, P is the period, t is the time since the initiation of surface temperature variation and λ is the wave length given by (Jaeger 1965)

$$\lambda = \sqrt{(4\pi\kappa P)} \tag{8}$$

differentiating equation (7) twice with respect to z and dividing the results by equation (7) we obtain

$$\frac{1}{V}\frac{\partial V}{\partial z}=\frac{2\pi}{\lambda}\left[\tan 2\pi\left(\frac{t}{P}-\frac{z}{\lambda}\right)-1\right] \tag{9}$$

and

$$\frac{1}{V}\frac{\partial^2 V}{\partial z^2}=\frac{-8\pi^2}{\lambda}\tan 2\pi\left(\frac{t}{P}=\frac{z}{\lambda}\right). \tag{10}$$

From equation (9) when $\partial V/\partial z=0$,

$$\tan 2\pi\left(\frac{t}{P}-\frac{z}{\lambda}\right)=1$$

or

$$\frac{t}{P}-\frac{z^1}{\lambda}=\frac{1}{8}, \tag{11}$$

where z_1 is the depth where the gradient in the disturbance is zero.

From equation (10) when $\partial^2 V/\partial z^2=0$,

$$\tan 2\pi\left(\frac{t}{P}-\frac{z_2}{\lambda}\right)=0$$

or

$$\frac{t}{P}-\frac{z_2}{\lambda}=0, \tag{12}$$

where z_2 is the depth where the second derivative first becomes zero. Some care has to be exercised here since the second derivative is sensitive to changes in thermal conductivity and to small errors of temperature measurement and may first swing through zero from one of these causes. This risk can be considerably reduced by smoothing over relatively large sections of, say 30 m.

Equations (11) and (12) may be solved for λ and t/P. The validity of assuming a form of surface temperature variation given in equation (7) can therefore be checked by substituting the values of λ, t/P and κ in equation (7) and solving for V_0 for each depth; the values of V_0 should be consistent if the assumptions are correct. An incorrect assumption in the value of κ will not affect the consistency of the V_0 values but only their magnitude.

A further check should be possible since from equations (9) and (10) we obtain

$$k^2+k\left(\frac{\partial V}{\partial z}\Big/V\right)+\frac{1}{2}\left(\frac{\partial^2 V}{\partial z^2}\Big/V\right), \tag{13}$$

where $k=2\pi/\lambda$.

Equation (13) can be solved for k and therefore λ, at any depth z, and hence P, t, and V_0 can be found from the above relations if λ is known and does not vary significantly.

Unless the diffusivity of rock has been measured it is usually assumed to be 0·01 cgs units for the purposes of rough calculations. In the present borehole, *in situ* measurements of thermal conductivity were made at 3- or 4-m intervals from 70 to 90 m using techniques similar to those of Beck & Beck (1965); the same experimental data yielded diffusivities ranging from 0·01 to 0·017. It is believed that much of this spread is due to experimental error since it can be shown (Blackwell, personal communication) that in this type of transient measurement the accuracy of the thermal conductivity determination is approximately five times better than that of the diffusivity determination; it is estimated that the error in conductivity is somewhere between 5 and 10 per cent so that errors of 25 to 50 per cent in the diffusivity figures can be expected.

Thus for the purpose of applying these tests to the upper part of the borehole the diffusivity is assumed to be constant and equal to 0·0137, the mean value of the eight values obtained. The results of applying these tests are shown in Table 1 for the depth range of 39 m, just above the base of the tills, to 107 m, below which depth the temperature differences between the observed temperatures and the mean undisturbed gradient, g, become too small to give valid results.

Table 1

Data for recent climatic changes by applying observed underground temperatures to equations (2) *and* (7). *In both cases diffusivity* = 0·0137, *undisturbed gradient* = 5·60 °C km^{-1}, *extrapolated surface temperature* = 8·5 °C

Depth (m)	t (yr)	Step function (equation (2)) V_0 (°C)	Periodic function (equation (7)) V_0 (°C)	
39·1	7	−3·24	−1·72	
42·1	8	−3·08	−1·56	
45·0	9	−2·76	−1·43	
47·9	10	−2·59	−1·33	
50·8	11	−2·34	−1·25	
53·8	12	−2·04	−1·20	
56·7	13	−1·76	−1·15	
60·1	17	−1·48	−1·13	
63·5	21	−1·29	−1·11	
67·2	24	−1·17	−1·09	
70·8	30	−1·07	−1·09	
74·3	35	−0·97	−1·10	p = 107 yr
77·7	36	−0·96	−1·09	t = 75 yr
81·0	39	−0·93	−1·11	λ = 241 m
84·3	34	−1·02	−1·10	
88·4	29	−1·23	−1·10	
92·0	27	−1·37	−1·06	
95·0	22	−1·76	−1·09	
97·9	20	−2·00	−1·11	
100·9	21	−1·91	−1·23	
103·9	22	−1·63	−1·41	
106·9	23	−1·40	−1·71	
Means	20	−1·6	−1·1	
95% confidence limits	9	0·3	0·1	

It can be seen that the data obtained by assuming a periodic function for the variation in the mean annual ground temperature are much more consistent than those obtained by assuming the step function. In fact the data for the periodic function from 50 m to 100 m are surprisingly consistent. The curvature that is apparent at each end of the range is simply an indication of the departure of the assumed ideal surface temperature function from the real function. The most objective conclusion that can be reached is that the mean annual ground temperature variation approximates a periodic function and is certainly a better approximation than a step function.

To check the validity of these results meteorological data on air temperatures were collected for Guelph, London, Stratford, Woodstock and Toronto, the first four cities being in the same climatic belt. Although Toronto is on the edge of a large lake it was included because its records go back for more than 125 years. Mean

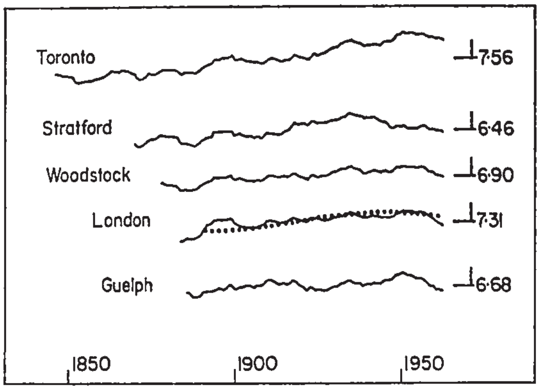


FIG. 3. Ten-year moving average temperatures for Guelph, London, Stratford, Toronto and Woodstock. Horizontal line indicates mean temperature for period of records taken; vertical bar indicates 1°C. + is theoretical curve (equation (7)) using $t = 75$ years, $P = 107$ years, $V_0 = \frac{1}{2}$°C.

monthly air temperatures were first obtained from the monthly reports of the Meteorological Service, but later checked against the original abstracts and re-computed data (known as ARDA 3) at the Department of Transport, Climatology Division, in Toronto.

After corrections were applied for small changes in station location and short periods of missing records, the data for each station were analysed to give the 10-yr moving average temperature and the mean temperature for each station since the readings were commenced. The results are shown in Fig. 3. For Toronto the fluctuations in temperature similar to those shown for the other cities are superimposed on a larger effect due to urbanization.

At Guelph, the average soil temperature at a depth of 1·5 m for the five years since ground temperatures have been measured is 8·5 °C while the mean air temperature for the same period is 6·2, a difference of 2·3 °C. At London the mean annual air temperature averaged over the last five years is 6·9; applying the Guelph correction to this value we obtain a figure of 9·2 °C for the mean annual ground temperature at London. The surface temperature obtained by extrapolating the undisturbed gradient over the range 100–200 m is 8·5 °C; the extrapolated surface temperatures from the Bullard plots (Bullard 1939) for the same range vary from 8·7 at the top to 8·2 °C at the bottom with a mean value of 8·4 °C. The agreement between these two is good, although the difference of 0·7 to 0·8 °C between this temperature and the present estimated mean annual ground temperature for London is only in reasonable, not good, agreement with the amplitude, V_0, in Table 1 for the periodic function. The agreement could be improved if the value for κ is decreased since, as pointed out earlier, the value of κ affects the magnitude but not the distribution of V_0.

It is therefore concluded that the deep inversion region in this borehole is primarily due to climatic variation (which approximates a periodic function) and is not a result of underground waterflows.

4. Thermal conductivity data

The core from the borehole has been sampled in considerable detail. The basic sample interval was approximately 4 m, although for separate studies in some regions of the hole samples were spaced as close as 30 cm and occasionally sets of discs from one piece of core were also prepared. Except for the samples for special studies, no attention was paid to the nature of the core; in this sense the sampling was random although the systematic sample interval prevents use of this term in its strictest sense. In this paper, only the single disc data is used.

From each core a disc was prepared in the usual manner and the thermal conductivity determined with a precision of 2 per cent on a divided bar apparatus similar to that described by Beck (1957).

The porosity and density of a sample were also measured to determine if there was any correlation between these two quantities and the thermal conductivity of the sample. The smoothed data, with appropriate conversions to facilitate comparison, are plotted in Fig. 2.

Although the overall correlation is reasonable there are two regions, one centred at 100 m in the Bois–Blanc formation and the other at 350 m at the base of the Salina, where an increase in the porosity and the reciprocal density is not accompanied by an increase of the formation resistivity. The later discrepancy corresponds with a region of low heat flow in the borehole, as will be discussed in the next section; it is unfortunate that the other region lies in the section of the borehole that is still disturbed by the long period surface temperature fluctuations.

5. Heat flow values

It appears from Fig. 2 that there is remarkably good agreement between the thermal resistivity and temperature gradient; however a close examination shows that this does not lead to a constant heat flow with depth. The heat flow across a 4-m interval was calculated by taking the product of the mean conductivity of two consecutive samples and the linear gradient between the two sample points; the results are plotted in Fig. 2 as the heat flow over 4-m intervals. The sharp ' spike ' at 446 m is genuine and is caused by an unfortunate coincidence of a short section of high conductivity material with the disturbed gradient at the end of the casing. The ' noise ' of the curve, due mostly to taking the temperature gradients over a 4-m interval, obscures any genuine variations of heat flow with depth. Although averaging the heat flow over 12-m intervals shows some improvement, it is only when the interval heat flow is averaged over 30-m, and the horizontal scale expanded four times, that the variations of heat flow with depth become obvious.

As a check on these results the heat flow and extrapolated surface temperature were also calculated from equation 14 (Bullard 1939) over the same intervals used to obtain the curve for interval heat flow averaged over 30 m,

$$T_z = T_0 + H\,\Sigma_i\, D_i/K_i\,, \tag{14}$$

where T_z is the temperature at depth $z = \Sigma_i D_i$, R_i is the thermal resistivity of the ith homogeneous section of thickness D_i, and T_0 is the surface temperature. It can be seen from Fig. 2 that, although there are differences in detail the overall characters of two curves are the same; as is to be expected, the plot of extrapolated surface temperature versus depth is almost a mirror image of the heat flow values versus depth for the Bullard plot.

For the interval heat flow averaged over 30 m the coefficient of variation (r.m.s. deviation expressed as a percentage) was usually around 4 or 5 per cent with the worst being a little over 8 per cent; for the Bullard plot the coefficients of variation were usually less than 1 per cent with the worst being 2 per cent.

The hole was divided into longer sections and the process repeated, the results being summarized in Table 2, which also shows the results after correcting for the Wisconsin glaciation which will be discussed next.

Table 2

Summary of heat flow values before and after correction for Wisconsin glaciation

				Before glaciation correction					After glaciation correction				
				From Bullard plot			From interval heat flow		From Bullard plot			From interval heat flow	
L	Depth range (m)		N	T_0	H	E	H	E	T_0	H	E	H	E
	192 to	262	16	7·86	0·68	1·5	0·71	4·7	7·94	1·09	1·3	1·23	4·6
	258	325	18	7·47	0·08	0·5	0·82	3·0	6·89	1·40	0·8	1·47	2·9
	320	391	20	7·99	0·67	0·6	0·68	3·5	7·34	1·29	0·3	1·31	1·9
70	389	458	23	7·42	0·79	1·9	0·73	3·7	8·04	1·14	1·7	1·18	3·3
	456	530	23	7·64	0·75	1·7	0·79	3·2	8·88	0·99	2·0	1·03	3·0
	527	596	19	6·87	0·86	1·3	0·87	4·7	7·98	1·11	1·3	1·12	4·1
	192	331	35	7·73	0·72	2·1	0·76	2·9	7·56	1·22	3·2	1·34	2·9
140	327	468	42	7·64	0·74	3·2	0·72	2·8	7·82	1·18	2·4	1·22	2·1
	465	596	37	7·45	0·78	3·2	0·82	3·1	8·69	1·02	3·1	1·07	2·8
400	192	596	110	7·58	0·76	4·5	0·76	1·8	7·95	1·13	9·4	1·21	1·8

L = approximate section length (m)
N = number of points
T_0 = extrapolated surface temperature in °C
H = heat flow in μ cal cm^{-2} s^{-1}
E = coefficient of variation (%)

Time on onset 100 000 yr ago
Time of retreat 10 000 yr ago
Amplitude of temperature step = 10 °C

6. Glaciation correction

The regularly increasing heat flow in the lower 100 m of the borehole suggests that at least part of the variation might be due to the onset and retreat of an ice sheet, with the other variations being superimposed on this. The major effect will be from the Wisconsin ice age with an onset time (t_1) approximately 100 000 years ago and a retreat time (t_2) of approximately 10 000 years ago (Dreimanis 1964). The onset and retreat is considered to be sudden and therefore a step function of temperature versus time, similar to equation (2), has been used (Jaeger 1965). The principal effect of applying this correction is to increase the heat flow values in any section of the borehole with slight improvements in the variance in some sections and a worsening of the variance in other sections. Corrections have been tried using other values for V_0, t_1 and t_2; again, the variance in some sections improves but the improvement is not general except in those cases where t_1 and t_2 are clearly in error on geological grounds. The numerical data are summarized in Table 2 and sample curves are given in Fig. 4.

7. Discussion of heat flow variation with depth

It is clear that the variations of heat flow with depth in this hole are real, the minimum value over a 30-m section being $0{\cdot}62\mu$ cal cm^{-1} s^{-1} between 365 and 395 m, and the maximum value being 0·92 at the bottom of the hole. There appear to be only four possible explanations for these variations.

First, underground waterflows are bringing cooler water (or warmer water, depending on which heat flow value is regarded as normal) into various regions of the borehole. Second, there are endothermic and/or exothermic chemical reactions taking place as part of the geological processes. Third, some sections of the borehole may contain more heat generating radioactive material than others, thus creating heat sources in these sections; this is more or less equivalent to an exothermic chemical reaction noted previously. Finally, the borehole may pass through or near geological structures with distort the heat flow lines.

The first possibility was dealt with by Judge & Beck (1967). Very briefly, they took the mean heat flow value for the borehole as being the normal value and argued that since the broad minimum in the heat flow contour correlated well with a maximum

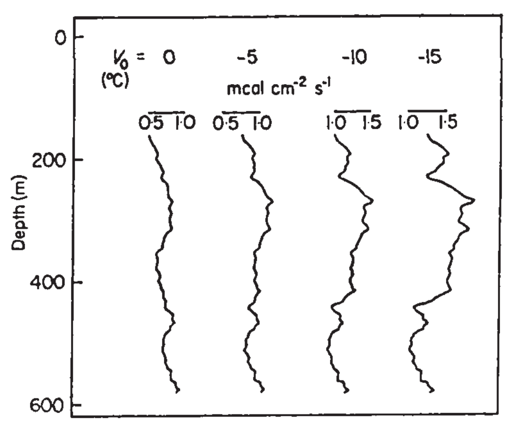


FIG. 4. Sample curves before and after correcting for the Wisconsin glaciation with assumed onset time of 100 000 years ago and retreat time of 10 000 years ago. V_0°C is the assumed amplitude of the temperature change at time of onset, with $V_0 = -10$ °C being the most likely value.

in the porosity curve and was centred close to the boundary of the Salina and the Guelph–Lockport formations, and since these two formations outcrop in regions that are 100–200 m higher in elevation than London, this would produce 400–500 m head of water in the formations under London. This combined with the possibility of seepage into the formation of cooler rainwater and snow melt waters seemed to provide the most likely explanation for the minimum.

A straightforward seepage of meteoric water into the sedimentary basin presents considerable problems. First, for a basin to recharge, it must be discharging; second, the *in situ* waters are generally saline and therefore of higher density than the seepage waters. A possible mechanism of regional significance is that after shallow penetration, the meteoric waters slowly leached out salt, becoming denser and therefore continuing to gradually move down the hydraulic gradient. A mechanism of more local significance might apply to the London region.

London lies in the marginal zone of Salina salt; the salt is fairly continuous in the downbasin direction to the west but has been largely leached away to the east. However, several isolated salt masses remain, one of them only a few kilometres to the east, near the village of Crumlin, and up the hydraulic gradient from London. Thus meteoric water seeping into the region of this salt body may increase in density and then, by displacing less saline water, move down through the geological horizons until stopped by the impermeable shales of the Rochester formation, and down the hydraulic gradient.

Whatever the details of the mechanism might be, the waterflow would have been going on for a very long period of time; thus the effects would not be concentrated in the layer of high porosity but would have disturbed the regions on either side of it due to conduction of heat into the porous layers. Although this mechanism seems to provide a reasonable explanation for the low heat flow region, it leaves unexplained the regions of high heat flow.

Little is known about the geological processes, such as dolomitization of limestone, but if there are any significant thermal effects they are more likely to be of an exothermic than an endothermic nature. However, dolomitization of limestone cannot be invoked to explain the high heat flow peak centred in the Clinton–Cataract formation since it is associated with a shaly region of very high thermal resistivity.

With regard to the third possibility, it seems unlikely that there would be significant differences in heat producing radioactive material in sedimentary rocks, although it is hoped to investigate this point at a later date.

The first three possible explanations deal essentially with a transient situation so that by repeating temperature measurements down the borehole every few years it may be possible to detect long-term changes in the temperature gradient at various depths.

The fourth possibility of some unknown structure close to the borehole affecting the results, cannot be adequately discussed. On a regional scale the geology of the area is well known and relatively simple, but it is known that on a small scale various beds and sub-units pinch out, with the Salina being one of the more complex formations in this regard. Unfortunately, this borehole is the only one in the London region of any significant depth and there is little point in discussing hypothetical structures at this stage until more detail of the immediate area within a 1-km radius is known.

Thus, to explain the variation of heat flow with depth we have to choose between four rather unsatisfactory possibilities, none of which are really capable of being proven without a great deal of detailed and expensive work. However, it is worth pointing out that if we accept the mean heat flow value for the whole borehole, uncorrected for glaciation effects, as $0{\cdot}76 \pm 0{\cdot}03$ then if we had made detailed measurements in a randomly selected 30-m section of the borehole the lowest possible value that could have been obtained would have been $0{\cdot}62 \pm 0{\cdot}03$, and the highest possible value that could have been obtained would have been $0{\cdot}92 \pm 0{\cdot}07$; that is, the heat flow value obtained in any randomly selected 30-m section would be in error by not more than 20 per cent. This is probably quite reasonable in a borehole where the conductivity varies from 3·0 to 14·2 mcal $cm^{-1}\ s^{-1}\ {}^{\circ}C^{-1}$.

Finally, it might be noted that heat flow values have been obtained in about 30 boreholes in southwestern Ontario, details of which will be given in a later paper, and that all of them give heat flow values, uncorrected for glaciation effects, lying between 0·9 and 1·1 h.f.u. with no observable regional variation of heat flow with depth. This is not really surprising since the U.W.O. borehole is the only one in which there has been complete core recovery and for which there has been the opportunity to make very detailed measurements of both thermal conductivities and temperatures.

8. Conclusions

It is concluded that:

1. Detailed temperature measurements, particularly in the upper levels of a borehole through reasonably uniform rocks, may yield useful information on climatic changes that have occurred in the past few decades.

2. The heat flow in a borehole may vary by ± 20 per cent with depth due to causes which are at present not clearly understood.

3. Notwithstanding close error limits that are often put on heat flow values the errors in a heat flow determination of a single borehole may be considerably greater than has been previously suspected.

4. A heat flow value determined from a randomly selected 30-m section of borehole may yield a useful heat flow value in the sense that a value of moderate accuracy is better than none at all. However, it would clearly be better if more than one 30-m section was available. This leads to the suggestion that it may be possible to use many oil wells in which cores have been taken from relatively short sections of the hole.

5. There is sufficient correlation between conductivity, density and porosity to warrant further investigation.

Acknowledgments

We are indebted to the Department of Energy, Mines and Resources for financing the drilling of the borehole; to the National Research Council for a grant in aid of research; to the Province of Ontario Graduate Fellowship Fund; to Professor R. W. Packer for pointing out the large amount of climatic information available for the region, and to the Climatology Division of the Department of Transport, Toronto, for giving us access to original records and unpublished data.

1969 *March*.

References

Beck, A. E., 1963. Lightweight borehole temperature measuring equipment for resustance thermometers, *J. Sci. Instr.*, **40**, 452–454.

Beck, A.E., 1965. Measuring heat flow on land, *Terrestrial Heat Flow*, pp. 24–57, ed. by W. H. K. Lee, American Geophysical Union, Washington.

Beck, A. E. Jaeger, J. C. & Newstead, G. N., 1956. The measurement of the thermal conductivities of rocks by observations in boreholes, *Aust. J. Phys.*, **9**, 286–296.

Beck, J. M. & Beck, A. E., 1965. Computing thermal conductivities of rocks from chips and conventional specimens, *J. geophys. Res.*, **70**, 5227–5240.

Bullard, E. C., 1939. Heat flow in South Africa, *Proc. R. Soc.*, **A173**, 474–502.

Diment, W. H. & Robertson, E. C., 1963. Temperature, thermal conductivity and heat flow in a drilled hole near Oak Ridge, Tennessee, *J. geophys. Res.*, **68**, 5035–5047.

Dreimanis, A., 1964. Notes on the Pleistocene time-scale in Canada, *Geochronology in Canada*, pp. 139–156, ed. by F. F. Osborne, Royal Society of Canada, Toronto.

Jaeger, J. C., 1965. Applied theory of heat conduction, terrestrial heat flow, pp. 7–23, ed. by W. H. K. Lee, American Geophysical Union, Washington.

Judge, A. S. & Beck, A. E., 1967. An anomolous heat flow layer at London, Ontario, *Earth planet. Sci. Lett.*, **3**, 167.

Lee, W. H. K. & Uyeda, S., 1965. Review of heat flow data, *Terrestrial Heat Flow*, pp. 87–190, ed. by W. H. K. Lee, American Geophysical Union, Washington.

Ratcliffe, E. H., 1959. Thermal conductivities of fused and crystalline quartz, *Brit. J. appl. Phys.*, **10**, 22–25.