THE ST. LAWRENCE VALLEY SYSTEM AND ITS TECTONIC SIGNIFICANCE

by

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A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfilment of the requirements for the degree of Doctor of Philosophy.

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March 1974

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ABSTRACT

The St. Lawrence Valley system is a branching group of narrow linear depressions, more than 2,000 km long, set largely on the platform and the Shield just west of the Northern Appalachians. The western half of the valley system is underlain by rifts some of which are well-defined grabens. The structure of the water-covered eastern half is less certain but the available geological information, although scanty, indicates that the submerged parts too lie along rift zones. Moreover, the gross pattern of the valley system is closely similar to the branching pattern of a large rift valley system. Therefore, it is hypothesized that the St. Lawrence Valley system is underlain by a rift system to which the name St. Lawrence Rift system is applied.

Many features of the St. Lawrence Rift system compare closely with those of the rift systems in East African, Baikal and Rhine regions. However, unlike the latter rifts, the St. Lawrence Rift system has little or no geophysical expression; it is not set along the crestal region of a discernible crustal swell; its fault troughs are devoid of Cenozoic volcanic products and have no noteworthy accumulations of Cenozoic sediments. These differences are thought to be largely due to long inactivity of the rift system.

In its present form, the St. Lawrence Rift system appears to have formed in the mid Mesozoic, possibly as a result of tensional stresses related to the opening of the Atlantic Ocean. However, there are indications that the pattern of mid-Mesozoic rifting was largely superimposed on pre-existing faults, also of tensional origin and possibly of Hadrynian age. These older faults are identified as structural elements inherited from the eo-Appalachian rift system of Bird-Dewey and Chidester-Cady models of Appalachian evolution.

RESUME

Le système de la Vallée du Saint-Laurent est formé d'un ensemble ramifié d'étroites dépressions linéaires, de plus de 2,000 km de longueur, situé en grande partie sur la plate-forme et sur le Bouclier Canadien à l'Ouest immédiatement des Appalaches Septentrionales. La demie Ouest de ce système est formée de failles dont certaines sont des grabens bien définis. Les structures de la demie Est sont submergées et moins bien définies, mais les données géologiques disponibles, bien que rares, nous montrent que ces parties submergées sont aussi des zones de failles. De plus, l'ensemble du système ressemble beaucoup au modèle ramifié d'un large système de vallées faillées. Ainsi, nous émettons l'hypothèse que sous le système de la Vallée du Saint-Laurent on trouve un système de failles auque nous donnons le nom de système de Vallées Faillées du Saint-Laurent.

Plusieurs charactéristiques de ce système de Vallées Faillées du Saint-Laurent se comparent de près avec celles des systèmes de vallées faillées de l'Afrique de l'Est, de Baikal et du Rhin. Cependant, contrairement à ces autres endroits, le système de Vallées Faillées du Saint-Laurent a peu ou aucune expression géophysique; il n'est pas situé le long d'une région de crête d'un gonflement perceptible de la croute terrestre; les fosses tectoniques ne renferment pas de roches volcaniques ni de dépôts sédimentaires importants d'age cénozoïque. Nous croyons que ces différences sont dues en grande partie à la longue inactivité du système de failles.

Dans sa condition actuelle, le système de Vallées Faillées du Saint-Laurent semble s'être developpé durant le Mésozoïque Moyen, peut-être due à des forces de tension reliées à l'ouverture de l'océan Atlantique. Cependant, il semble que le modèle de failles du Mésozoïque Moyen a été en grande partie superposé sur des failles pré-existantes aussi d'origine de tension et peut-être d'age Hadrynien. Ces failles plus anciennes sont des éléments de structure qui découlent d'un système de failles eo-Appalachien d'après les modèles de l'évolution des Appalaches proposés par Bird-Dewey et Chidester-Cady.

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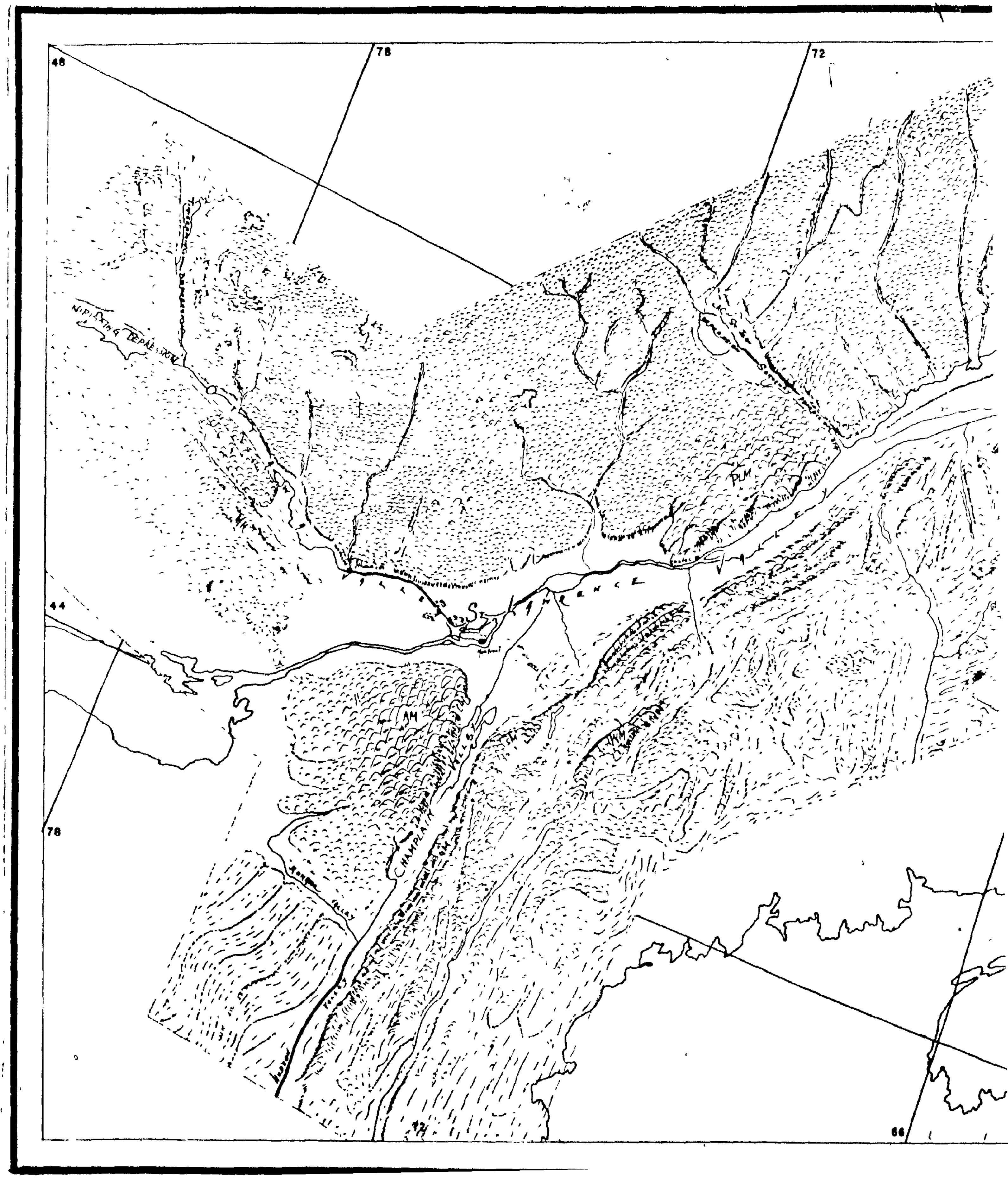
Chapter I: INTRODUCTION

GENERAL STATEMENT

"St. Lawrence Valley system" as used here refers to a branching group of narrow, linear, topographic depressions in northeastern North America (Fig. 1). For convenience of reference, this group of depressions is best divided into eight principal parts. Four of these contain large lakes - Lake Timiskaming, Lake Nipissing, Lake Champlain and Lac. St. Jean - and will be referred to as the Timiskaming Depression, Nipissing Depression, Champlain Valley and Saguenay-Lac St. Jean Depression respectively. The two easternmost parts are narrow submarine troughs and are known as the Laurentian and Esquiman Channels (also referred to as the St. Lawrence and Belle Isle troughs in older literature). The two remaining parts of the St. Lawrence Valley system consist of topographic lows straddling the Ottawa and the St. Lawrence Rivers and will be referred to as the St. Lawrence and Ottawa Valleys respectively. The part of the Ottawa Valley west of the city of Ottawa is commonly called the Upper Ottawa Valley, and the part between Ottawa and Montreal Island, the Lower Ottawa Valley. The term St. Lawrence Region as used in this thesis refers to the general area that provides the setting for the St. Lawrence Valley system. A locality index map of this region is given in Figure 2.

The chain of depressions consisting of the Champlain Valley, the St. Lawrence Valley, the Laurentian Channel (excluding the outer part)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78
72
44
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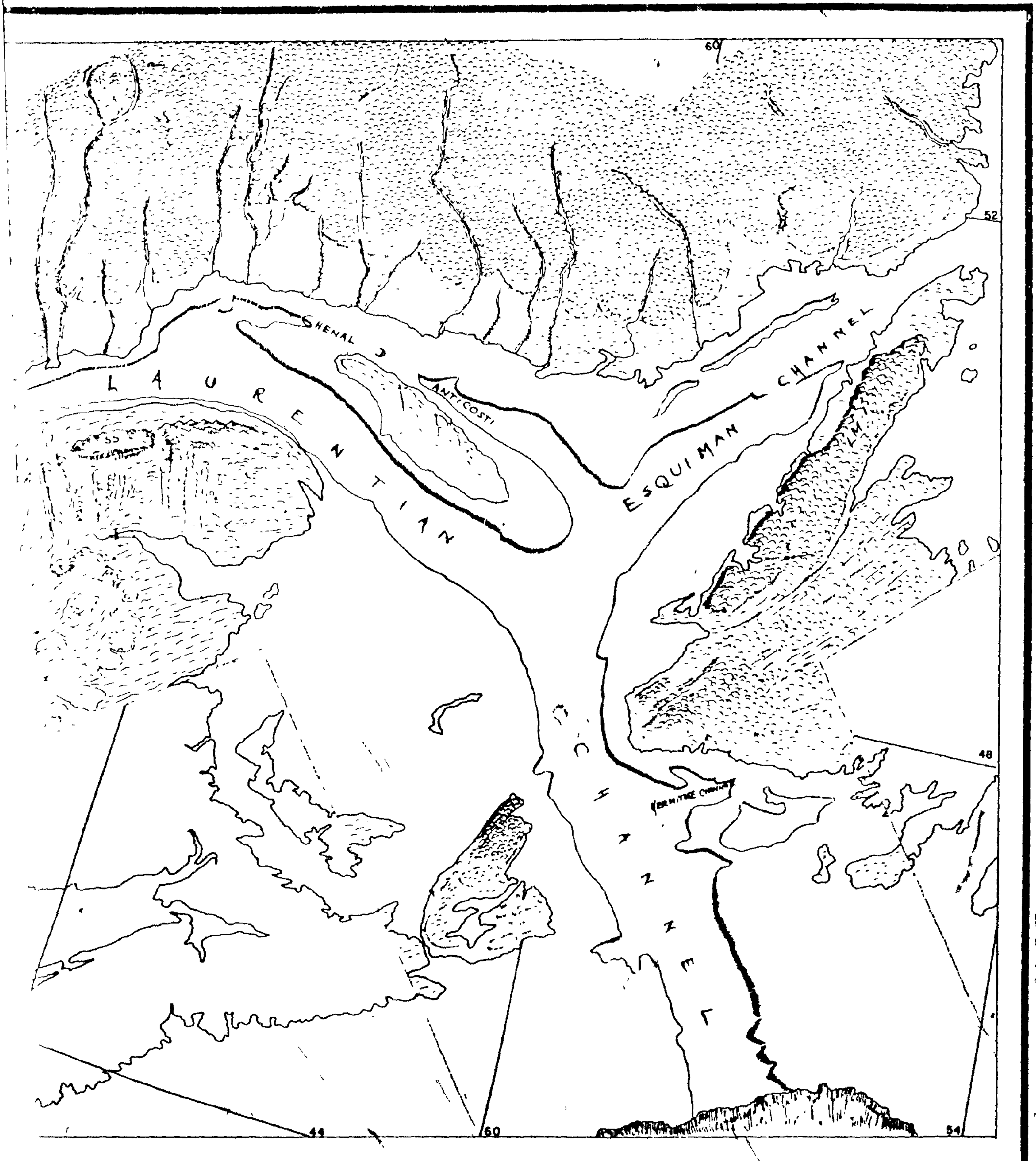
66
42
NIPISSING DEPRESSION
ST. LAWRENCE
VALLEY
PLM
AM
CHAMPLAIN
Hudson



: representation.
- Long Range
ı Mountains

KEY TO ABBREVIATIONS IN FIG.2

AI - Anticosti Island
AM - Adirondack Mountains
BF - Bay of Funday
BS - Betsiamites River
CR - Connecticut River
CS - Cabot Strait
EC - Esquiman Channel
GM - Green Mountains
GR - Gatineau River
GP - Gaspé Peninsula
HAR - Hamilton River
LC - Laurentian Channel
LCH - Lake Champlain
LN - Lake Nipissing
LM - Lake Melville
LRM - Long Range Mountains
LSJ - Lac St. Jean
LT - Lake Timiskaming
M - Montreal
MER - Mécatina River
MH - Madawaska Highlands
MOR - Moisie River
MR - Manicouagan River
NFLD- Island of Newfoundland
NM - Notre Dame Mountains
NR - Natashquan River
NY - New York
O - Ottawa
OR - Ottawa River
PLM - Parc des Laurentides massif
Q - Quebec City
RR - Romaine River
SBI - Strait of Belle Isle
SI - Sept Isles
SJR - St. John River
SLR - St. Lawrence River
SM - Sutton Mountains
SMR - St. Maurice River
SR - Saguenay River
WM - White Mountain

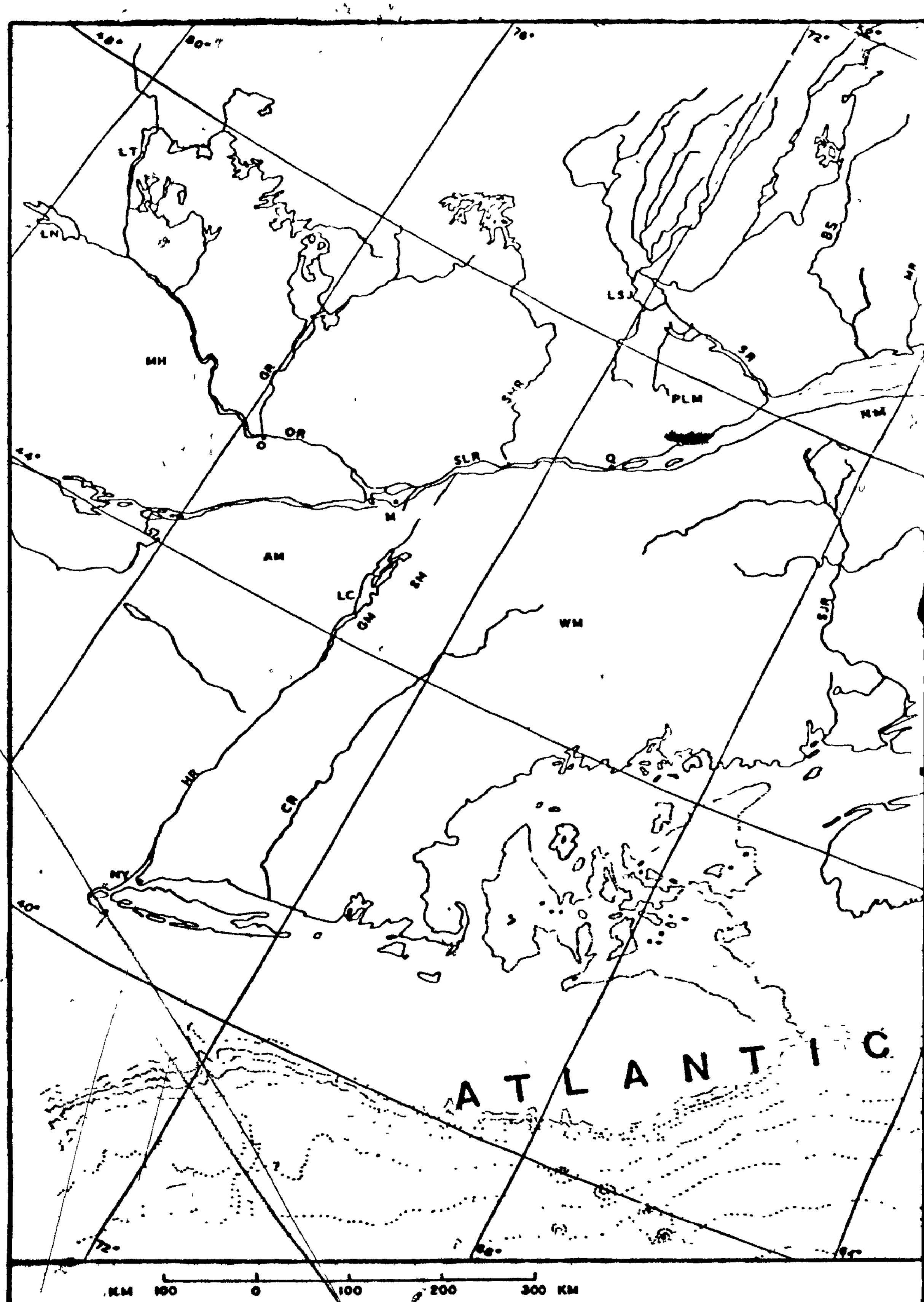


Fig.2.Locality index map of the St.Lawrence Region. Key to abbreviati

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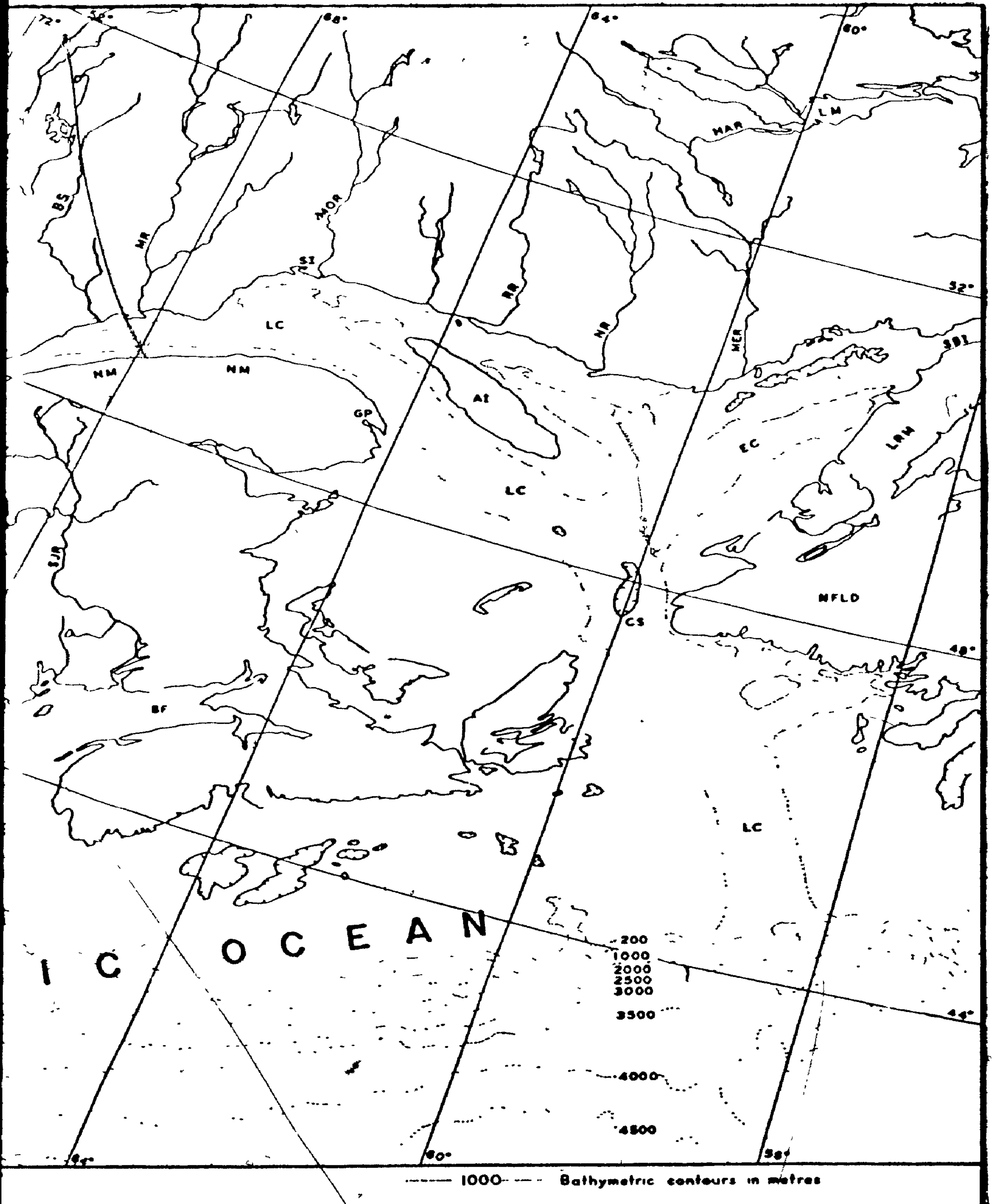

abbreviations on opposite page.

2 of 2

and the Esquiman Channel lies marginally with respect to the north and west boundary of the Northern Appalachians (Fig. 3). For simplicity, this part of the St. Lawrence Valley system will be referred to hereafter as the Marginal Segment. The branching depression consisting of the Ottawa Valley, the Timiskaming and Nipissing Depressions extends across parts of the Canadian Shield and will be referred to as the Shield Segment A;the Saguenay-Lac St. Jean Depression, which also extends across parts of the Canadian Shield will be referred to as the Shield Segment B.

The St. Lawrence Valley system provided early explorers and fur traders the main routes of travel into the interior of North America. Hence, some of the earliest observations on the geology and physiography of the northeastern part of the continent were made along some of these depressions. Early writings on these depressions are of a descriptive nature and hardly touch on the problem of their origin. With continuing studies, however, it has generally been recognized that they are not purely erosional features, but reflect in varying degrees their underlying structures.

A.W.G. Wilson (1903) was probably the first to suspect that some of the depressions may be structurally controlled. He suggested that the Timiskaming, Nipissing and Saguenay-Lac St. Jean Depressions are geomorphic expressions of grabens. He also suggested that the Cambro-Ordovician platformal rocks of the St. Lawrence Valley were laid down in a Precambrian depression of the graben type. Later, Kindle and Burling (1915) proposed that the Ottawa Valley and the

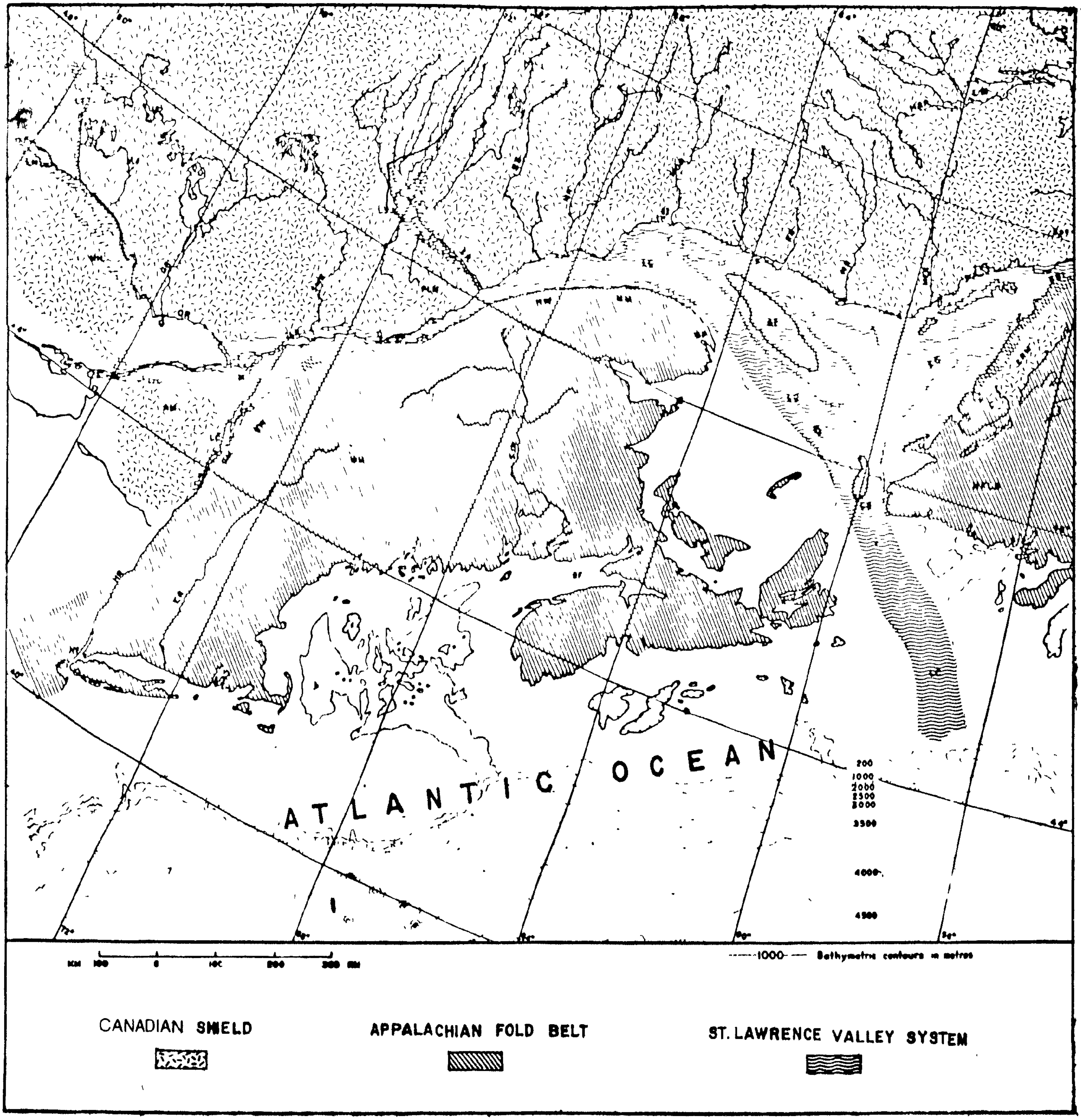

Fig.3. Map of the St.Lawrence Region showing the setting of the St.Lawrence Valley system with respect to major tectonic elements.

St. Lawrence Valley lie along structurally depressed zones of normal faulting. Further studies in the Upper Ottawa Valley by Kay (1942), Lower Ottawa Valley by Wilson (1946, p. 34) and St. Lawrence Valley by Osborne (1956) have shown that Kindle and Burling's view is essentially correct. Furthermore, Kay (1942) has demonstrated that the structure of the Upper Ottawa Valley is a graben. Kay (1942, p. 613) also has shown that the northern boundary fault of this graben extends westwards as the key fault of a structural low along the Nipissing Depression. Detailed studies of the Champlain Valley by Quinn (1933) and parts of the Saguenay-Lac St. Jean and Timiskaming Depressions by Dresser (1916) and Hume (1925) respectively, have revealed that these topographic lows too, are largely underlain by block-faulted, structurally depressed zones. Following the above studies, the structure of the Timiskaming and Saguenay-Lac St. Jean Depressions have gradually come to be accepted as grabens (e.g. Wilson 1959, p. 316;Ollerenshaw and MacQueen 1960;Lovell and Caine 1970). The geology of the Esquiman and Laurentian Channels is largely concealed. However, onshore studies in Esquiman Channel area by Cumming (1967, 1972) indicate that the channel at least in part lies along a zone of downfaulting. High-angle faults also probably dominate the structures of the inner part of the Laurentian Channel, because Anticosti Island, which is the only sizeable island in the Channel, is traversed by numerous high-angle faults (Roliff 1968). The question whether the outer part of the Laurentian Channel is underlain by a major structure or not is a matter of contemporary debate (e.g. King and MacLean 1970a).

Of the St. Lawrence Valley system depressions, therefore, the Timiskaming and Saguenay-Lac St. Jean Depressions and the Upper Ottawa Valley are topographic lows coincident with structures that are generally accepted as grabens. The Nipissing Depression, the Lower Ottawa Valley, the St. Lawrence Valley, the Champlain Valley and the Esquiman Channel also lie along structurally downfaulted zones.

Although graben structures have been known from different parts of the world for quite some time, they did not, until about 1960, receive the attention they now seem to deserve. Most geologists, influenced by the Contraction Hypothesis of the earth, paid more attention to the study of structures such as folds and overthrusts in orogenic belts, which were regarded as manifestations of tangential compression on a contracting earth. They dismissed, as secondary in importance, the tensional structures such as those of large graben systems, although the East African Rift system was known to extend for more than 6,000 km or a 52-degree sector of the circumference of the earth. However, following the discovery in the very early part of the century, that radioactive elements in rocks are a major source of the earth's heat, it has been gradually realized that the earth need not be cooling and contracting as advocated by proponents of the Contraction Hypothesis. Moreover, during the last 15 years or so some workers have given serious attention to the hypothesis that the earth may even be expanding (Egyed 1956, 1957;Carey 1958;Jordan 1971).

The discovery of the world-girdling sub-oceanic ridge system, which for the most part has crack-like crestal grabens (Heezen 1960), has shown that graben structures have developed in the earth's crust on a truly grand scale. Also, it has been argued that at least some of the continental graben systems are extensions of the sub-oceanic system (Girdler 1964, pp. 147-149). The proponents of the earth expansion theory regard large graben structures as manifestations of earth expansion, just as proponents of the Contraction Hypothesis regarded structures in fold mountains as manifestations of global contraction. Also, with the increased interest in the hypothesis of Continental Drift, thermal convection in the mantle has been regarded by many as providing a possible driving force for drifting continents and large graben structures are believed to be localized along zones of upwelling of convection currents (Wilson 1963). The outward flow of currents from these zones is believed to drag the crust apart and produce graben structures. Furthermore, according to the recently proposed hypothesis of plate tectonics (Isacks et al. 1968; Le Pichon 1968; Morgan 1968) the crestal grabens of the sub-oceanic ridge system overlie zones of plate growth that are presumably related to sea floor spreading, and some continental grabens are "failed arms" of plume-generated triple junctions, the other two arms of a triple junction having combined to form a plate boundary (Burke and Dewey 1973).

Thus, graben structures have assumed fundamental importance in the context of theories of earth expansion, mantle convection and plate tectonics and hence are receiving increased attention. This is reflected in the fact that one of the three international programs selected for emphasis by the International Upper Mantle Committee (IUMC), at its meeting in Moscow in 1964, was the world system of large grabens. Since the 1964 meeting, the following meetings were held under the sponsorship of IUMC: a seminar on the East African Rift System was held in Nairobi, in April 1965 (UMC-UNESCO seminar 1965); two symposia on the World Rift System, one in Ottawa in September 1965 (see Irvine, ed. 1966) and the second in Zurich in September 1967 (some papers presented are published in Tectonophysics 8, 1969) and an international rift symposium (mainly on the Rhine graben) in Karlsruhe in 1968 (Illies and Mueller, eds. 1970).

Beyond the purely scientific reasons for the greatly expanded interest in large graben structures, their study has been accelerated because of the recognition of a close temporal and spatial relationship between continental graben structures and alkaline-carbonatite magmatism. Carbonatites contain economic concentrations of some elements, particularly of niobium, rare earths and thorium. Also many carbonatite provinces are kimberlite provinces and kimberlite pipes are the primary source of natural diamonds in the world.

THE THESIS

The graben structures along the Upper Ottawa Valley, the Timiskaming and Saguenay-Lac St. Jean Depressions are not isolated structures but are parts of a system of elongate fault troughs which are crudely coextensive with the St. Lawrence Valley System. This system of fault troughs not only resembles the gross pattern of a large graben system but also has many of the geomorphological and geological characteristics of such a system. These similarities suggest that the St. Lawrence Valley system is underlain by a large continental graben system - the St. Lawrence Rift System (Kumarapeli and Saull 1966a). This, in essence, is the subject of the present thesis. This view of the tectonic significance of the St. Lawrence Valley system should be of interest to those who are concerned with the study of large graben systems. It will be argued that grabens along the St. Lawrence Valley system represent a piece of evidence vital to our understanding of the geological evolution of the Appalachian foldbelt. The graben concept also helps clarify the problem of seismicity and alkaline-carbonatite magmatism and a host of other geological and geomorphological peculiarities of the St. Lawrence Region.

PURPOSE, SCOPE AND NATURE OF THE STUDY

The writer has three main aims in the present study:

1. To synthesize selected aspects of the geological and other characteristics of the St. Lawrence Valley system and its environs (Chapters II to XIII). These syntheses do not constitute a treatment of the entire spectrum of geological and other characteristics of these respective areas, but represent rather a discussion of various aspects (e.g. high-angle faulting) which to the writer appear to be relevant to the graben problem. Other characteristics are either briefly mentioned or omitted altogether.

2. To compare and contrast the various characteristics of the St. Lawrence Valley system structures with those of the rift systems of the three classical areas: Rhine, Baikal and East Africa (Chapter XIV).

3. To speculate on the age and origin of the structures along the St. Lawrence Valley system (Chapter XV).

The St. Lawrence Valley system extends over a very large area. Its length from the north end of Lake Timiskaming to the mouth of the Laurentian Channel is about 2,300 km. Published geological information on this vast region is voluminous. Much has been written, mainly during the first half of this century, on the geomorphology of the region. These writings are largely concerned with the interpretation of landscapes of the region from a Davisian point of view. Geophysical investigations began in the early fifties (e.g. Fitzpatrick 1953;Press and Beckmann 1954) and have since been gathering momentum.

Field studies by the writer included two summers (1966 and 1967) work in the St. Lawrence Valley. This work consisted mainly of 10 gravity traverses over the Appalachian - platform boundary on the east side of the valley (Appendix III) and one traverse over the shield - platform boundary on the west side of the valley (Appendix IV). Visits of two to three days duration were made to several other areas of the St. Lawrence Valley system: the Upper Ottawa Valley in summer 1966, the lower St. Lawrence Valley and the Saguenay-Lac St. Jean Depression in summer 1967, the Lower Ottawa Valley, the Nipissing and the Timiskaming Depressions in summer 1968. During these short visits the main emphasis was on the study of geomorphological and structural features.

In the winter of 1966, three weeks were spent at the National Airphoto Library, Ottawa, studying vertical aerial photographs of the St. Lawrence Valley system and its environs mainly to collect evidence of faulting along and near the margins of the depressions.

PREVIOUS WORK

In 1966, the writer published with V.A. Saull (Kumarapeli and Saull 1966a) evidence supporting the view that the overall structures along the St. Lawrence Valley system is that of a large rift system to which the name St. Lawrence Rift system was proposed. This publication contained the preliminary results and conclusions of the present research project. Further results and updated conclusions are given in the present paper. The 1966 publication has been quoted widely and appears to have generated a considerable amount of interest among researchers, and to have helped focus the attention of some on the graben problem of eastern North America (e.g. Poole 1967;Saull 1967; Doig and Barton Jr. 1968; Florensov et al. 1968; Gerencher and Gold 1968; Solonenko 1968a; Voight et al. 1968;Woollard 1969; Voight 1969; Currie 1970; Doig 1970; Lovell and Caine 1970; Lumbers 1971; Clark 1972). Since the publication of the preliminary report, the writer also has published on topics related to the present research problem (Kumarapeli and Saull 1966b; Kumarapeli 1969; Kumarapeli and Sharma 1969; Kumarapeli 1970).

Prior to 1966 (Kumarapeli and Saull 1966a) the St. Lawrence Valley system as a whole had not been viewed as a geomorphic feature coextensive with a large graben system. Parts of the valley system, however, had been recognized as reflecting graben structures and the previous work on these parts has been cited earlier in this chapter.

Some workers have been critical of certain conclusions contained in the preliminary report. Sheridan and Drake (1968), Keen (1969, 1972), King and MacLean (1970a) have criticized the postulation of a regional structure along the outer part of the Laurential Channel. These criticisms are discussed in Chapter XIII (see pp.182-187). Voight *et al.* (1968) and Voight (1969) have criticized the concept of a presently active St. Lawrence Rift system, dominated by extensional tectonics. This criticism appears valid and is incorporated in the present work (see pp.259-260). One other aspect that has been criticized (e.g. see Northwood 1966 in discussion of Kumarapeli and Saull 1966b) is the postulated extension of a branch of the St. Lawrence Rift system through Lake Superior and thence southwards (Kumarapeli and Saull 1966a, p. 650). This extension was perhaps the most speculative aspect of the original paper. Because no additional evidence has since come to light, and because there are indications that the branch in question terminates in the northern part of Lake Huron (see p.147), such an extension now appears to the writer as improbable and it therefore will not be discussed further in the present work.

SOME DEFINITIONS

The widely used term *rift* is applied in this paper to structures of regional extent formed as a result of extension, without reference to the actual geometry of such structures (Beloussov 1969, p. 539).

The term *graben* is used strictly in a structural sense. It is a rift consisting of an elongate, relatively depressed crustal unit or block that is bounded by faults on both sides (American Geological Institute, Glossary of Geology 1972). Amongst natural examples, symmetrical grabens are rare; all degrees of structural asymmetry are usually present. The structures of large rift systems for the most part are <u>complex grabens</u> of longitudinal, tilted blocks, minor grabens and horsts combined to produce the larger structural depressions.

The term rift valley is used strictly in a geomorphic sense. It is used to denote a topographic depression coincident or nearly so with a graben structure (Dennis 1967).

Paleogeographic directions are referred to the present day geographic directions.

ACKNOWLEDGEMENTS

This study originated from discussions with Professor V.A. Saull on the tectonic framework of seismicity in the St. Lawrence-Great Lakes region (Kumarapeli and Saull 1966a). Prof. Saull acted as the thesis director and Chairman of the Thesis Committee, the other members being Profs. J.A. Elson (1966-1972), L.P. Geldart (1966-1967), E.W. Mountjoy (1969-1974) and H. Helmstaedt (1972-1974). The continued encouragement and help given by the Thesis Committee and by Prof. C.W. Stearn, are gratefully acknowledged. Special thanks are due to Prof. Saull for accompanying the writer on field trips to Upper Ottawa Valley, Lower Ottawa Valley, Lower St. Lawrence Valley and Saguenay-Lac St. Jean Depression and for assistance given in geophysical surveys; to Prof. Mountjoy for bringing to the writer's attention Marshall Kay's (1942) work on the Ottawa-Bonnechere graben; to Profs. Helmstaedt, Mountjoy and Saull for critically reading the August 1972 thesis submission and the present submission; to Prof. W.M. Telford for critically reading the section on gravity studies over Logan's Line (Appendix III); to A.K. Goodacre of the Earth Physics Branch, Department of Energy, Mines and Resources, Ottawa, for computing the gravity data on profiles across Logan's Line and for critically reading the Chapter XIII on the outer part of the Laurentian Channel and to Bijon Sharma, former graduate student of the Department of Mining Engineering and Applied Geophysics, McGill University, for collaboration in the interpretation of gravity data on the profile across the Shield margin. In addition to discussions with various members of the Thesis Committee, the writer also benefited from discussions with several other members of the McGill staff, particularly Profs. T.H. Clark, R. Doig, J.E. Gill, E.H. Kranck, and A.R. Philpotts (formerly at McGill). Numerous other geologists or geophysicists who have contributed unpublished information, suggestions and criticisms including R.A. Gibb and A.K. Goodacre of the Earth Physics Branch, Department of Energy, Mines and Resources, Ottawa and W.H. Poole of the Geological Survey of Canada. Thanks are also due to Prof. N.A. Forensov, Department of Geosciences, Academy of Sciences of the USSR for supplying the writer with extensive literature on the Baikal Region and to S.K. Mahajan, graduate student, Department of Geological Sciences, McGill, for helping in field work.

The writing of this thesis in its present form was carried out while the writer held a faculty position at Sir George Williams University, Montreal. The encouragement given by my colleagues, especially Prof. A.N. Déland was very helpful.

Financial assistance in the form of a McGill University Graduate Fellowship (1966-1967), a Carl Reinhart Scholarship (1967) is gratefully acknowledged. Field, travelling and related expenses during the period May 1966 to August 1967 were paid from National Research Council of Canada grant A-4348 and Geological Survey of Canada grant 3-58 held by Profs. Saull and Telford.

I am thankful to Lorraine Bertrand, Elsie Brawley, Anahid Morsi and Lilian Mikhail for typing the manuscript and to Milan Gilmore and Norman Massé for assistance with draughting.

CHAPTER II: THE APPALACHIAN REGION

INTRODUCTION

The term "Appalachian Region" as used in the present work applies to the northern part of the Appalachian foldbelt, extending from about latitude 43°N to its presumed termination off the north-east margin of the Island of Newfoundland. This part of the foldbelt is bounded on the west and north sides by the Marginal Segment of the St. Lawrence Valley system; the outer part of the Laurentian Channel extends across it. A brief description of the tectonic framework of the region and a summary of views on its tectonic evolution are given here to orient the reader with respect to certain aspects of the regional geology. A brief description of the physiographic characteristics of the region is given in Appendix I.

TECTONIC FRAMEWORK

The Appalachian geosyncline in the problem area developed from late Precambrian to Permian (Poole 1967). The foldbelt consists of a group of distinct mobile zones that were subjected to episodes of deformation, including two main phases referred to as the Taconic (Ordovician) and Acadian (Devonian) orogenies. Of these two, the Acadian orogeny was of wider regional extent; the Taconic orogeny appears to have affected only a relatively narrow zone, about 100 km wide, along the north and west flanks of the foldbelt. Also, the Acadian orogeny led to the development of larger systems of nappes and deep-seated thrusts, to more intense metamorphism and was accompanied by a larger scale of plutonism (Zen 1972, p. 50).

The foldbelt is laterally zoned, stratigraphically as well as tectonically. Several schemes of tectono-stratigraphic zones have been proposed (Schuchert 1930; Bird and Dewey 1970; Williams 1972a; also see Zen 1972). They are basically similar; differences lie mainly in the degrees of complexity of the schemes. A brief and generalized description of the tectonic framework of the Appalachian Region is given here using the three-fold subdivision proposed by Bird and Dewey (1970; also see Schuchert 1930), illustrated in Figure 4.

Zone A (Fig. 4) consists of a belt, first deformed intensively during the Taconic orogeny and later affected to a minor extent by the Acadian orogeny. Bird and Dewey (1970) divide Zone A into two sub-zones: a northwestern strip called Logan's Zone (whose north and west boundaries underlie the Marginal Segment of the St. Lawrence Valley System) and a southeastern strip called the Piedmont. Logan's Zone was a part of the miogeosyncline (dominated by carbonate deposition) before the Taconic orogeny and then became successively a linear zone of block-faulting, exogeosynclinal subsidence (Zen 1972, also see Cady 1969, p. 33) and westward thrusting (Zen 1967,

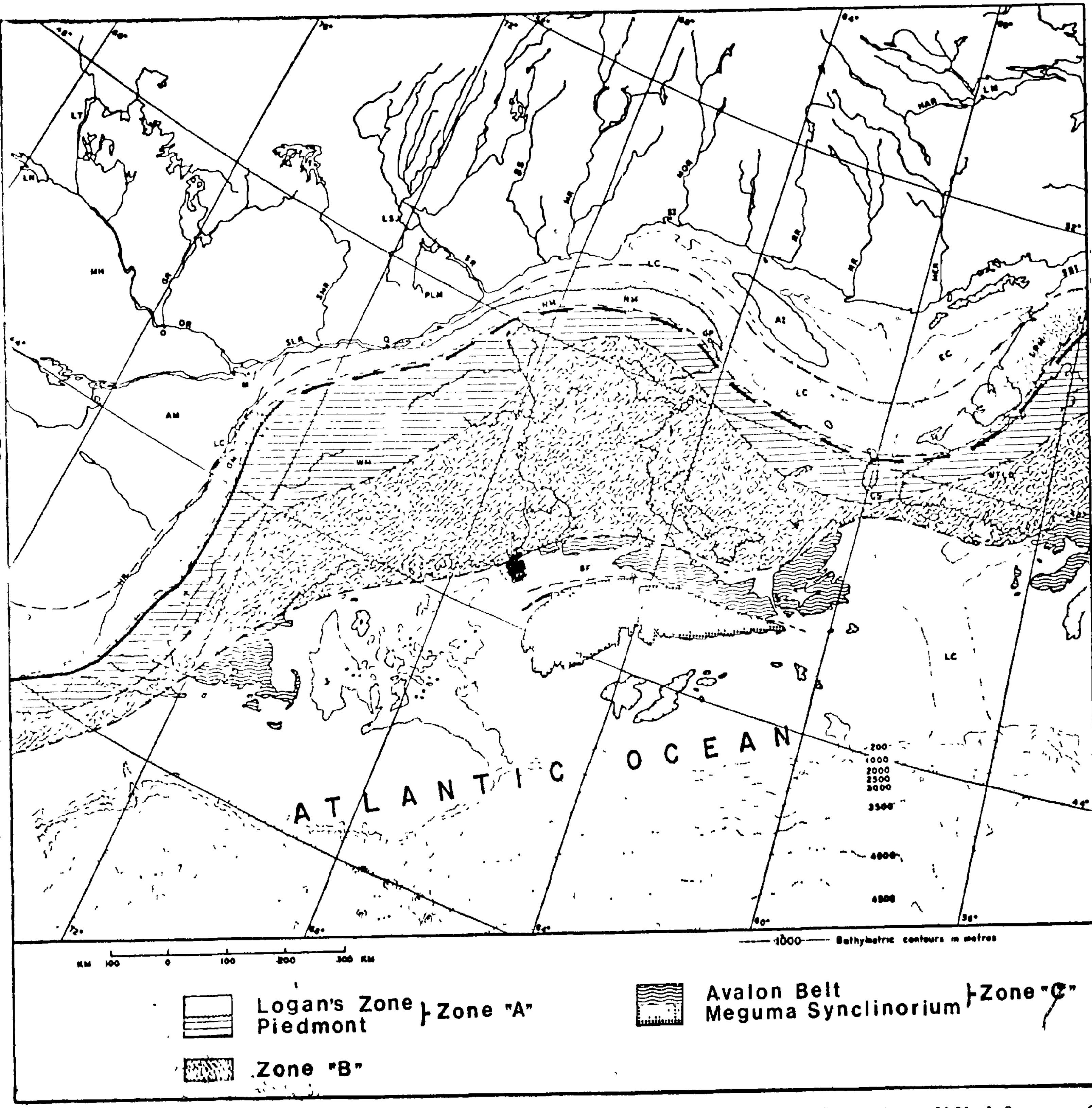

Fig.4. The three main tectono-stratigraphic zones of the Northern Appalachians. Slightly modified from Bird and Dewey (1970). Heavy line indicates the continental margin in Cambrian and early Ordovician times as postulated by Rodgers (1968).

1972; Rodgers and Neale 1963; Williams 1972b) receiving from the southeast large klippen and Taconic flysch followed by coarser clastic sediments. This tectono-sedimentation regime seems to have prevailed over the entire area now occupied by the Marginal Segment of the St. Lawrence Valley System (see Zen 1972). Along the east sides of the St. Lawrence and Champlain Valleys the soles of Taconic thrust sheets are believed to form a more or less continuous feature, known as Logan's Line* (Logan 1863, Clark 1951) in the former area and the Champlain thrust (Welby 1961, p. 193) in the latter. Northwest of Logan's Zone are remnants of a once-extensive platform cover, the St. Lawrence Platform (Poole 1967), which represents an extension of the shelf environment of Logan's Zone. The platformal rocks overlie a Precambrian basement of infra-crustal rocks with K/Ar ages ranging from 800 to 1100 million years approximately (Stockwell 1968). The southeastern margin of Logan's Zone lies along an axis of great structural uplift: the composite Green Mountain - Sutton Mountain - Notre Dame Mountain - Indian Head Range - Long Range Mountain - anticlinoria. This zone was apparently uplifted during the Taconic orogeny and was probably the source area of at least some of the klippen that moved to the north and west (Zen 1972, p. 41). Grenville crystalline rocks form the cores of this line of anticlinoria in the Green Mountain,

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*It is not clear from Bird and Dewey's (1970) description of Logan's Zone whether its western margin includes Logan's Line. The writer has assumed this however.

Indian Head Range and Long Range Mountain areas, suggesting that Logan's Zone is also underlain by Grenville rocks. The Piedmont is a prism of late Precambrian to Ordovician eugeosynclinal deposits (clastic sediments and volcanics; Bird and Dewey 1970) that were deformed and metamorphosed probably during the Taconic orogeny. Thus, the boundary between Logan's Zone and the Piedmont was approximately the site of a major facies change before the Taconic orogeny: a Cambrian - early Ordovician carbonate/ orthoquartzite shelf facies in Logan's Zone changes southeastward to a eugeosynclinal clastic-volcanic facies (Cady 1968; 1969, pp. 15-17). Rodgers (1968) has speculated that this facies boundary was a bank edge, similar to the margin of the present Bahama Banks, and that the drop off from the bank to the southeast may have marked the edge of the Cambro-Ordovician North American continent. Close to this boundary, the Piedmont contains a persistent belt of ultramafic intrusions, probably of Early Ordovician age (Zen 1972, p. 8; Chidester and Cady 1972). The only known rocks of Grenvillian age in the Piedmont occur as thoroughly mobilized equivalents in Chester and related gneiss domes (Faul et al. 1963, p. 7).

Zone B is made up of Cambrian to Lower Devonian sedimentary-volcanic assemblages of eugeosynclinal character. These were deformed during the Acadian orogeny. Representatives of rocks of Grenvillian age are not known in this zone (Bird and Dewey 1970).

Zone C of Bird and Dewey (1970) comprises (i) a northwestern strip called the Avalon "belt" (ii) a southeastern area called the Meguma synclinorium (Fig. 4). The Avalon "belt" is outlined by a series of areas exposing late Precambrian rocks. These areas, if not the entire "belt", acted as geanticlines throughout the Paleozoic (Rodgers 1972, p. 514). The rocks consist of great thicknesses of weakly to moderately metamorphosed clastic, volcano-clastic and volcanic sequences of Hadrynian age (Poole 1967, pp. 14-16; Poole <u>et al</u>. 1970, pp. 231-235; Rodgers 1972, pp. 512-514) intruded locally by granite which give radiometric (Rb/Sr) ages close to Cambrian-Hadrynian boundary (McCartney <u>et al</u>. 1966; Cormier 1972). Three interpretations have been given to the Avalon "belt".

1. It represents outliers of a platform - the Avalon platform of Poole (1967).
2. It is a post-Grenvillian rift zone of the Basin and Range type (Papezik 1970).
3. It represents groups (or possibly one long line) of volcanic islands, flanked by originally deep water basins, formed in the Hadrynian and probably consolidated into a line of sialic nuclei by the beginning of the Paleozoic (Hughes and Brückner 1971; Rodgers 1972).

The results of recent studies of the Avalon "belt" appear to fit best the volcanic island hypothesis (Rodgers 1972; Wiebe 1972; Hughes 1973; Helmstaedt and Tella 1972, 1973).

The Meguma synclinorium was apparently the site of a deep-water basin of eugeosynclinal character that lay to the southeast of the Avalonian "belt". Great thicknesses of Cambro-Ordovician sediments and volcanics that accumulated in this basin were deformed during the Acadian orogeny (Poole 1967; Poole et al. 1970). Thus, the tectono-stratigraphic development of the Meguma trough appears to be similar to that of Zone B.

The three tectono-stratigraphic zones above, when viewed broadly, outline two arcuate belts (Rodgers 1970, p. 4). One of these extends from the southern limits of the region through southern Quebec to Gaspé and is convex to the northwest. The other arc spans the Island of Newfoundland and is convex to the southeast. The area in between the two arcs is covered by the waters of the Gulf of St. Lawrence. Just how the two arcs are interrelated through the water covered area is rather obscure. Any scheme of correlation of the two arcs across the Gulf has to take into account two main peculiarities. One is that on the Gaspé-New Brunswick side of the Gulf, Zone B ~~is~~ about 250 km wide, whereas on the Newfoundland side, it is only about 60 km wide. Williams et al. (1970) have suggested that, if there was no original constriction of the geosyncline, the unusual narrowness of Zone B in Newfoundland can be explained by a combination of folding and transcurrent faulting. The other peculiarity is that, in geological maps, the Taconian and Acadian structures of the Newfoundland arc look askew in relation to those of the mainland arc. This skewness has been explained by some as reflecting an original feature that existed during the developmental stage of the geosyncline (Poole 1967, p. 44; Sheridan and Drake 1968); others however, favour the idea that the Newfoundland arc may be offset to the right by faulting along the present site of the Laurentian Channel (King 1951, p. 91; Drake and Woodward 1963; Rodgers 1970, p. 4). The actual movement of the arc is believed to be an anticlockwise rotation of about 30 degrees (Du Bois 1959; Nairn et al. 1959; Black 1964), the pivot being in the Strait of Belle Isle area. The suggested offset must have taken place in the Devonian or earlier because the Carboniferous and later rocks in the Gulf of St. Lawrence are apparently not disrupted (King 1970, p. 96) The apparent skewness of the Newfoundland arc will be discussed further in Chapter XV.

Other important features of the tectonic framework of the Appalachian Region are post-Acadian successor basins in which great thickness of sediments accumulated (Belt 1968). The youngest of these basins are clearly tensional features as exemplified by the Late Triassic Fundy graben. This graben is one of a chain of similar fault troughs extending as far as Florida and Alabama (Rodgers 1970, pp. 203-210). The overall arrangement of these fault troughs can be likened to a large rift system, comparable in size and style to the East African Rift System of today (Bain 1957; Sanders 1960).

Northeastward, the Appalachian foldbelt strikes out to sea along the coast of Newfoundland and probably continues to the edge of the continental shelf (Sheridan and Drake 1968). Southeastward, it is covered by a platformal cover of Mesozoic and Tertiary strata (McIver 1972).

GEOPHYSICAL EXPRESSION AND THE TECTONIC FRAMEWORK

Results of deep crustal seismic studies (Barret et al. 1964; Ewing et al. 1966; Keen and Loncarevic 1966; Dainty et al. 1966; Sheridan and Drake 1968) indicate that Logan's Zone and possibly also the Avalon "belt" have a single layer "normal" crust 30 to 35 km thick, whereas the Piedmont and Zone B has a thicker (maximum about 45 km) crust with an "intermediate layer" (Dainty et al.1966). The "intermediate layer" has P-wave velocities of 7.3 to 7.5 km/sec and is underlain by an upper mantle with high P-wave velocities of 8.5 to 8.7 km/sec. The crust and upper mantle characteristics of the Meguma trough are not known.

The general level of the Bouguer gravity field over the Appalachian Region is tens of milligals higher than that of the adjacent shield and platform areas to the north and west (Keen 1972). This difference can be related to differences in crust and upper mantle parameters as indicated by deep crustal seismology (Keen 1972; Innes and Argun-Weston 1967). The change from thicker, denser crust in Zone B and the Piedmont to thinner, lighter crust in Logan's Zone must take place roughly along the boundary between Logan's Zone and the Piedmont. This boundary is also roughly coincident with a composite line of gravity highs which is one of the most persistent gravity features of the Appalachian Region. Fitzpatrick (1953), Innes and Argun-Weston (1967), and Diment (1968) conclude that this gravity high cannot be explained solely on the densities of exposed rocks. They explain it by postulating an upwarp of the uppermost mantle material, roughly wedge-shaped in cross-section. Where the gravity ridge attains its highest amplitude, the wedge-shaped rise of mantle is believed to be some 20 km wide at the base and to have 25 km relief (Innes and Argun-Weston 1967, p. 74). During early Paleozoic times, this line of crust/upper mantle change was also roughly coincident with an important paleotectonic line (Rodgers 1968). During the Taconic orogeny it was an axis of structural uplift (Green-Sutton-Notre Dame-Indian Head Range-Long Range Mountain anticlinoria) and intrusion of ultramafic rocks. These tectonic-magmatic activities and the concept of a rise of the uppermost mantle material as postulated by Innes and Argun-Weston (1967) are mutually compatible. Before the Taconic orogeny the zone of crustal change was probably coextensive with the boundary between miogeosynclinal and eugeosynclinal basins.

In as much as the structural uplift along the northeastern margin of Logan's Zone is characterized by a gravity ridge, the structurally depressed western part of Logan's Zone is characterized by a series of gravity troughs. Compared with the gravity highs, the lows are more open features. Along the east side of the St. Lawrence Valley, a gravity trough parallels the St. Lawrence River. Thompson and Garland (1957, p. 127) have suggested that the gravity trough there may be correlated with the presence of low density sedimentary rocks. Innes and Argun-Weston (1967, p. 74) however, take a different view and suggest that the gravity trough in this area may reflect changes in deeper crustal parameters.

SPECULATIONS ON TECTONIC EVOLUTION

Several recent syntheses of the tectonic evolution of the Appalachian Region have been formulated within the framework of the plate-tectonic hypothesis (Isacks et al. 1968; Le Pichon 1968; Morgan 1968). This hypothesis has become increasingly popular and has quickly acquired many supporters but it also has been the target of severe criticism. Its strongest supporters believe that by means of this hypothesis most tectonic-magmatic phenomena of the planet can be explained. The most severe critics (Beloussov 1970; Meyerhoff and Meyerhoff 1972 a & b) on the other hand suggest that the hypothesis is seriously in error, while still others question its global applicability to tectonic-magmatic phenomena (Gilluly 1972; Ilich 1972). As a generalization, however, the plate tectonics hypothesis helps explain many features of the ocean floors and of the circum-Pacific continental margins. It unifies the tectonic and geophysical characteristics of the earth more than any other hypothesis. It has been used in the synthesis of tectono-stratigraphic evolution of fold-mountain belts (e.g. see Dewey and Bird 1970, 1971; Dickinson 1971 a & b). However, caution advocated by some workers against its applicability to all fold-mountain belts seems relevant. Gilluly (1972, p. 407) states that the hypothesis seems applicable to many of the earth's fold-mountain belts - especially those containing considerable volumes of ophiolites - but the features of other fold mountains seem difficult or even impossible to explain through plate models.

Syntheses Based on Plate Tectonics

Recent plate-tectonic syntheses (commonly also referred to as models) of the Northern Appalachians (Dewey 1969; Stevens 1970; Bird and Dewey 1970; Church and Stevens 1971; Schenk 1971) are basically similar. Of these, the one by Bird and Dewey (1970) is the most pertinent for the Appalachian Region and will be referred to as the Bird-Dewey model. Bird and Dewey (1970) support this model by a detailed discussion and interpretation of the tectono-stratigraphic characteristics of Newfoundland and New England areas.

Bird-Dewey Model

It deals primarily with the pre-Devonian evolution of Zones A and B. The following summary description of the model is taken from Bird and Dewey (1970).

> "The Appalachian orogen evolved through a sequence of interrelated sedimentation - deformation - metamorphism patterns within a tectonic belt situated along the eastern margin of a North American continent". Its "stratigraphic-tectonic zones and deformation sequences are related to Late Precambrian to Ordovician expansion, of a proto-Atlantic Ocean. This oceanic opening and closing was achieved by initial extensional necking and graben-like rupture of a single North American-African continental plate and then further expansion of the rift to oceanic dimensions by lithospheric plate accretion, followed by the formation of a Benioff zone and contractional plate loss along a trench-island arc complex marginal to the drifted North American continent. Pre-orogenic Appalachian sedimentation patterns were essentially the same as those found along modern continental margins".

According to the above model, Logan's Zone with its "normal" continental crust and steep gravity gradients represents the pre-Devonian continental slope and part of the shelf; the Piedmont with its two-layer crust, the continental rise, and Zone B, also with two-layer crust, the telescoped part of the proto-Atlantic Ocean. The two main orogenic episodes that deformed Zones A and B are believed to record the collision of North America first with an island arc system (Taconic) and subsequently with Africa (Acadian).

Bird and Dewey (1970) do not discuss the evolution of Zone C in their plate-tectonic scheme. They simply assume (p. 1047) the Zone B/C interface as a continental margin of indeterminate type. Schenk (1971) speculates that the entire Zone C is of African origin: it became attached to the North American continent when the proto-Atlantic Ocean (Wilson 1966) closed in the Devonian and remained attached when the Atlantic reopened in the Mesozoic. Recently, Schenk's synthesis has been severely criticized, on the grounds that his trans-Atlantic correlations are tenous (Hollard and Schaer 1973). Perhaps the most viable interpretation of Zone C in terms of plate tectonics has been proposed by Rodgers (1972). He thinks that the Avalon "belt" may represent a discontinuous chain of offshore volcanic islands formed in the late Precambrian, probably consolidated into sialic nuclei by the beginning of the Paleozoic, and incorporated into the foldbelt when the proto-Atlantic Ocean closed in the Devonian. According to this scheme, the Meguma synclinorium like Zone B, represents a part of the telescoped scar of the proto-Atlantic Ocean.

Post-Acadian successor basins and in particular the Triassic grabens are extensional features. They are generally believed to be related to Mesozoic continental rifting, as a prelude to the opening of the present day Atlantic Ocean (Dewey and Bird 1970, p. 2633).

A survey of recent literature clearly shows that Bird-Dewey model for the evolution of the Appalachian Region is popular among many workers. However, there are those who think that the geological characteristics of the region cannot be fitted into this model. The foremost among these is Cady (1972). He argues that the faunal, stratigraphic, structural and petrologic evidence is contrary to the concept of relict subduction zones as implied in the Bird-Dewey model and that there is no real basis for postulating an ensimatic origin for the geosyncline, except perhaps in Newfoundland where complete ophiolite complexes are present amongst ultramafic rocks (Church and Stevens 1971; Dewey and Bird 1971; Upadhyay et al. 1971). The alpine type ultramafic rocks elsewhere in the foldbelt were apparently intruded into continental crust (Chidester and Cady 1972). Accordingly, an alternative to Bird-Dewey model has been proposed (Cady 1972, Chidester and Cady 1972) and is briefly described below.

Chidester-Cady Model

According to this model, the Appalachian eugeosyncline evolved largely along the site of a continental rift zone which initiated late in the Precambrian across a single North American-African continent, 300 to 800 km northwest of the line along which Africa and North America were to separate by continental drift beginning in the Mesozoic. At its northeast end, presumably off the north coast of present day Newfoundland, the rift zone was connected to an oceanic rift-ridge system. The protracted pre-Acadian orogenic events (including the Taconic orogeny) were related chiefly to extension and differential vertical movements. The apparently compressive Acadian orogeny is explained by a partial closure of the Appalachian geosyncline, by flexural folding of the sialic basement and the geosynclinal contents.

A principal difference between the Bird-Dewey model and the Chidester-Cady model is that according to the former the Appalachian foldbelt is ensimatic and pericontinental whereas according to the latter it is largely ensialic and intracontinental. This debate on ensialic vs ensimatic origin of the Appalachian geosyncline is likely to be a continuing one. Data as commonly presented to support one point of view or the other seem to have large elements of interpretative bias. A part of this bias is perhaps peculiar to our discipline. The absence of clearly defined ophiolite sequences, except in Newfoundland, however, seems to present a big stumbling block to the Bird-Dewey model.

According to both models the Appalachian geosynclinal activity began with rifting of a single North-American-African continent late in the Precambrian and the line of rifting lay along the Appalachian eugeosynclinal zone. It will be hypothesized in Chapter XV that the St. Lawrence Rift system is a feature inherited largely from the initial rift system as suggested in Bird-Dewey and Chidester-Cadey models.

SUMMARY AND CONCLUSIONS

The boundary between Logan's Zone and the Piedmont is roughly coincident with a line of crustal change separating a two-layer 40-45 km thick crust to the southeast from a single layer, 30-35 km thick, crust to the northwest. This line of crustal change was also an important paleotectonic line. Before the Taconic orogeny, it was overlain by a major facies boundary between miogeosynclinal (on the northwest side) and eugeosynclinal basins. During the Taconic orogeny it became a zone of axial uplift. Grenville crystalline rocks form inliers in separated areas along this zone of axial uplift. One interpretation is that the edge of the Cambro-Ordovician North American continent lay just to the southeast of this line of Grenville inliers.

A feature common to recently proposed models of the Appalachian evolution is that the geosynclinal activity began with late Precambrian (Hadrynian) rifting of a single North American-African continent and that the line of rifting lay along the Appalachian eugeosynclinal zone.* It will be proposed later that faults related to this initial rift system acted as ancestral structures for the St. Lawrence Rift system.

*Bird and Dewey (1970, pp. 1043-1044) place the line of rifting within the Piedmont, just southeast of the Logan's Zone - Piedmont boundary whereas Chidester and Cady (1973, p. 4) place it further to the southwest in the eugeosynclinal zone.

CHAPTER III: THE SHIELD REGION

INTRODUCTION

The southeastern part of the Canadian Shield provides the setting for a major portion of the St. Lawrence Valley system (Fig. 5), and is referred to in this paper as the Shield Region. The northwestern limit of the Shield Region can conveniently be placed along the watershed of the Ottawa-St. Lawrence River system. Thus, except for the lobe-like extension of the Adirondack Mountains, the Shield Region comprises of a strip of territory approximately 400 km wide and extending from Georgian Bay (Lake Huron) northeastwards to the Atlantic coast, a distance of about 2000 km. Some aspects of its geological characteristics are briefly discussed below. Some physiographic characteristics of the region are described in Appendix II.

GEOLOGICAL CHARACTERISTICS

General

The Canadian Shield has been divided into seven structural provinces (Stockwell 1964; Idem 1968; pp. 692-698), each province consisting of a part of the Shield with a broad unity of structural trends and of isotopic ages. The provinces are separated from one another by uncomformities and/or structural-metamorphic fronts. The Shield Region includes almost the whole of the Grenville Province, and a small part of the Superior Province in the northern part of the Timiskaming Depression (Fig. 5).

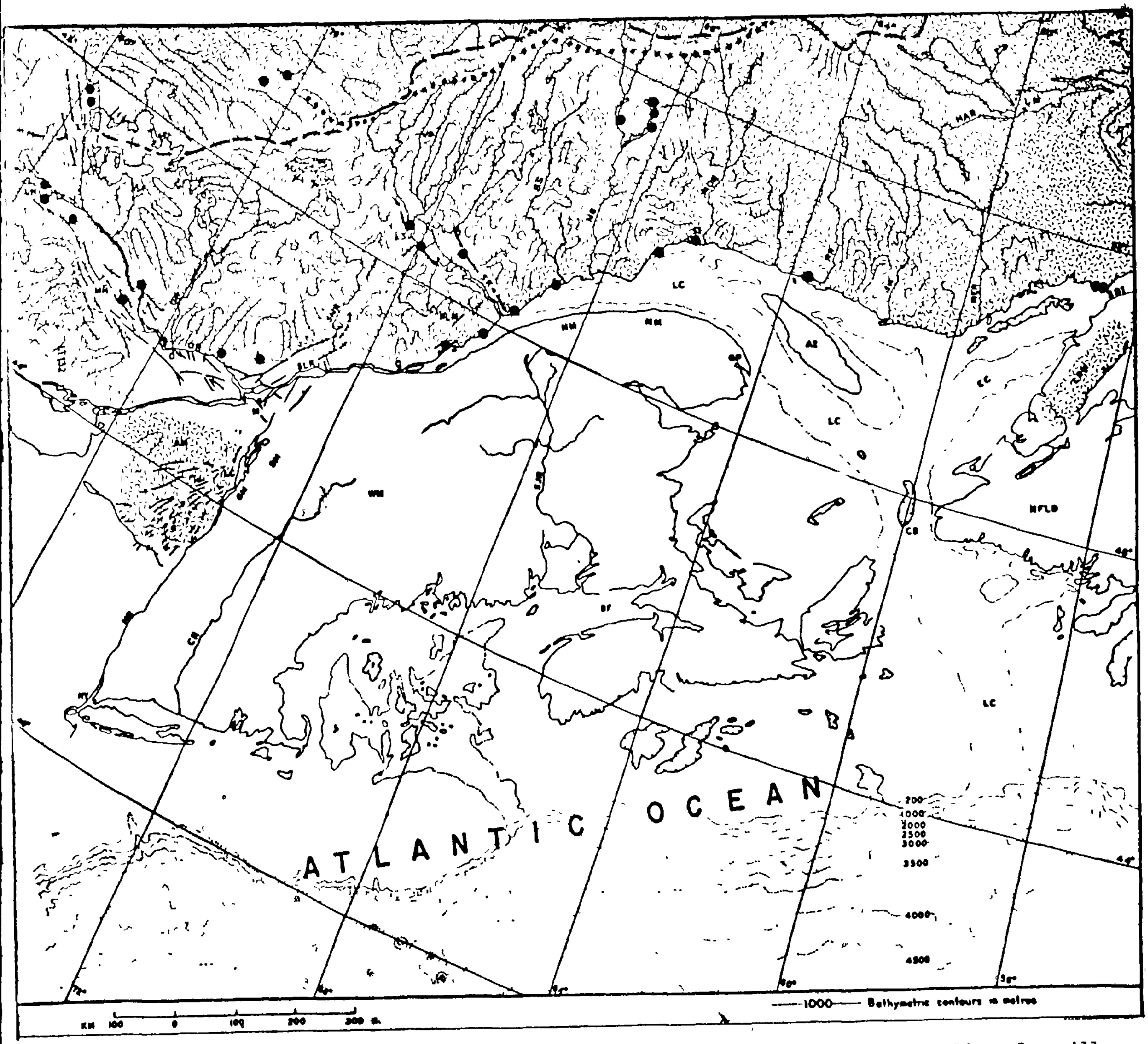


Fig.5. Southeastern part of the Canadian Shield showing selected features. Thick broken lines-Grenville Front. Heavy dotted line-Gravity low along the Grenville Front. Intermediate lines-High-angle faults. Fine lines-Generalized trends of gneissic foliation. Filled circles-Paleozoic outliers. Compiled from Geological and Tectonic maps of Canada (Can.Geol.Surv. Maps 1250A & 1251A) and other sources.

Both the Grenville and Superior Provinces possibly evolved through Precambrian orogenies. Isotopic age dating (K/Ar) has shown that the last orogeny in the Superior Province took place some 2300 to 2700 m.y. ago and this is referred to as the Kenoran orogeny. K/Ar values for the corresponding event in the Grenville Province - the Grenvillian orogeny - are much lower, ranging from 800 to 1100 m.y. (Helikian). Yet the Grenville Province is more deeply eroded to expose high-grade metamorphic rocks (upper amphibolite or granulite facies) formed in the catazonal environment, possibly at depths of 15 to 20 km (Wynne-Edwards 1972, p. 325), indicating that the area has since been greatly uplifted. Wynne-Edwards (1972, p. 322) thinks that erosion to the general level of the catazone was complete about 800 m.y. ago. At the present level of erosion, rocks consist mainly of a plutonic-gneissic complex contianing numerous bodies of granite, anorthosite and some distinctive metasedimentary rocks (including marbles, quartzites and pelitic metasediments) to which the name Grenville Supergroup has been assigned (Wynne-Edwards 1972, p. 290). The structural-metamorphic discontinuity that marks the northwest boundary of the Grenville Province is known as the Grenville Front. Wynne-Edwards (1972, Figure 4) has postulated that in general, the Grenville Front is a southeast dipping thrust fault. Northeast-trending structures predominate along the Grenville Front. Elsewhere, within the Grenville Province, such trends are not always immediately apparent; curved trends of great complexity are the most conspicuous (Wynne-Edwards 1969).

Compared with the Grenville Province, the Superior Province seems to expose a higher level of an orogenic edifice. In the problem area, the oldest rocks are Archean volcanic-sedimentary rocks that were probably deposited in a eugeosynclinal environment. They have dominantly east-west trends and have been intruded by numerous bodies of granite, and commonly metamorphosed to amphibolite and greenschist facies. In much of the problem area, they are overlain by a cratonic cover also of Precambrian age. Diabase dikes are common in the Grenville and the Superior Provinces and in some areas form dike swarms (Fahrig 1970).

High-Angle Faults

Several zones of high-angle faults of regional extent are known to affect the Shield Region. These faults are largely restricted to the steep edges of the region along the margins of the St. Lawrence Valley system depressions and are discussed in detail in Chapters IV through to XII, but for the present some of their main features are listed as follows: (i) they are high-angle faults (ii) it is impossible to determine precisely their age with the present data (iii) the most obvious effects of movement along them are (a) relative structural subsidence of crustal blocks along the St. Lawrence Valley system (b) uplift and tilting of crustal blocks in the areas of Madawaska Highlands, Adirondack massif and Parc des Laurentides massif (see Appendix II) which appear to be block mountains (iv) the faults have a tendency to conform to a regional lineament pattern (discussed below) oriented prevalently NE and NW.

Lineament Pattern

Numerous lineaments in the form of narrow rectilinear or zig-zag valleys are present in the Shield Region and range in length from a few kilometers to several tens of kilometers (Fig. 6). These can be clearly seen in vertical airphotos. Only a small portion of these lineaments can be considered as reflecting bedding and foliation directions, for many of them cut across these trends, and therefore seem to be fracture (fault and/or joint) controlled.

The lineaments in the Shield Region produce a remarkably persistent pattern. They prevalently fall into two sets, one that trends approximately northeast and the other approximately northwest (Fig. 7). That they reflect a regional fracture pattern is also indicated by the fact that these directions coincide with two of the three dominant directions of diabase dikes in the Shield (Lespérance 1948, p. 10).

The faults shown in Figure 6 conform or tend to conform to the regional lineament pattern. In the discussions of high angle faults in Chapter IV through to XII it will be shown that the tendency of fault directions to conform with the regional lineament pattern is a characteristic feature of many areas of the St. Lawrence Valley system. This might imply that the faults have developed along the trends of earlier fractures or that the fracture pattern is a product

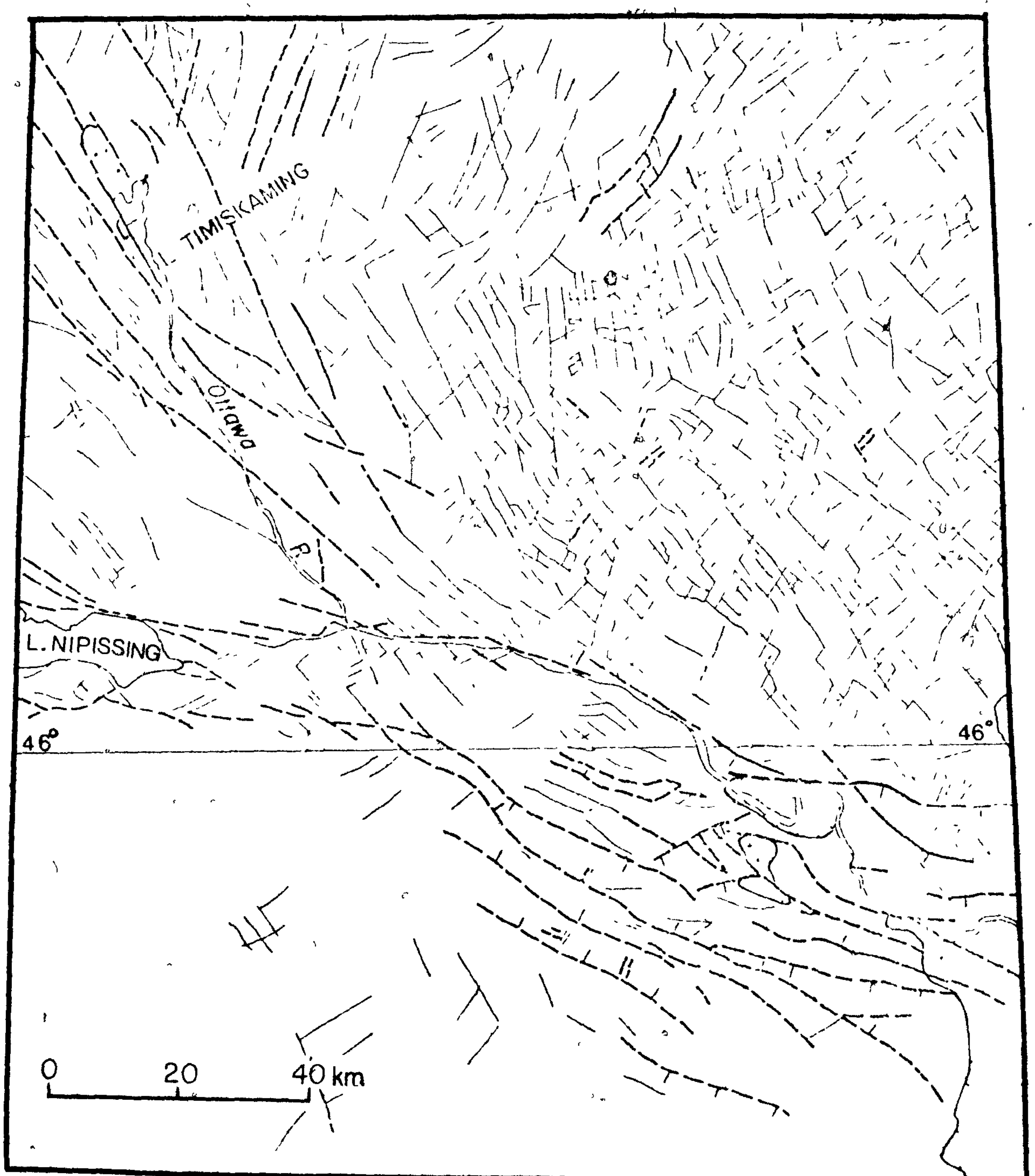


Fig.6. Airphoto lineaments in a part of the Shield Region. Thin lines-Airphoto lineaments. Thick dashed lines-Mapped and interpreted high-angle faults, after Kay 1942, Lovell and Cain 1970, Lumbers 1971. Stipples-Areas of Early Paleozoic platformal rocks.

of the same stresses that produced the faults. The faults are largely restricted to the steep edges of the Shield Region along the St. Lawrence Valley system whereas the lineament pattern appears to be present over much of the Shield Region. Moreover, as lineaments do not appear to have a greater frequency in proximity to the faults, it is thought that the fault trends are a result rather than a cause of the general lineament pattern. This lineament pattern will in future be referred to as the "regional fracture pattern" of the Shield Region.

Deep Drainage Lines and their Possible Structural Significance

Drainage characteristics of the Shield Region are described in Appendix II. In addition to the poorly-established drainage lines there are deeply entrenched river valleys which attain depths of 300 m or more but even the deepest of them is filled to a greater or lesser extent with glacial debris. These deep drainage lines are transverse to the Shield Region and their rivers in the upper courses flow against the regional slope. They commonly have zig-zag trends the linear segments being parallel to the "regional fracture pattern" (see above) and hence they appear to be controlled by joints and/or faults. It will be hypothesized later (see Chapter XIV) that these deeply entrenched valleys lie along tension fractures which developed in the Shield Region, as part of events that led to the formation of ancestral structures related to the St. Lawrence Rift system.

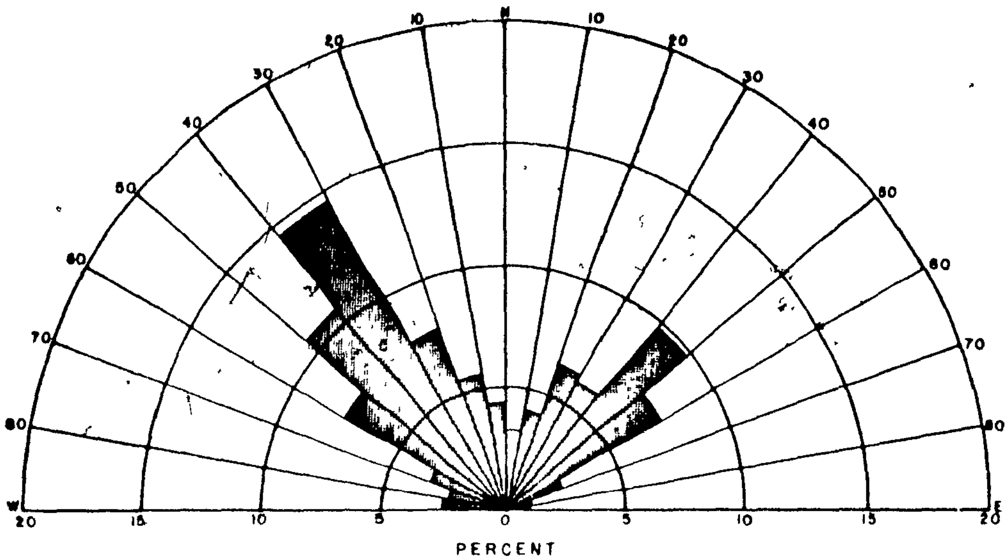


Fig.7. Strike frequency diagram of lineaments shown in the part north of latitude 46°in Fig.6.

SUMMARY AND CONCLUSIONS

The Grenville Province, that makes up the greater part of the Shield Region, has been strongly uplifted and deeply eroded to expose catazonal metamorphic rocks, possibly formed at depths of 15 to 20 km. Erosion to the general level of the catazone may have been complete about 800 m.y. ago.

The surface of the region is characterized by a prevalently northeast and northwest oriented lineament pattern which probably reflects a regional fracture pattern. It is thought that this fracture pattern controlled the directions of faulting and also the courses of the deeply entrenched river valleys of the region.

High-angle faults of the Shield Region are largely confined to the margins of the St. Lawrence Valley system depressions. Movements on these faults have led to relative structural subsidence of the St. Lawrence Valley system and to uplift of Madawaska Highlands, Adirondack Massif and Parc des Laurentides massif areas to form block mountains.

CHAPTER IV: MARGINAL SEGMENT:

THE ST. LAWRENCE VALLEY

INTRODUCTION

The term "St. Lawrence Valley" refers to the northeasterly trending linear depression whose plain-like floor straddles the St. Lawrence River, from its confluence with the Ottawa River to where it is joined by the Saguenay River (Fig. 8). As so defined, the St. Lawrence Valley is about 450 km long and varies in width from about 25 km to about 100 km. This depression is a part of the Marginal Segment of the St. Lawrence Valley system. It is

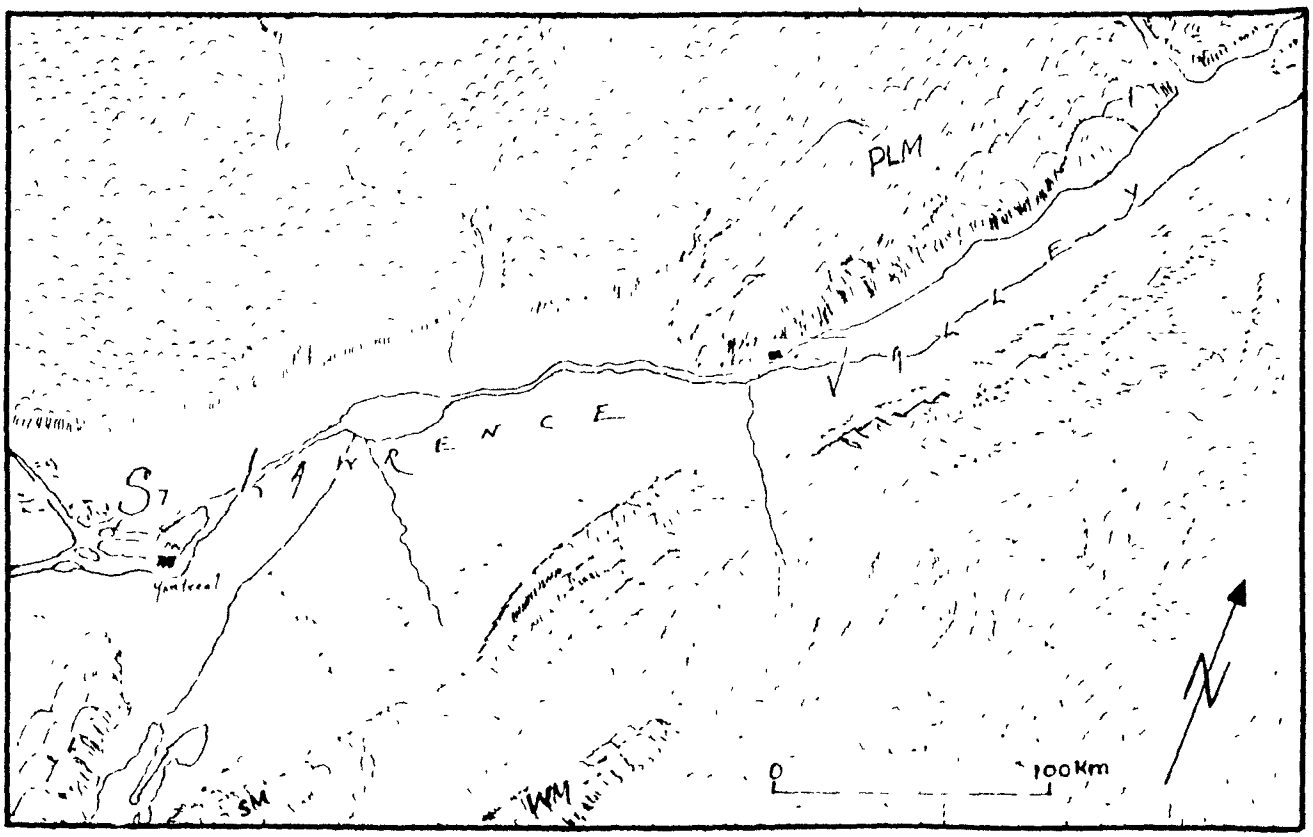


Fig.8. Physiographic sketch diagram of the St.Lawrence Valley area. Key to to abbreviations: PLM-Parc des Laurentides massif; SM-Sutton Mountains; WM-White Mountains.

bounded on the northwest by the Laurentian Highlands and on the southeast by the highlands of the Appalachian Region and is largely cut into a strip of relatively soft (higher erodibility) rocks that lies between the Shield and the Appalachian foldbelt.

Between Quebec City and the Saguenay outlet the valley is a moderately deep trench, 25 to 30 km wide and is largely occupied by the St. Lawrence River. In this section, the valley floor lies about 600 m below the adjacent areas of the Laurentian Highlands (Parc des Laurentides massif) and about 300 m below the bordering areas of the Appalachian Region. The rest of the St. Lawrence Valley is shallower, wider and has a plain-like floor. The bordering areas of the adjacent highlands rarely rise to more than 300 m above the valley floor. Southwest of Quebec City, the valley widens to about 50 km in a distance of about 75 km through several step-like egressions of the north boundary. Further southwestwards, it continues to widen gradually, to attain a maximum width of about 100 km at the south end (Fig. 8).

The St. Lawrence Valley of this thesis has sometimes been called the St. Lawrence River valley. This latter term is inappropriate because there is no evidence that the topographic low was carved out by the St. Lawrence River, although the river has undoubtedly modified the valley floor to some extent. Also, some workers refer to the feature as the St. Lawrence Lowlands, but this term is generally applied to the plain-like lowlands coincident with the present St. Lawrence Platform (see Bostock 1969). Thus, the St. Lawrence Lowlands include a major portion of the St. Lawrence Valley, but the term lowlands fails to convey the actual form of the feature which is partly a shallow flat-bottomed depression and partly a well-defined trench.

GEOLOGICAL CHARACTERISTICS

The St. Lawrence Valley straddles the boundary between the St. Lawrence Platform and Logan's Zone of the Appalachian foldbelt, although in the section southwest of Quebec City, the valley is mostly on the platform. Along its northwest margin Grenville rocks emerge from beneath the platform cover and form the north wall of the valley.

General Geology and some Paleogeographic Implications

The strip of Appalachian rocks on the southeast side of the valley is 10 to 25 km wide and consists of flyschoid allochthonous sequences of nappes, thrust slices and klippen of Logan's Zone. The rest of the valley is largely underlain by platformal rocks of Cambro-Ordovician age. Between Quebec City and the Saguenay outlet these rocks are exposed in patches along the northwest margin of the valley but may be continuous along the channel of the St. Lawrence River.

The platformal succession ranges in thickness from 0 to about 2600 m (Hofmann 1972) and overlies Grenville crystalline rocks. A

Grenville inlier of about 115 km^2 in area occurs immediately north of the confluence of the Ottawa and St. Lawrence Rivers and forms a hilly area called Oka Mountain. The platformal succession begins with the Potsdam group which as far as is known is restricted to the part of the valley southwest of the St. Maurice River. The basal unit of this group, the Covey Hill formation is believed mainly to represent alluvial plain deposits (Lewis 1971, p. 875) and is typically composed of unfossiliferous coarse reddish and greyish felspathic sandstones, quartz sandstones and conglomeratic sandstones, ranging in thickness from 0 to probably as much as 600 m (Hofmann 1972, p. 4; Clark 1972, pp. 19-34), the maximum known thickness being in the southernmost parts of the valley (see Clark 1972, p. 12). The Covey Hill formation is disconformably overlain and overlapped by the Chateauguay formation, the lower part of which is a pure quartz blanket - a light-coloured, well-sorted, quartz sandstone about 250 m thick, with low content of feldspar and containing some marine fossils - deposited by transgressing seas in a subtidal, shallow water environment. A late Cambrian age has been assigned to the Chateauguay formation, but the older Covey Hill may be in part Precambrian (Hofmann 1972, p. 4). Lewis (1971, p. 873) has suggested that the depositional basin in which great thicknesses of Covey Hill sediments accumulated was a fault-bounded "gravity sag" or "half graben" with faults along what is now the northwest margin of the St. Lawrence Valley. The upper part of the Chateauguay formation contains dolomitic sandstone formed in an environment transitional to the stable shelf environment of carbonate deposition, which became established over the entire area and lasted through the Early Ordovician and a part of the Middle Ordovician (Trenton). Then, in the late Middle Ordovician, possibly related in cause to the uplift of the Taconic Mountains, and to crustal subsidence and emplacement of klippen in Logan's Zone (Zen 1972, p. 49), the area subsided and received sediments of black shale facies (Utica), followed in the late Ordovician times by coarser clastics (Queenston), apparently derived from the destruction of Taconic Mountains. There is no definite record of Silurian rocks in the area, but on St. Helen's Island, near Montreal, a diatreme breccia presumably related to Monteregian igneous activity (to be described later), contains blocks of Lower Devonian limestone (Logan 1863) indicating that the original platformal sequence contained rocks as young as Early Devonian.

Two systems of crustal dislocations, of regional extent and of contrasting character, occur in close association in the St. Lawrence Valley area. One of these consists of dislocations collectively referred to as Logan's Line (Fig. 9). Structures comprising Logan's Line are believed to be southeast-dipping, low-angle thrusts representing the western limit of allochthonous sequences of Logan's Zone (see St. Julien 1972, p. 1). The other is a system of high-

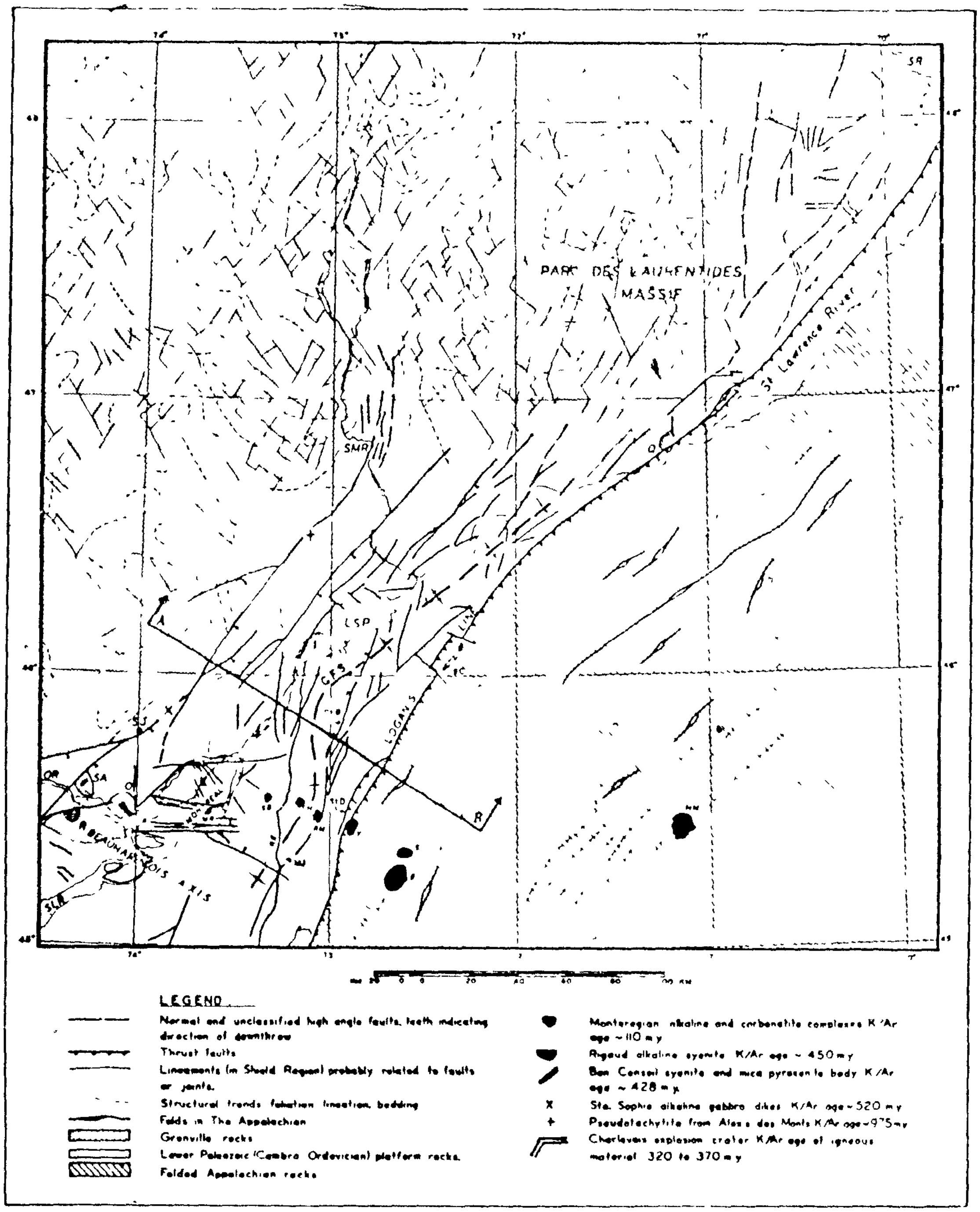


Fig. 9. Map showing selected features of the St. Lawrence Valley. Data from Houde and Clark [illegible], Rondot [illegible], Clark [illegible] and other sources. Lineaments of the Shield Region from vertical airphoto study. Map compiled by P.S. Kumarapeli. Key to abbreviations: B-Brome, BC-Bon Conseil; Ch-Chambly-Fortierville syncline; LSP-Lac St. Pierre; MM-Mount Megantic; MR-Mount Royal; O-Oka Mountain; OR-Ottawa River; R-Rigaud Mountain; RM-Rougemont; RR-Richelieu River; S-Shefford; SA-St. André; SB-St. Bruno; SH-St. Hilaire; SJ-St. Jérome; SLR-St. Lawrence River; SMR-St. Maurice River; SR-Saguenay River; St.B-St. Barnabé Fault; St.D-St. Dominique Fault; Y-Yamaska.

angle faults that affects much of the valley area and the shield margin to the northwest (Fig. 9).

The platformal rocks of the valley although much broken by high-angle faults have a relatively simple structure and are largely autochthonous. An exception is a belt of intensely crumpled, presumably parautochthonous rocks 5 to 10 km wide, on the foot-wall side of Logan's Line (this is the St. Germain Complex of Clark 1956, p. 4). Elsewhere, the platformal strata are horizontal or nearly so. Dips rarely exceed 10° and are generally no more than 2 or 3 degrees. The prism of platformal rocks has the gross structure of a doubly plunging syncline - the Chambly-Fortierville syncline. The axis of this syncline lies closer to Logan's Line than to the shield margin (Fig. 9).

Products of a Cretaceous igneous event, referred to as the Monteregian igneous activity (Gold 1968, pp. 288-302) occur in the southern part of the valley. They consist of stocks, dikes, sills and plugs of alkaline igneous rocks and carbonatite. Igneous rocks which are petrochemically similar to these, occur typically in the environment of continental rifting (Heinrich 1966, pp. 24-27). Accordingly, these igneous rocks and also the fault systems in the area are of particular interest to the graben problem of the thesis and are described below in some detail.

<u>Monteregian Rocks</u>

Monteregian Rocks are a closely related group of alkaline igneous rocks. Detailed accounts of these rocks are found in the works of (Adams 1903, pp. 239-282), Graham (1944, pp. 455-482), Gold (1968 pp. 288-302) and Hodgson (1968).

The largest concentration of Monteregian igneous bodies occurs in a well-defined zone about 45 km wide, extending approximately S75^{O}E from St. André for a distance of about 225 km (Fig. 9). This zone can be regarded as the Monteregian Petrographic Province "proper" Outside the province " proper" probably comagmatic rocks occur in Mt. Megantic and vicinity (Osborne 1935 in Graham 1944, p. 459; Lowden 1961; Fairbairn et al. 1963), and in the Champlain Valley region (Zartman et al. 1967). The following brief discussion applies to the igneous rocks of the Monteregian Province "proper" and of Mt. Megantic area. Rocks in the Champlain Valley region will be discussed later (p. 71).

Including Mt. Megantic, eleven main centres of intrusions are exposed (Figs. 9 & 10). A twelfth is only slightly exposed but its presence has been demonstrated by geophysical means (Kumarapeli et al. 1968) and diamond drilling. There is geophysical evidence that other intrusive centres, not yet exposed by erosion, are present in the area (e.g. see Philpotts 1970, p. 400). The main plutons are crudely circular in plan with areas ranging from about 0.5km^2 to over 75 km^2. They are believed to be cylindrical or funnel-shaped bodies. Genetically related dikes, sills and small plugs intrude the province intensively (Fig. 10).

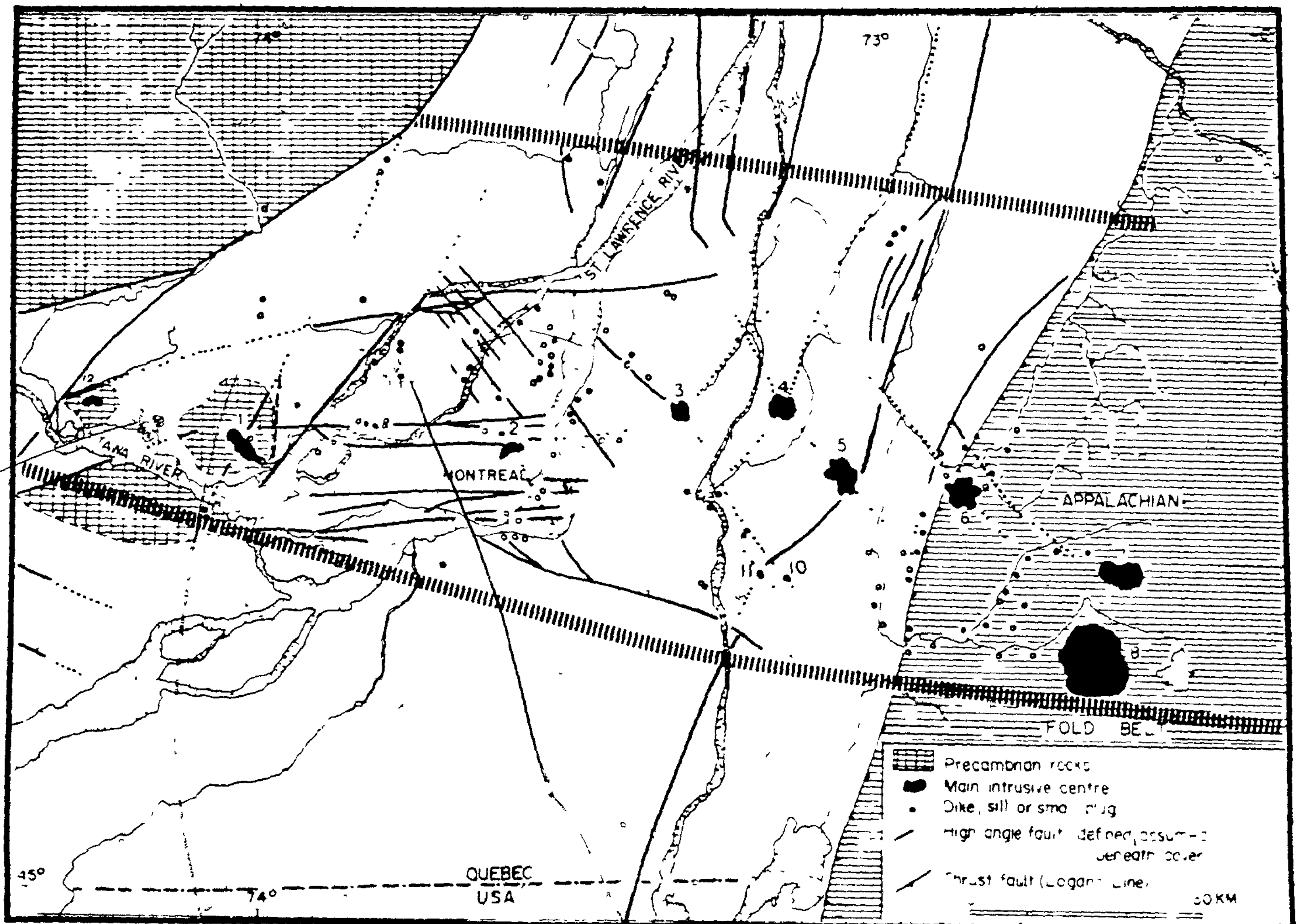


Fig.10. Map showing the igneous body distribution of the Monteregian Province "proper" and the possible relations of the main intrusive centres to high-angle faults. Igneous body distribution from Hodgson 1968 and other sources. Main intrusive centres: 1-Oka, 2-Mt.Royal, 3-St.Bruno, 4-St.Hilaire, 5-Rougemont, 6-Yamaska, 7-Shefford, 8-Brome, 10-Mt.Johnson, 11-Grand Bois, 12-St.André. Wide broken lines-possible north and south boundaries of the Monteregian Province "proper".

The most unifying petrochemical character of the Monteregian igneous rocks is that they are strongly alkaline. Their alkali-lime indices (see Peacock 1931), as far as have been determined, fall well below the 51% level (Gold 1968, p. 293), which separates the alkaline from alkali-calcic rocks. Despite this unifying character, Monteregian rocks differ widely, chemically as well as mineralogically. In the main complexes, for instance, the initial phase is usually ultrabasic or basic and later phases are more silicic. At one extreme are peridotites, pyroxenites and titanaugite gabbro and at the other extreme are strongly alkaline syenitic rocks, the two being apparently linked through normal gabbro and essexite, monzonite and syenodiorite (Gold 1968, p. 290). The above variations seem to reflect trends of magmatic differentiation, probably in local chambers.

Monteregian rocks also show variations of a regional character. For instance, westward along the province there is a progressive decrease in SiO_2 (Hodgson 1968, p.102). The easternmost pluton in Mt. Megantic has a considerable amount of granitic rock associated with it (Reid 1961) whereas the westernmost plutons, the Oka and St. Andre (see Gleeson and Cormier 1971) complexes are largely made up of carbonatite and other extremely undersaturated rock types. Fitzpatrick (1953) states that gravity highs as large as 19 milligals are associated with individual plutons but the size of the anomaly depends on the proportion of mafic material and the size of the plutons. The regional gravity high associated with them parallels the trend of the hills and decreases in amplitude to the east, the direction in which the plutons also become decreasingly mafic. Another variation of a regional character is the westward increase of K_2O/Na_2O (Hodgson 1968, p.102). Potassic ultrabasic rocks such as alnöite and kimberlite (Marchand 1968; Gold and Marchand 1969) are restricted to the western part of the province, west of the Island of Montreal. This part of the province is also characterized by numerous diatreme breccia pipes, (Gold et al. 1972), probably representing a gas-rich pipe drilling phase. Hodgson (1968, p. 131) suggests that the probable explanation for the westward variations of Monteregian rocks is that, in this direction, magmas were derived from increasingly greater depth within the upper mantle.

Isotopic ages of Monteregian rocks vary from about 90 m.y. to about 130 m.y., (Lowden 1960, 1961; Hurley et al. 1959; Fairbairn et al. 1963; Shafiqullah et al. 1970; also see Clark 1972, p. 153) indicating that the magmatic event took place in the Cretaceous. Results of paleomagnetic studies (Larochelle 1959, 1968 and 1969) and of isotopic age dating are mutually consistent.

Regional fault systems: Logan's Line

Except for a few minor offsets, Logan's Line is shown on most small scale geological maps as a continuous thrust fault that extends from the east side of Lake Champlain (Missisquoi Bay) to the vicinity of Quebec City and beyond (Fig. 9). However, outcrops indicating its position and nature are exposed mainly in the Lake Champlain and Quebec City areas. In the former area the fault is well exposed on the east shore of Missisquoi Bay. Cambro-Ordovician rocks (Philipsburg series) have been thrust westwards over Middle Ordovician (Trenton) shales along a fault plane that dips about 20° eastwards (Clark 1951, p. 20). In the Quebec City area, however, dips of faults are steeper, ranging between 40° to 70°, but are again towards the Appalachians (Ells 1888, p. 15k; Raymond 1913; Graham and Jones 1931; Riva 1972). There too, Cambro-Ordovician rocks have been thrust westwards over Middle Ordovician rocks. In between the two above localities which are about 200 km apart, Logan's Line is placed to explain stratigraphic anomalies and wide brecciated zones and its exact nature and position are in doubt (Clark 1964a, p. 75).

Kumarapeli and Saull (1966a, p. 665) discussed the nature of Logan's Line in the context of possible graben subsidence along the St. Lawrence Valley and emphasized the problematic nature of structures included in Logan's Line. They suggested the possibility that the scale of thrusting postulated by some of the earlier workers may be exaggerated, implying that the thrust faults may steepen downwards and extend into the Precambrian basement i.e. Logan's Line is essentially a structure involving "thick-skinned" tectonics. In order to test this hypothesis, the writer used gravity traverses to study the inferred trace of Logan's Line on the southeast side of the St. Lawrence Valley. The studies included ten gravity traverses across this trace. The gravity work was planned on the following assumptions: (i) the subsurface structure of Logan's Line includes a basement step, the downdrop on which is towards the St. Lawrence Valley (ii) there is a significantly high density contrast between the basement and cover rocks. Contrary to the assumption (i) above, concensus has grown, overwhelmingly in recent years, in support of possibly gravity driven "thin-skinned" thrust sheets on Logan's Line (Williams 1972b, pp. 192-194; Zen 1972, p. 13). In view of this, the residual anomalies over Logan's Line have been interpreted in this thesis, in terms of density differences of rock masses on either side of an east-dipping thrust fault, rather than in terms of a basement step (Appendix III).

Regional Fault Systems: High-Angle Faults

The presence of high-angle faults of the St. Lawrence Valley has been inferred largely from geological mapping (Béland 1961; Clark 1952,1955, 1964a, b & c, 1966, 1972; Osborne and Clark 1960; Rondot 1966, 1969; also see Dufresne 1948; Houde and Clark 1961, Osborne 1956) and from geophysical studies (Hosain 1965; McDonald 1965; Roliff 1968; Sharma 1968; Kumarapeli and Sharma 1969; Frey 1973). Over the years, more and more faults have been interpreted to give the braided fault pattern shown in Figure 9. Opportunities for direct examination of these faults are rare because the bedrock over much of the valley area is concealed by unconsolidated Pleistocene sediments.

The faults are believed to be mostly of the normal type because:

1. minor faults exposed in road cuts and other embankments are normal faults
2. the faults outline a down faulted terrain indicating that the movements were controlled by gravity
3, there is no evidence for large-scale strike-slip movements.

The writer examined an E-W trending fault (Cheval Blanc Fault? see Clark 1972, p. 162) which was intersected by the north-south line of the Montreal subway tunnel system; it consists of a composite zone of normal faults about 12 m wide and dipping at 70°S.

It is, however, possible that some of the high-angle faults in the eastern part of the valley close to Logan's Line are reverse faults. For instance, the fault on the east side of Rougemont (Fig. 9) is an east dipping high-angle reverse fault (Philpotts 1972, p. 15). What appears to be the northward continuation of the fault is St. Barnabé fault (Fig. 9) which also may be an east-dipping high-angle reverse fault rather than a west-dipping normal fault (Kumarapeli and Saull 1966a, p. 645). Thus, it appears that in the southeast side of the valley, normal faults, high-angle reverse faults and thrust faults occur in close association.

The high-angle faults of the St. Lawrence Valley can be grouped into three sets according to their strike trends:

1. a principle set of northeast striking faults; a few of the faults included in this set strike north-northeast
2. an E-W set
3. a northwest set.

Another way of classifying these faults is:

1. longitudinal faults: the northeast striking faults are nearly parallel to the valley trend
2. transverse faults: the northwest and E-W trending faults make large angles with the valley trend.

The E-W set of faults is best known from the southern part of the valley, and they appear to represent the eastward continuation of longitudinal high-angle faults of the Lower Ottawa Valley (see Fig. 21). Movements along them appear to have produced a slight crustal sag in the southern part of the St. Lawrence Valley (see Clark 1972, Fig. 18).

Faults striking northwest are known mainly from the Montreal and Lac St. Pierre areas. Compared with other high-angle faults, the northwest-trending faults appear to be relatively short. The northeast and northwest-trending sets of faults conform with the principal directions of the "regional fracture pattern" of the Shield Region (Fig. 9) indicating, as discussed earlier (see p.40-42), that the development of these faults may have been controlled by an older fracture pattern. In the southern part of the valley all three sets of faults are present, they produce an approximately E-W trending zone of intense block faulting (Fig. 10). This area of block-faulting has been the site of much of the Monteregian igneous activity. The main intrusive centres are probably located at the loci of intersection of faults as illustrated in Figure 10.

Ages of Faults

There is no general agreement on the age of Logan's Line. Most workers think that it formed as an integral part of the Taconic orogeny (e.g. see Zen 1972, p. 13). Others think that the available data do not provide an unequivocal answer (e.g. see Cady 1969, pp. 63-64, 161-162) and that most, if not all, movements along it may have taken place in Late Devonian, during the Acadian orogeny (Poole 1967, pp. 34-35). Recently Zen (1972) has discussed the nature of Logan's Line and related structures; he thinks that the thrusting events can be dated, and assigns a late Early or Middle Ordovician age to them.

A simple answer also does not seem to exist to the age question of the high-angle faults. As discussed below, the indications are that the faults as they now appear contain a sequence of movements impressed on them since their formation sometime before the platform cover began to form in Cambrian or possibly in late Precambrian times. Any attempt to answer this question has to take the following lines of evidence and their implications into account.

1. The lithological and sedimentalogical characteristics of the Covey Hill formation (Cambrian and possibly partly Precambrian) indicate that the Covey Hill

sediments accumulated in a subsiding fault trough (asymmetrical graben?) coextensive with the southern part of the St. Lawrence Valley (Lewis 1971, p. 873). In the southwestern part of the valley, the Potsdam is considerably thick even at the valley margin (Clark 1952, p. 17; Kumarapeli and Sharma 1969) but no Potsdam is found on the shield surface. Yet an outlier of residual limestone west of Echo Lake (about 15 km west of the valley margin; see McGerrigle 1939, p. 43) lies on the Grenville surface such that its position with respect to its correlatives within the valley appears to be in normal stratigraphic position (Osborne 1937, p. 8). These observations suggest that a step, probably representing a fault or fault-line scarp, existed along the present valley margin when the Potsdam began to be deposited and that the fault was active during the accumulation of the Covey Hill sands.

2. Just northeast of Lac St. Pierre, roughly in line with the channel of the St. Maurice River, drilling for oil has indicated a pre-Middle Ordovician basement step of nearly 400 m amplitude (Belyea 1952, p. 31), which can best be explained as a feature due to faulting.

3. At St. Alexis des Monts (for location see Fig. 9), near the northwest margin of the valley, a glassy pseudo-tachylite from one of the faults that cut Grenville

rocks has yielded a K/Ar age of 975 ± 50 m. y. (Philpotts and Miller 1963), although what appears to be the continuation of this same fault, displaces Ordovician rocks within the valley.

4. In the St. Sophie area (for location see Fig. 9), an alkaline gabbro dike swarm perhaps related to faulting along the valley margin, has been K/Ar age dated at 520 ± 2 m. y. (Doig and Barton Jr. 1968, p. 1403).
5. Subsidence of the carbonate shelf (and the establishment of deeper water conditions) in late Middle Ordovician times, may have been achieved by movements on high-angle faults (Zen, 1972, p. 49).
6. Seismic reflection data apparently indicate that at least three periods of early Paleozoic normal faulting took place in the St. Lawrence Valley and that thrusting on Logan's Line took place after the formation of large normal faults (Frey, 1973). The pre-Logan's Line high-angle faulting, however, could be pre-Middle Ordovician or pre-Late Devonian, depending on which age is assigned to Logan's Line.
7. High-angle faults that affect the Valley floor cut and displace the Cambro-Ordovician platformal sequence (Houde and Clark 1961), including the youngest Upper Ordovician rocks (Becancour River) present. In a locality in

Quebec City high-angle faults cut thrust faults of the parautochtonous sequence (J. Riva quoted by E.W. Mountjoy).

8. There is a good space correlation between the area of block-faulting in the southern part of the St. Lawrence Valley and the distribution of Monteregian igneous bodies of the western half of the Monteregian Province "proper" (Fig. 10). The eastern half also lies along the possible eastward continuation of this zone of block-faulting. These space relations suggests that the Cretaceous-magmatic event and block-faulting may also be chronologically related (Kumarapeli 1970). Such an inference seems reasonable in view of the common space and time association, in continental shield and platform areas, of alkaline magmatism with zones of major faulting (McCall 1959; King & Sutherland 1960; Ginzburg 1962; Bailey 1961, 1964).

9. The St. Lawrence Valley area is characterized by mild earthquake activity (Smith 1967) which can be attributed to movements along some of these faults at the present time (Kumarapeli and Saull 1966a, p. 646).

The main conclusions consistent with the above evidence and lines of reasoning are as follows.

1. The St. Lawrence Valley was an area of high-angle fault movements in post-Ordovician times. The greater part of these movements may have taken place in the mid-Mesozoic synchronously with the Monteregian magmatism.
2. Before the platformal sequence began to form in Cambrian or possibly in late Precambrian times, the present St. Lawrence Valley area, at least in part, had a fault-lined structural framework. Also there are indications that some high-angle fault movements took place during the deposition of the platform cover. These movements and also the post-Ordovician movements possibly represent renewed movements on older faults.
3. At least some of the faults may be in an unhealed state at the present time.

The foregoing discussion gives some insight into the chronology of high-angle faulting in the St. Lawrence Valley. Even a pretence of attempting a complete answer to the age question is unrealistic, mainly because, excepting Pleistocene deposits, stratified rocks younger than Ordovician are almost completely absent in the area. This should be kept in mind when speculating whether a major part of the fault movements took place during a particular time period or in several discrete time periods.

Regional Fault Systems and the Tectonic Framework

If the thrust faults of Logan's Line are, in fact, structures involving "thin-skinned" tectonics (p.55), then it is considered likely that these faults became inactive soon after their formation during the Taconic and/or Acadian orogenies. At least some high-angle faults on the other hand, have apparently remained active and have dominated the tectonic framework of the valley area during the entire Phanerozoic time and possibly even earlier. In places the high-angle faults seem to have involved the entire crust and upper mantle, as indicated by the close spatial association of some faults with Monteregian alkaline magmatism.

Structural Effects of High-angle Faulting

Interpretation of the regional structures produced by high-angle faults is complicated by certain unsolved problems. In the area of Logan's Line, for instance, dislocations of both tensional and compressional origins (normal faults, east dipping high-angle and low-angle reverse faults) occur in close association. The interrelations of these faults are largely unknown, but the compressional dislocations are best viewed as structural elements created by orogeny (Appalachian). Elsewhere in the valley, faulting appears to be less complex in that only high-angle faults are present and these faults are probably of the normal type. They combine to form an elongate zone of downfaulting i.e. a rift zone. The structural subsidence is largely controlled by the northwest marginal fault zone and the high-angle faults to the southeast of the zone may be synthetic and/or antithetic in nature. The overall structure is not unlike that of the Hessische Graben belonging to the Saxonic faulted area (see De Sitter 1964, p. 126). In the southern part of the valley where the rift zone appears to be widest (∿55 km) it consists of a series of longitudinal fault blocks downdropped to the east in step-like fashion (Fig. 11).

Between the south end of the valley and Quebec City, the longitudinal shape of the rift zone is cradle-like. This shape has been achieved, at least in part, by movements on transverse faults. The rise of the basement towards the south end of the valley culminates in an arch-like structure - the Beauharnois axis - that extends across the south end of the valley (Fig. 9). The structurally highest part of this axis seems to be at its west end, where the Grenville outliers of Oka Mountain (and also of Rigaud Mountain and of St. André, in the nearby areas of the Ottawa Valley) are located.

Following the line of earlier discussions on the ages of faulting, there is conclusive evidence that the rift zone formed, at least in part, in post-Ordovician times, after the deposition of the present platform cover, although it may have developed partly during the deposition of the cover rocks as well. The post-Ordovician structural subsidence may have taken place largely in the mid-Mesozoic, synchronously with the Monteregian alkaline magmatism (Kumarapeli 1970). However, an ancestral structure appears to have existed, possibly as a zone of high-angle faulting, before the platform cover began to form.

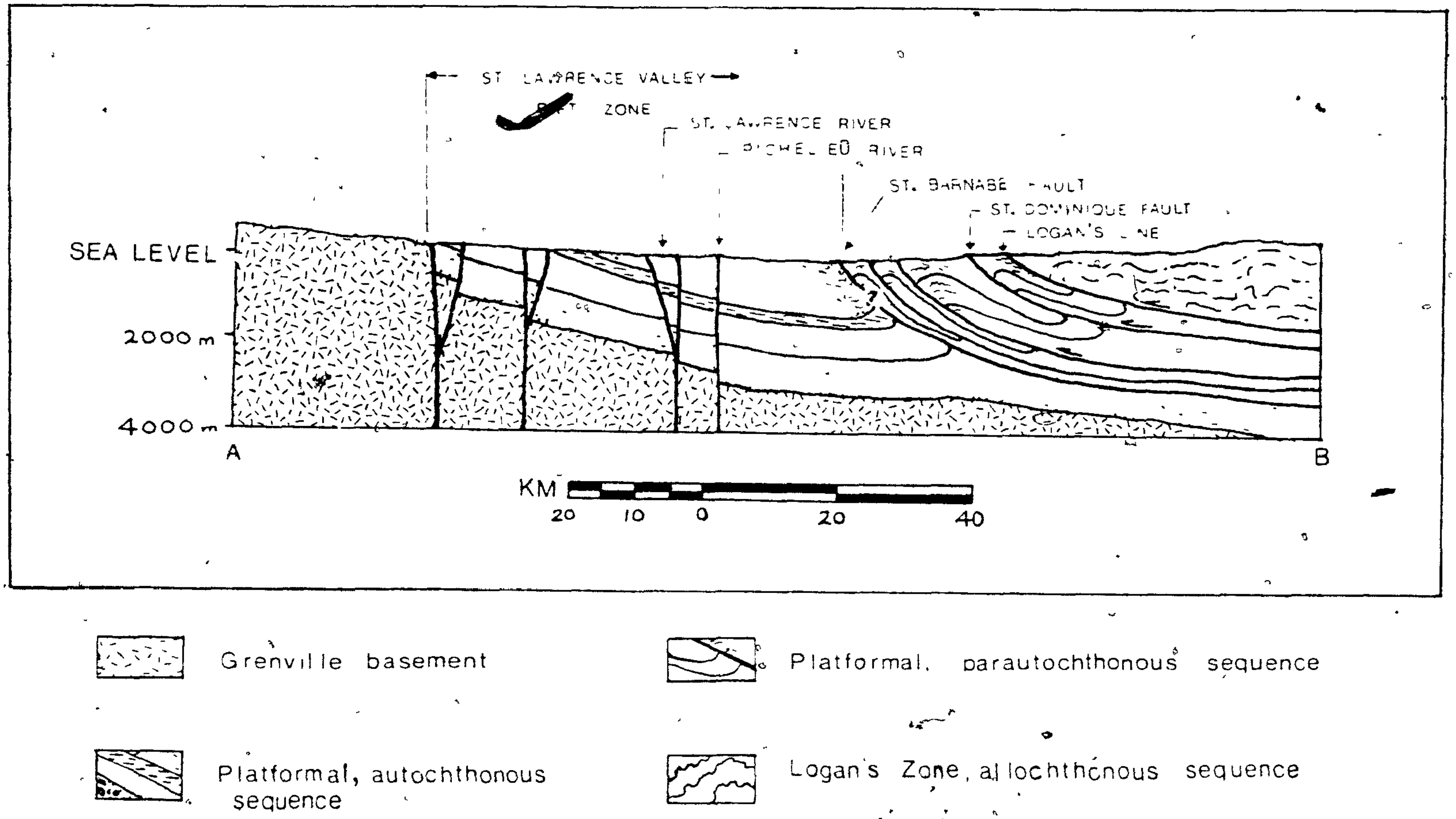


Fig. 11. Possible structure section across the St. Lawrence Valley. For section location see Fig. 9.

SUMMARY AND CONCLUSIONS

In the St. Lawrence Valley, the part that is underlain by autochthonous platformal rocks, is a zone of block-faulting consisting of a major set of longitudinal faults and a minor set of transverse faults. Play of movements on these faults has produced a linear zone of structural subsidence which at least in the southern part of the valley has a graben-like cross section. Between the southwest end of the valley and Quebec City, the downfaulted zone is longitudinally cradle-like. The rise of the basement towards the southwest end of the valley culminates in a transverse arch (Beauharnois axis), the north flank of which is severely block-faulted. This block-faulted area has been largely the site of a major episode of alkaline-carbonatite magmatism (Monteregian) in the early Cretaceous.

There is conclusive evidence that structural subsidence along the valley took place in post-Ordovician times. Much of this subsidence may have taken place synchronously with the early Cretaceous alkaline-carbonatite activity. There are strong indications that the post-Ordovician movements represent, at least in part, renewed movements on older faults which appear to have been in existence during the early development of the Appalachian geosyncline.

CHAPTER V: MARGINAL SEGMENT: THE CHAMPLAIN VALLEY

INTRODUCTION

The topographic low of the St. Lawrence Valley divides at its south end into two branches that follow the north and east sides of Adirondack Mountains. The branch on the east side of the Adirondacks is largely occupied by Lake Champlain and is referred to as the Champlain Valley (Fig. 12). The Champlain Valley extends northwards from the divide between the Hudson River valley and the Lake Champlain basin, approximately for a distance of 175 km where it merges with the St. Lawrence Valley. In outline, the valley is trumpet-shaped; it is about 10 km wide at the south end and widens northwards gradually, except at its north end (Fig. 12) where it opens out to attain a width

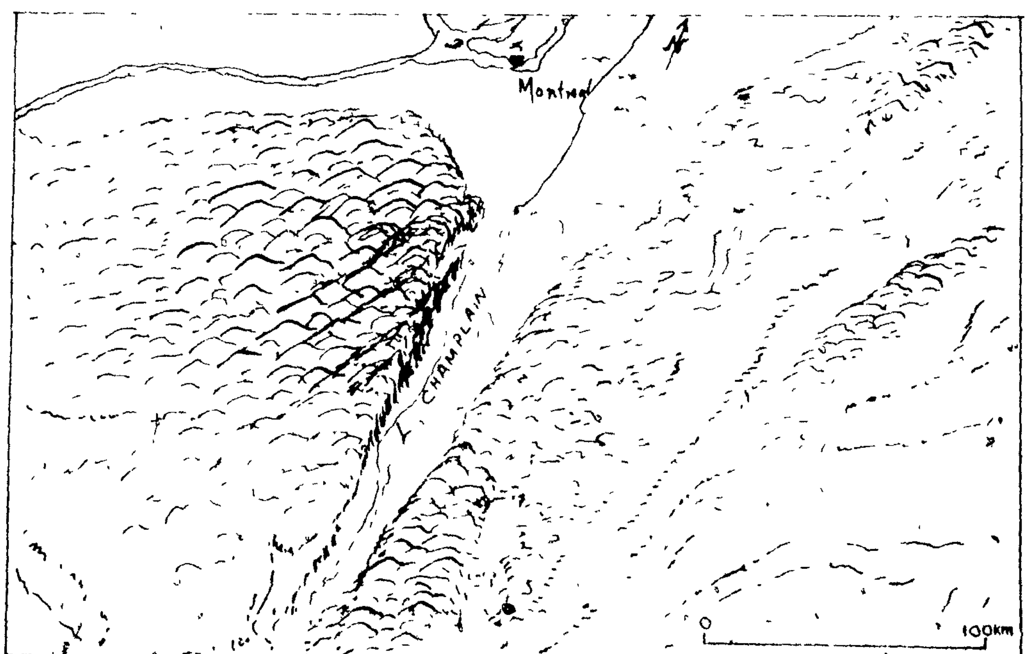


Fig. 12 Physiographic sketch of the Champlain Valley and vicinity.

of about 70 km. It is bounded for the most part by abrupt scarps, 200 to 300 m high. The Champlain Valley, like the St. Lawrence Valley, is largely cut on a strip of rock of relatively higher erodibility, that lies between the Shield and the Appalachian foldbet.

GEOLOGICAL CHARACTERISTICS

Tectonic Setting

Like the St. Lawrence Valley, the Champlain Valley is set along the boundary between the Appalachian foldbelt (Logan's Zone) and the platform, the larger part being over the platform. The eastern boundary of the Appalachian allochthon, which in the Champlain Valley area is called the Champlain thrust (Cady 1945, p. 565; also see Doll et al. 1961; Cady 1969, pp. 63-64) extends along the east side of Lake Champlain (Fig. 13). Although the thrust zone is generally marked by a minor scarp, the rise to the Green Mountains is 10 to 20 km east of the thrust zone. Therefore on the east side, a narrow strip of the Champlain Valley extends on to the Appalachian foldbelt.

The Adirondack Mountains on the west side of the valley are a part of the Grenville structural province. At least parts, if not all, of the Adirondack Mountains were once covered by Paleozoic platformal rocks, for remnants of these rocks still persist in some places as downfaulted outliers (N.Y. State Mus. and Sci. Ser.- Geol. Surv. map and Chart serv. No. 5).

General Geology

The platformal autochthonous sequence of the Champlain Valley is essentially similar to that of the St. Lawrence Valley. It begins with the Cambrian Potsdam sandstone, overlain by dolostones of the Beekmantown group (Lower Ordovician). Discomformably above the Beekmantown is a Middle Ordovician sequence, consisting of a lower limestone (several thin units), and a thick upper black shale that becomes sandy towards the top. This succession is about 750 m thick on the west side of the valley (Rodgers 1970, p. 72) and increases in thickness towards the east.

On the foot-wall side of the Champlain thrust, the platformal rocks are generally crumpled and flanked on the east side by minor thrust slices such as the Highgate Springs slice (Kay 1958). This crumpled belt is actually the southward continuation of the par-autochthonous St. Germain Complex of the St. Lawrence Valley. The belt is absent at the south end of the valley but widens to about 15 km at the north end of the lake (Hawley, 1967). West of the crumpled belt, the platformal rocks are not folded but they are broken by numerous high-angle faults, mostly of the normal type (Quinn 1933; Welby 1961, pp. 199-210). The strata in this area are either nearly flat-lying or are tilted at varying angles. Along the west margin of the valley, they have fault contacts or overlapping relationships with the Grenville crystalline rocks of the Adirondack Mountains.

The hanging-wall side of the Champlain thrust is in places characterized by thrust slices (e.g. Rosenberg slice, Rodgers 1970, p. 73) in the form of synclinoria, containing rocks as old as Lower Cambrian which have overridden the crumpled shales (Middle Ordovician) of the platformal sequence.

The Paleozoic rocks as well as the Grenville crystalline rocks in the Champlain Valley area have been intruded by a closely related group of alkaline igneous bodies, mainly in the form of dikes. These alkaline igneous rocks and the high-angle faults in the area are of special importance to the present problem and hence will be described in some detail below.

Alkaline Igneous Rocks

Numerous igneous bodies of petrochemically related types occur in the Champlain Valley region (Kemp and Marsters 1893; Welby 1961, pp. 186-190; Woodland 1962). These igneous rocks, like those of the southern St. Lawrence Valley area, have strongly alkaline affinites, and hence, some workers prefer to include the Champlain Valley area in the Monteregian Petrographic province (Gold 1968, p. 290).

The igneous bodies are mostly dikes; a few sills and plugs also occur. The majority of them are in the platformal and Appalachian rocks of the valley, but a few of them occur in the adjacent areas of the Green Mountains and Adirondack Mountains (Fig. 13).

The igneous rock types fall into two general types, light coloured bostonites and dark coloured lamprophyres. Bostonites are

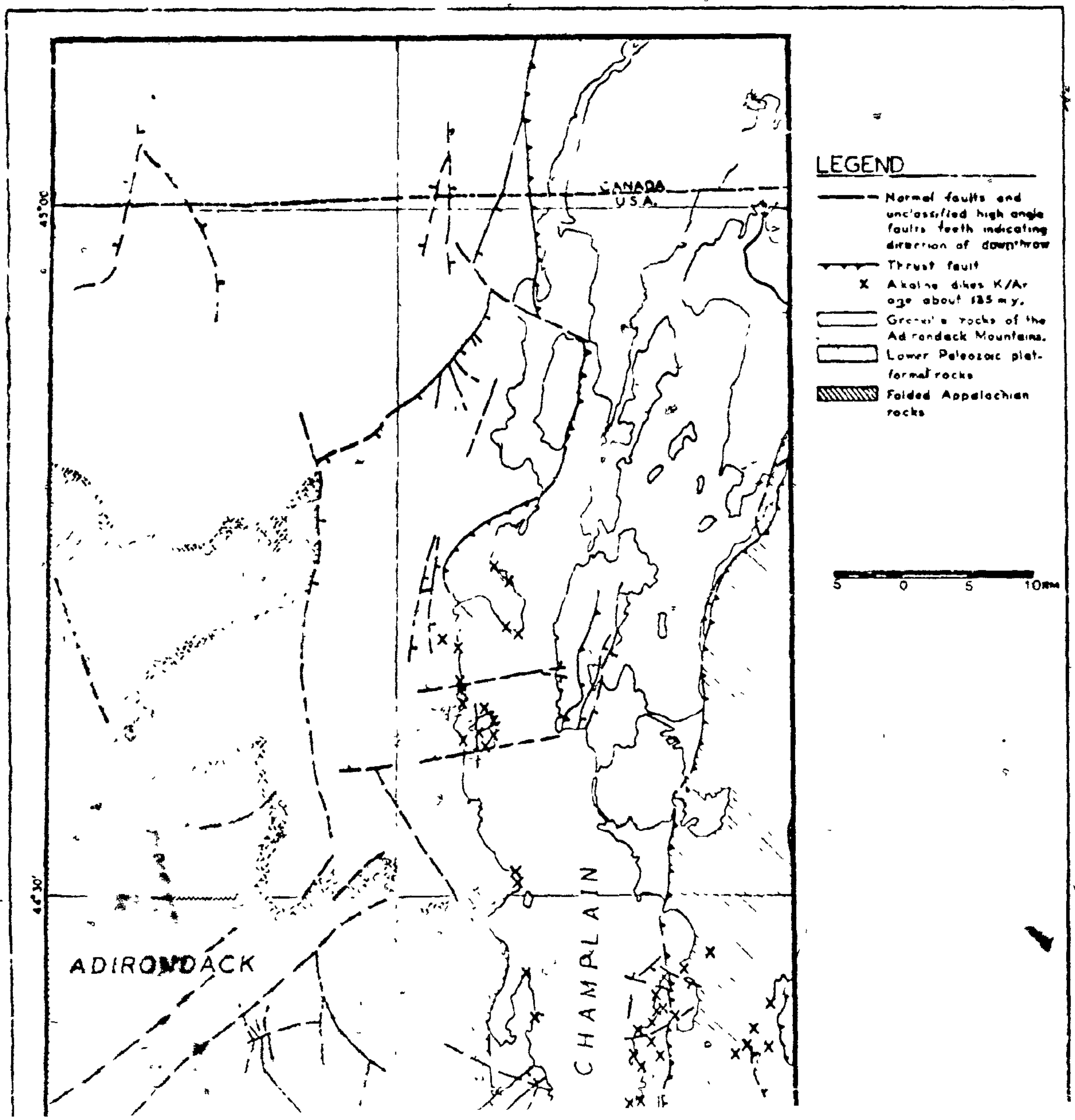
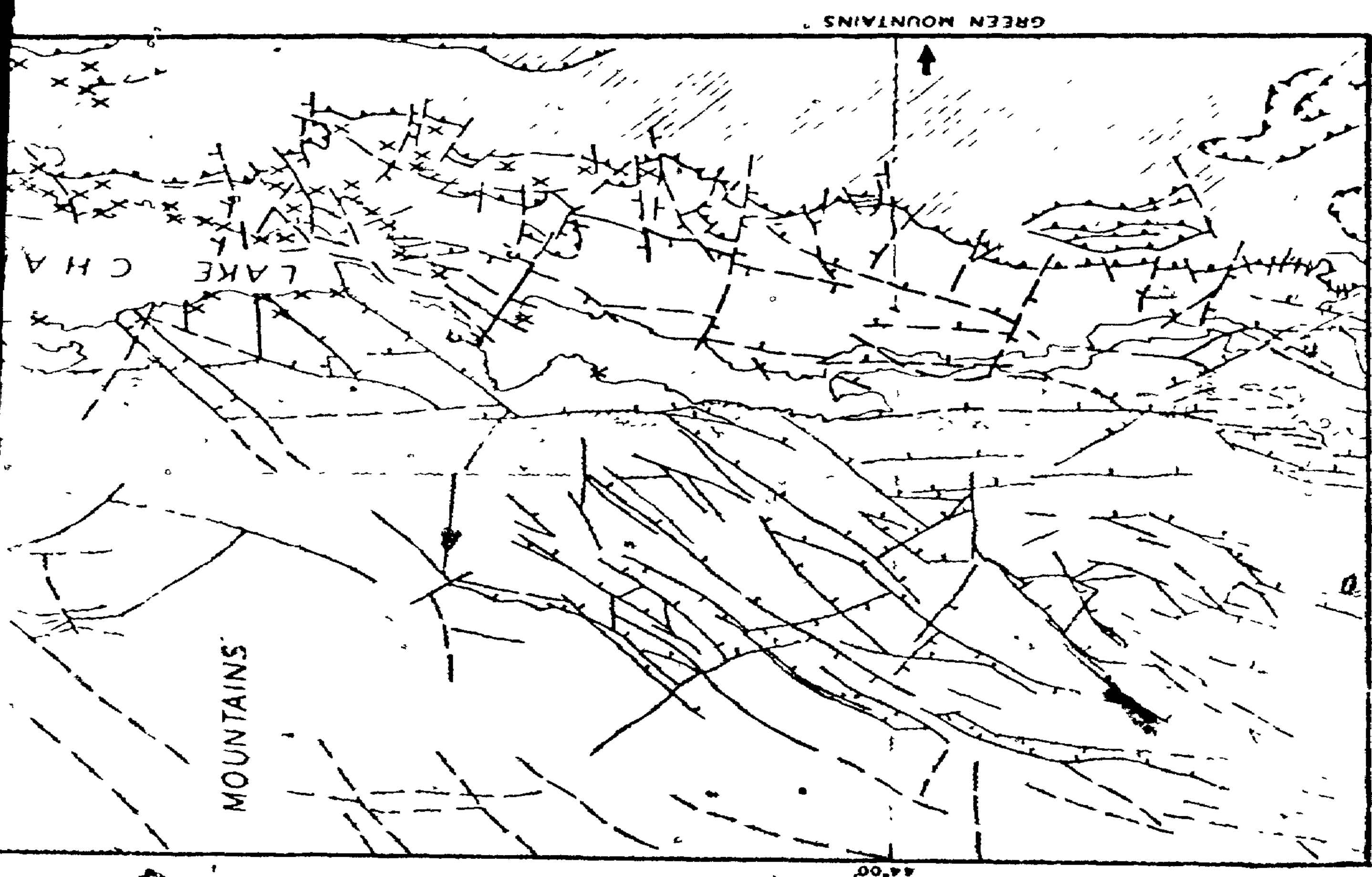


Fig. 13. Map showing selected geological features of the Champlain Valley area. Data compiled from Geological map of New York State(N.Y.State Mus. & Sci. Ser.- Geol.Surv. map & chart ser.no.5,1961), Welby 1961, Doll et al.(1961).



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the more common and form thicker intrusions. The thickest bostonite dike in the area is about 12 m in width (Kemp and Marsters 1893, p. 59). Many of the so-called bostonites are in reality bostonite breccias containing blocks of earlier formed bostonites or of country rock.

Of the lamprophyres the most common types approach the composition of camptonite while others can be classified as monchiquites. The lamprophyres generally form thinner bodies than do the bostonites. The thickest lamprophyre dike known is about 6 m (Kemp and Marsters 1893, p. 59).

It is generally believed that the bostonites followed the lamprophyres. Most of the cross cutting relationships, seen in the field, confirm this view, although instances where lamprophyres cutting bostonites also have been reported (Welby 1961, p. 189). It is probable that the two types of rocks are consanguinous and were emplaced during the same general period of magmatism.

Two K/Ar dates of Champlain Valley alkaline rocks have been published. An age of 136 ± 7 m. y. has been obtained by Zartman _et al._ (1967, p. 862) for biotite from a lamprophyre dike. Zartman _et al._ think that this age value is probably the true K/Ar age, although they were not able to check the result by the Rb/Sr method because of an unfavourable Rb/Sr ratio for the biotite. The other age is a value of 111 ± 2 m. y. for the syenite of the Barber Hill stock (Armstrong and Stump 1971) which is associated with bostonites. These Jurassic-Cretaceous ages are similar to those determined for the Monteregian rocks of the St. Lawrence Valley area.

High-Angle Faults

High-angle faults (Swinnerton 1932; Quinn 1933; Rodgers 1937, p. 1683; Doll et al. 1961; Welby 1961, pp. 199-210; Cady 1969, pp. 91-92; Rodgers 1970, p. 72) affect the autochthonous and parautochthonous rocks of the valley and in places cut the allochthonous rocks further to the east (Welby 1961, pp. 204-209). Numerous high-angle faults also cut the Grenville rocks of the Adirondack Mountains on the west side of the valley. The distribution of faults is shown in Figure 13. They form a well defined pattern which apparently splays out southwards, to involve a considerable area of the Adirondack Mountains. The faults can be traced into the Mohawk Valley area, beyond which they have not been traced (Rodgers 1970, p. 74).

The fault traces are rarely exposed; only in a few places, have fault-planes been actually observed (e.g. see Welby 1961, p. 199; Fisher and Hanson 1951, p. 809). However, some of the faults have been well authenticated by geological mapping, although many others, especially those believed to affect the floor of Lake Champlain and the Adirondack Mountains, have been inferred on physiographic evidence. Some of the fault traces appear to be rather sinuous or curved, but on the average, the fault trends fall into two sets: one principal set trending north-northeast and hence more or less parallel to the valley and the other set approximately transverse to the valley direction.

The longitudinal faults are best known from the west side of the valley and from the adjacent areas of the Adirondack Mountains, and are believed to be normal faults (Quinn 1933; Rodgers 1970, p. 72). Many of the faults of the Adirondacks strike northeast, instead of north-northeast as in the valley. The faults on the west side of the valley generally have throws up to about 600 m in Paleozoic rocks and in one extreme case a throw of about 1200 m (See Quinn 1933, p. 120) has been estimated. The fault planes, where observable, dip between 60° and 70° but some are almost vertical. Many have down-throws to the east, although, the reverse is sometimes the case. Where the faults are downthrown on the east, the blocks tend to be tilted west and vice versa. Welby (1961, p. 197) states that although majority of the longitudinal faults on the east side of the valley are normal, some are reverse. Further to the south, Zen (1972, p. 27) also reports both normal and reverse faults from the autochthonous shelf rocks on both sides of the Taconic Klippe.

The cross-faults intersect longitudinal faults of the valley almost at right angles and are known to extend eastwards into the allochthonous rocks on the west flank of the Green Mountains. They cut the Champlain thrust and the flanking thrust slices (Welby 1961, pp. 205-210; Rodgers 1970, p. 73). The cross-faults are believed to be nearly vertical normal faults (e.g. Welby 1961, p. 204). Throws in Paleozoic rocks up to about 600 m have been reported along them (Welby 1961, p. 204). The cross-faults, together with the longitudinal faults, break up the Champlain Valley floor into a mosaic of blocks. As mentioned earlier, the majority of these blocks are tilted west along longitudinal faults; they are tilted north and south along cross-faults.

No conclusive evidence has yet come to light concerning the relative ages of longitudinal and cross-fualts, although evidence of an indirect nature seems to indicate that, in some instances, longitudinal faults are off-set by cross-faults (Quinn 1933, p. 120; Welby 1961, pp. 210-211). Nevertheless, most workers agree that the two sets of faults are probably related in time and cause.

The high-angle faults of the Champlain and St. Lawrence Valleys although described separately in this paper, should be regarded as parts of a single fault system because:

1. faults in the two areas are similar
2. the tectonic setting of faults in the two areas is identical
3. presumably comagmatic rocks are associated with some of the faults of both areas
4. some of the faults in the Champlain Valley extend into the St. Lawrence Valley
5. the faults in the two areas appear to have similar evolutionary histories (see below).

Ages of High-Angle Faults

The high-angle faults of the Champlain Valley cut the youngest Ordovician rocks present, which are late Middle Ordovician (see Rodgers 1970, p. 72). They also cut the Champlain Thrust, but these movements can be post-Ordovician or post-Devonian, depending on which age is assigned to the Champlain Thrust. Wiesnet (1961) has pointed out that the progressively greater abundance of felspar in the Potsdam, towards the eastern edge of the Adirondack massif, may indicate pre-Potsdam normal faulting in the Precambrian basement. He suggests that high-angle faults in the Champlain Valley area may be related to movements along buried pre-Potsdam faults. Zen (1972, p. 28) thinks that large vertical movements took place on these faults in Middle Ordovician times, roughly synchronousely and mutually connected in cause with the Taconic orogeny and that the faulting resulted in crustal subsidence that led to deeper water conditions and deposition of black shale in the area. Evidence for Middle Ordovician and/or pre-Middle Ordovician normal faulting has been observed by Fisher and Hansen (1951, pp. 808-809).

Some of the high-angle faults in the area cut mafic dikes (Hudson and Cushing 1931, p. 101) and both cut and are cut by felsic dikes (Buddington and Whitcomb 1941, p. 26). Because the two types of dikes appear to be consanguinous, the above cross cutting relationships indicate mid-Mesozoic faulting. Mild seismicity and possible evidence for post-glacial faulting in the area (Oliver et al. 1970) indicate that some of the faults may still be in an unhealed state. Thus, apart from uncertainities introduced by the lack of record, the chronology of high-angle faulting of the area appears to be similar to that of the St. Lawrence Valley area, and can be summarized as follows.

1. The faults were active in post-late Middle Ordovician times. Some of this faulting may have taken place in the mid-Mesozoic synchronously with the alkaline igneous activity in the area.

2. Significant fault movements also may have taken place roughly synchronously with the establishment of deeper water conditions in the Middle Ordovician times.

3. High-angle faulting may have occurred along the west margin of the valley before the deposition of the Potsdam, the upper part of which is late Cambrian.

4. Some of the faults may still be in an unhealed state.

Structural Effects of High-angle Faulting

The part of the valley overlying the Champlain thrust has a complex structure. There, normal faults, high-angle reverse faults and east-dipping low-angle thrusts occur in close association. The interrelationships of these faults are poorly known. To the west of this area where autochthonous platformal rocks are cut by normal faults, the structure consists of a longitudinal belt of structurally down-dropped graben-horst terrain (Fig. 14). Along the greater part of the Adirondack margin, structural subsidence has taken place on en echelon faults which in places overlap to produce step faulting. The overall structure of the downfaulted zone is probably that of a complex graben. Its east margin is rather indefinite but the normal faults have not been traced to any significant distance, east of the Champlain Thrust. Thus the width of the downfaulted zone may be about 50 km at its north end and gradually decreasing southwards to about 20 km at the south end of the Champlain Valley. The length of the downfaulted zone is about 200 km and it together with the zone of downfaulting along the St. Lawrence Valley forms a continuous rift zone about 650 km long.

In view of the earlier discussions on the ages of high-angle faults it appears that significant structural subsidence along the Champlain-St. Lawrence Valleys took place in post-Ordovician times. A major part of this subsidence may have taken place in the mid-Mesozoic, synchronously with the Jurassic-Cretaceous alkaline-carbonatite activity. There are indications that the post-Ordovician movements, at least in places represent renewed movements on earlier faults which appear to have existed during the early stages of Appalachian evolution.

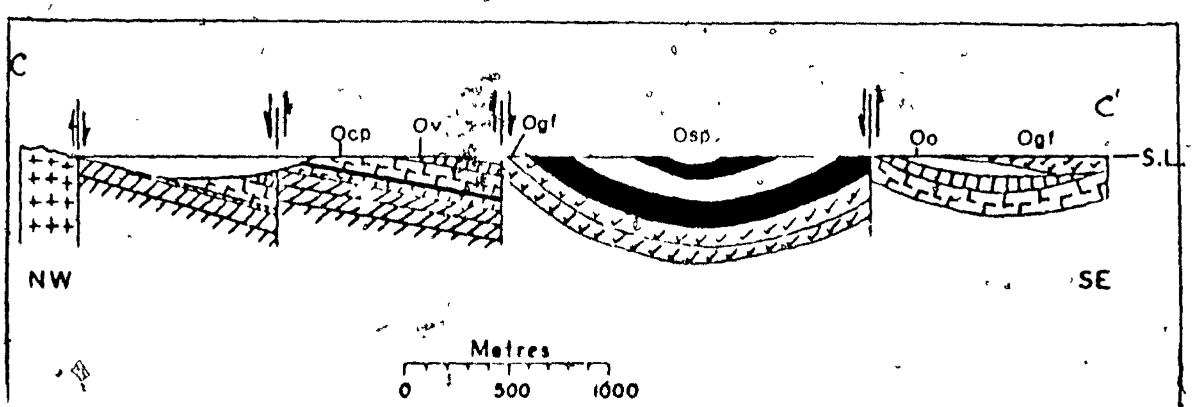

Fig. 14. Section across a part of the Champlain Valley showing grabens and horsts in platformal rocks. After Welby (1961). For location of section see Fig. 13. Key to abbreviations: Ocp-Crown Pt. Ls.; Ov-Valcour fm.; Ogf-Glens Falls Ls.; Osp-Stony Pt. Sh.; Oo-Orwell Ls..

SUMMARY AND CONCLUSIONS

The Champlain Valley is largely underlain by a structurally ~~downdropped~~ horst-graben terrain whose gross structure appears to be that of a complex graben. Southwards, the faults related to the structure appear to terminate in the Mohawak Valley area. Northwards it is continuous with the zone of downfaulting along the St. Lawrence Valley, and together these zones form a continuous rift structure about 650 km long. This structure is characteristically restricted to the narrow platform along the western margin of the Appalachian foldbelt. Significant structural subsidence along this rift zone took place in post-Ordovician times. A large part of this subsidence probably took place in the mid-Mesozoic, synchronously with the Monteregian alkaline-carbonatite activity. The post-Ordovician movements may be related to renewed activity on older faults which possibly existed during the early stages of the Appalachian evolution.

CHAPTER VI: MARGINAL SEGMENT:

THE ESQUIMAN CHANNEL AND VICINITY

INTRODUCTION

The Esquiman Channel, which is also called the Belle Isle Trough in older literature is the northernmost part of the Marginal Segment of the St. Lawrence Valley system. It is the largest tributary of the Laurentian Channel, and branches off from the latter just southeast of Anticosti Island (Fig. 15). Thence it extends under the St. Lawrence Gulf, between Newfoundland and the mainland and loses its identity as a distinct topographic feature before reaching the Strait of Belle Isle. The Esquiman Channel itself is set within a much wider topographic low, lying between the Long Range Mountains of Newfoundland where elevations of 300 m to 700 m above sea-level are common, and the Mecatina plateau of the Shield Region where elevations generally range from 200 m to 500 m. This wider topographic low will be referred to as the Esquiman Channel Area and is the topic of discussion in this chapter. The topographic low includes not only the area under the Gulf of St. Lawrence but also the strip of lowlands on the southeast side of the Gulf - the Newfoundland Coastal Lowlands.

The Esquiman Channel is a trough-like submarine depression. Where it joins the Laurentian Channel 400 m depths are common. Similar depths continue only for about 25 km along the channel, but 200 metre

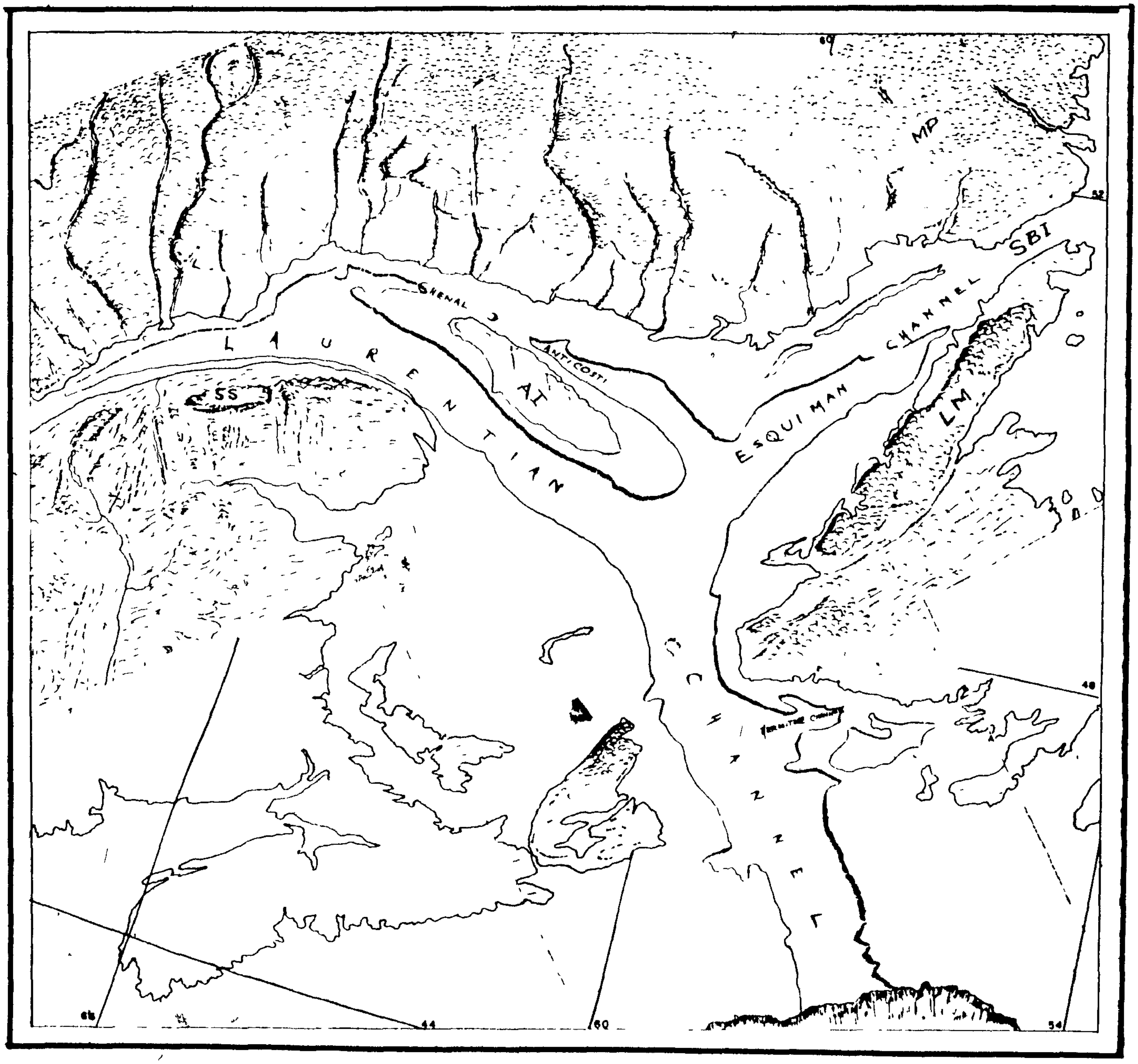

Fig. 15. Physiographic sketch of the Laurentian and Esquiman Channel areas. Key to abbreviations: AI-Anticosti Island; MP-Mecatina Plateau; LM-Long Range Mountains; SBI-Strait of Belle Isle; SS-Shickshock Mountains.

depths are common for a distance of about 250 km. The Esquiman Channel is widest (about 100 km) where it joins the Laurentian Channel and tapers off in the direction of the Strait of Belle Isle. The Esquiman Channel Area has an outline similar to that of the channel itself but has a maximum width of about 250 km. The rise towards the Mecatina plateau is gradual but the rise to the Long Range Mountains commonly takes the form of an abrupt scarp, 200 to 300 m high.

GEOLOGICAL CHARACTERISTICS

Tectonic Setting

The tectonic setting the Esquiman Channel Area is similar to that of the Champlain - St. Lawrence Valleys, being mainly a narrow platform, lying between the Appalachian foldbelt and the shield.

General Geology

The Esquiman Channel Area contains a thick sequence of Paleozoic strata which is believed to overlie a Grenville basement. Grenville rocks are exposed along most of the mainland coastline (Fig. 16) and Grenville-like rocks also occur in Long Range Mountains (Clifford 1969), the Indian Head Range (Williams 1967) and on Belle Isle (Williams and Stevens 1969). The cover rocks consist mainly of - (a) an autochthonous shelf sequence ranging in age from Early Cambrian or possibly from late Precambrian to Middle Ordovician, (b) an allochthonous eugeosynclinal sequence of about the same age as the shelf sequence, and (c) Carboniferous successor basin deposits.

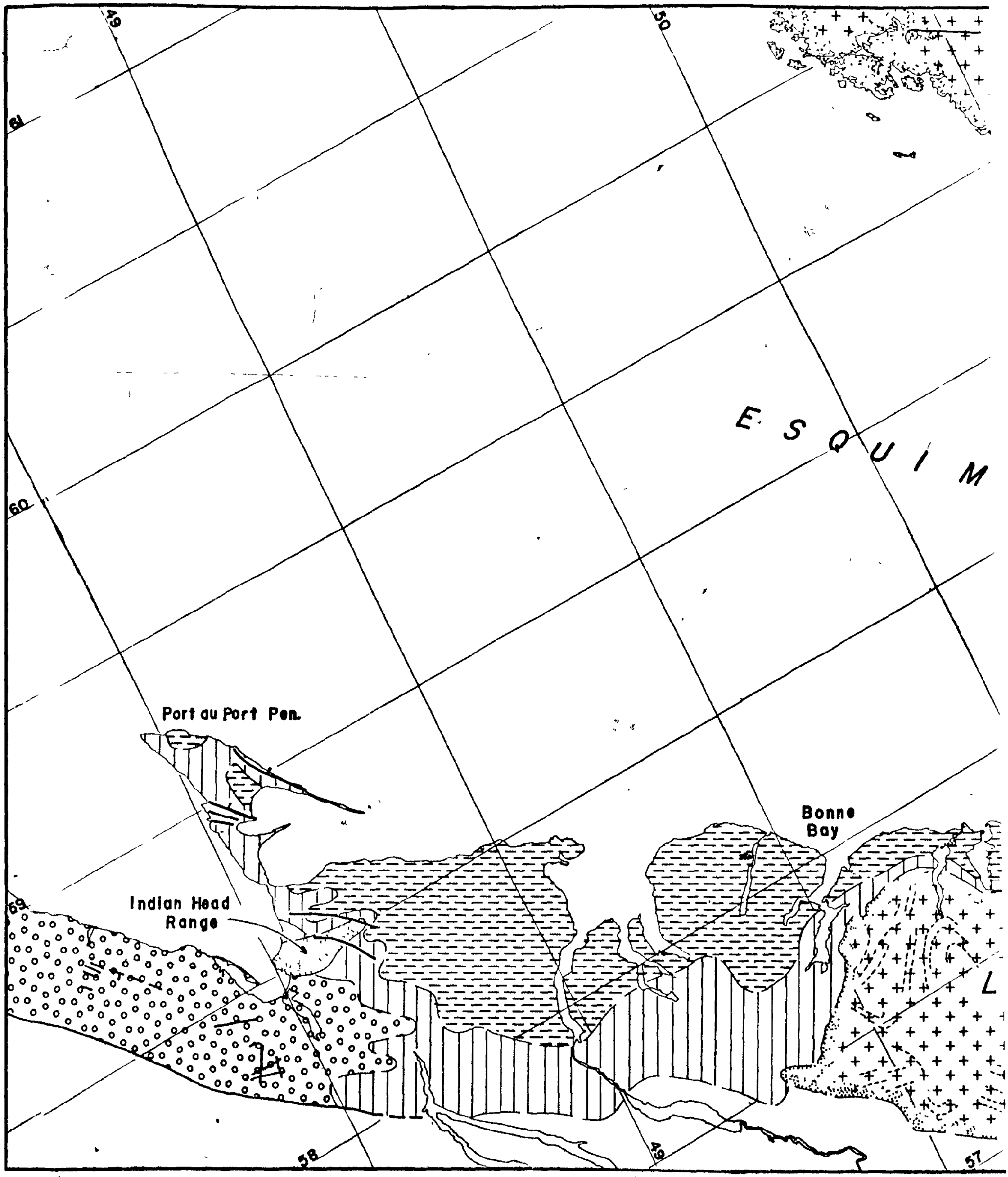

Fig.16.Geology of the Esquiman Channel Area.Data from Clifford 1969; Cumming 1967,1972; Davies 1963,1965, 1968; Strong and Williams 1972 and other sources.

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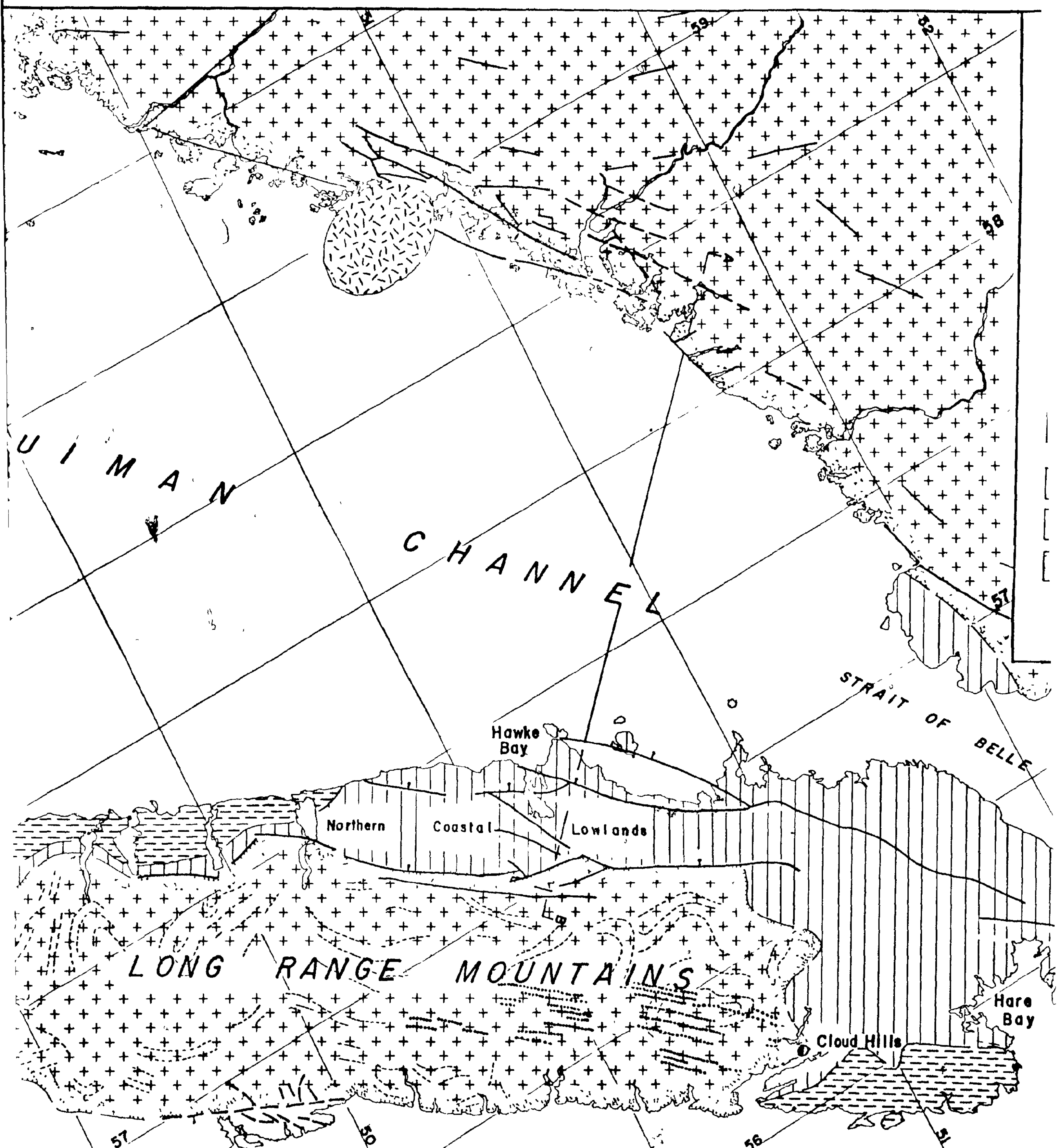

ries 1963,1965a&b,

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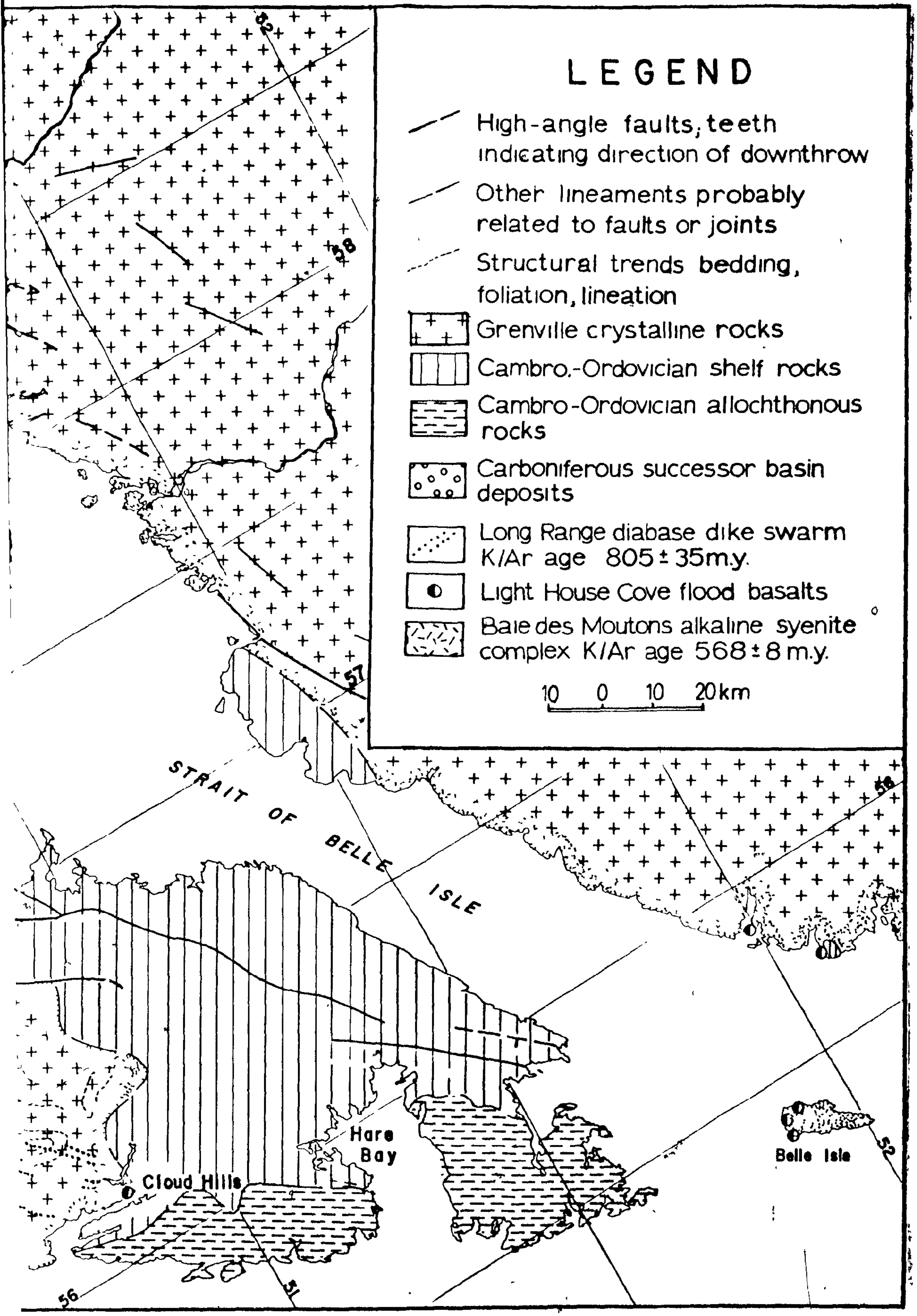
LEGEND
High-angle faults; teeth indicating direction of downthrow
Other lineaments probably related to faults or joints
Structural trends bedding, foliation, lineation
Grenville crystalline rocks
Cambro.-Ordovician shelf rocks
Cambro-Ordovician allochthonous rocks
Carboniferous successor basin deposits
Long Range diabase dike swarm K/Ar age 805±35m.y.
Light House Cove flood basalts
Baie des Moutons alkaline syenite complex K/Ar age 568±8 m.y.
10 0 10 20km
STRAIT OF BELLE ISLE
Hare Bay
Cloud Hills
Belle Isle
52
58
57
56
56
51
52

The basal portion of the autochthonous shelf sequence (Labrador group) consists of Lower Cambrian clastic rocks ranging from white and pink quartzite and reddish arkosic sandstone to dark shale; minor beds of limestone or dolostone occur nearly everywhere in the sequence (Rodgers 1970, p. 149). On the mainland, rocks belonging to this group, are exposed along the northwest share of Strait of Belle Isle. A sequence of less than 150 m of rock is preserved there. Dips are gentle to the southeast, away from the shield. Across the Gulf, on the northwest side of the Long Range Mountains, dips are steep to the northwest, away from the Precambrian core. The Labrador group as a whole becomes thicker and more shaly to the southeast. Basaltic volcanic rocks (Light House Cove formation) and also probable feeder dikes occur at the base of the Lower Cambrian strata in several northeastern areas (Williams and Stevens 1969, pp. 1151-1152). Also from Belle Isle, Williams and Stevens (1969) have described boulder conglomerate and arkosic sandstone overlain by thick white quartzite (Bateau Formation) at the base of strata known to be lower Cambrian. These rocks as well as the basaltic voncanics may be in part late Precambrian (Williams and Stevens 1969, p. 1147). The major rock unit overlying the basal clastic sequence is a thick (600 to 1200 m), massive carbonate (St. George group), mainly dolostone and some limestone in the upper part. These rocks contain Lower Ordovician fossils. Disconformably overlying the carbonate sequence are lower Middle Ordovician limestones, commonly argillaceous, and becoming more shaly upwards, grading into graptolitic black shale (Table Head formation). The upper shale unit of this formation was derived from the east (Williams 1972b, p. 192), and is believed to represent the eastern part of a westward transgressing flysch wedge. In the Esquiman Channel Area black shales appear somewhat lower in the Middle Ordovician than in the Champlain and St. Lawrence Valleys.

In the area south of the Strait of Belle Isle, the Cambro-Ordovician shelf sequence is nearly flatlying, whereas in the Northern Coastal Lowlands and further to the southwest it is considerably deformed and broken by high-angle faults. According to Lilly (1966), the deformation of these rocks dies out northwestwards within a short distance from the shoreline. Southeastwards, the deformed shelf rocks terminate with fault-contact against the Precambrian massif of the Long Range Mountains whose structure is essentially horst-like, being bounded on its east and west sides by high-angle faults. The shelf sequence thickens southeastwards, but because of deformation, a maximum thickness figure cannot be determined. This thickening seems to be reflected in the pattern of the magnetic field over the area. It is highly irregular on the northwest side of the Esquiman Channel Area, indicating that in this area the more highly magnetic Grenville basement is buried, if at all, only under a thin cover of shelf rocks (Keen et al. 1970, p. 271). On the southeast side of the Channel Area, the magnetic field is relatively smooth presumably reflecting the greater depth of the basement there. The thickness of the shelf sequence along line A-B of Figure 16 has been estimated to be about 1.5 km (Sheridan and Drake 1968, Fig. 10) by seismic refraction work.

The paleographic implications of the shelf sequence in the Esquiman Channel Area, are similar to those of the Champlain and St. Lawrence Valleys and can be summarized as follows.

1. The southeast thicknening wedge of shelf rocks, at least in the area to the west of the Long Range massif, was deposited in a graben-like trough (asymmetrical graben?) which began its development before the deposition of the basal beds in the Early Cambrian and possibly in the Hadrynian (Clifford 1969, p. 651). In the basal units, the presence of immature arenites and boulder conglomerates, presumably of local derivation indicate high relief of the adjacent Precambrian areas.
2. The carbonate sequence of the St. George group, suggests that in the Early Ordovician the Esquiman Channel Area became a stable carbonate shelf.
3. The deposition of early Middle Ordovician flysch indicate deeper water conditions possibly brought about by subsidence of the carbonate shelf. The influx of flysch and movement of klippen (discussed below) from the southeast, are probably related to uplift in the geosyncline (Equivalents of the Taconic Mountains?). The subsidence of the carbonate shelf

in the Esquiman Channel Area appears to have taken place slightly earlier than in the Champlain and St. Lawrence Valleys.

The belt of shelf rocks extends southwestwards from the North Coastal Lowlands as far as the Indian Head Range area, but in between it is partly buried under a eugeosynclinal sequence of about the same age as the shelf rocks (Fig. 16). The eugeosynclinal rocks are believed to be allochthonous (Rodgers and Neale 1963; Kay 1969) forming one or two klippen that moved to their present position (from a source area to the east) in Middle Ordovician times, a situation similar to that of the Taconic Klippe area, southeast of the Champlain Valley. The allochthonous rocks include ultramafic rocks (e.g. see Smith 1968) which are believed to be basal parts of ophiolite complexes (Williams et al. 1972, p. 236).

Southwest of the Indian Head Range area, the Cambro-Ordovician shelf rocks are overlain by Carboniferous successor basin deposits, laid down in a rift zone (Fundy Basin Rift, see Belt 1968). Also in the northwest coast of Port-au-Port Peninsula, some 450 m of Early Devonian (possibly also Late Silurian) strata composed mainly of red beds and lighter clastic rocks (Calm Bank formation) are present.

Two aspects of the geology of the Esquiman Channel area (i) high-angle faulting (ii) post-Grenville igneous activity (which may be chronologically related to faulting), are of special interest to the graben problem as conceived in this paper. These two aspects are discussed in some detail below.

Post-Grenville Igneous Activity

Baie-des-Moutons Central Complex and the Associated Dike Swarm

Spatially associated with the faults of the mainland coast are post-Grenville igneous bodies of alkaline affinities. One main intrusive centre, referred to as Baie-des-Moutons intrusion (Fig. 16), is known. It is a roughly circular pluton, about 25 km in diameter and is composed mainly of coarse-grained syenitic rocks (Davies 1968). Closely associated with the syenitic complex is a dike swarm of varying composition, but of distinctly alkaline character. Like the faults in the area, the majority of dikes are parallel to the coastline. The dikes have been divided, according to their age relations with the central complex, into two main groups (Davies 1968). The group, consisting of basic lamprophyres and aplitic dikes, seems to be comagmatic with Baie-des-Moutons complex. The other group as a whole is younger, but consists of dikes of at least four different ages (Gerencher and Gold 1968). They range in composition from alkaline gabbro to syenite and to carbonatite.

The Baie-des-Moutons complex has been age dated (K/Ar on biotite in the syenite and in a related carbonatite dike) at 568 $\pm$ 8 m. y. (Doig and Barton Jr. 1968, p. 1403). Of the younger dikes, a single date of 470 m.y. (Rb/Sr on biotite) from an alkaline gabbro dike (Davis 1968, p. 217) is available. Alkaline gabbro dikes are the oldest of the younger group of dikes.

Flood Basalts and Dike Swarms in the Belle Isle Area

In the Belle Isle area, thin undeformed sequences of flood basalts are exposed in two areas (Clifford 1965; Williams and Stevens 1969; Strong and Williams 1972): one on the mainland coast, northwest of Belle Isle, the other on Belle Isle (Fig. 16). Similar flood basalts also occur at Cloud Hills on the east side of the Long Range massif. The lavas in the three areas have been correlated because of their similar petrochemistry and stratigraphic position, and have been named the Light House Cove formation by Williams and Stevens (1969). It is believed that these lava occurrences are the remnants of one or more once extensive plateau basalt fields. Diabase dikes are ubiquitous in the general area of the flood basalt occurrences and in the northeastern part of the Long Range Mountains make up a well-defined northeast trending dike swarm (Fig. 16). Some of the dikes are considered to be feeders of the existing basalt flows (Williams and Stevens 1969).

As mentioned earlier, stratigraphic evidence indicates that the lavas are either early Cambrian or late Precambrian (Williams and Stevens 1969). However, diabase dikes from the Long Range swarm have been age dated at 805 ± 35 m. y. (Pringle et al. 1971). This may be the age of the Light House Cove formation. The Bateau Formation at the base of the shelf sequence in Belle Isle is cut by diabase dikes similar to those dated as 805 m. y. The dikes and lavas are, of course, post-Grenville.

High-Angle Faulting

The coastal belt, southeast of the Gulf is broken by northeast trending high-angle faults (Fig. 16). This longitudinal fault system is best known from the Northern Coastal Lowlands, where it dissects the shelf sequence. Both normal faults (steeply northwest dipping to vertical) and high-angle reverse faults (steeply southeast dipping to vertical) are present (Cumming 1967, p. 18, Riley 1962, pp. 46-50). The faults at the contact between the Precambrian rocks of the Long Range massif and the shelf rocks, where observable, are southeast dipping high-angle reverse faults (Clifford 1969, p. 651). Movements on most faults are dip-slip, although Riley (1962) has observed strike-slip and oblique slip movements on some faults of the Stephenville Map-Area. The downthrows are always towards the channel. Cumming (1967, p. 12) states that the degree of faulting appears to be much greater towards the shoreline, indicating the possibility that faults may be present in the water covered Gulf area as well.

What is possibly a part of the same fault system described above, is present along the coastal belt of the mainland. For instance, Cumming (1972, p. 3) has described a northeast trending normal fault, some 30 km long, extending roughly along the Precambrian-Paleozoic boundary on the north side of Strait of Belle Isle (Fig. 16). Also according to Davies (1968, pp. 268-271, also see Davies 1963, 1965a & b) most, if not all, of the many linear valleys that characterise the coastal belt, are fault-line valleys. Davies (1968, pp. 270-271) points out that the precise nature of the underlying faults is difficult to determine, but the offsets (of lithic units) produced by some of the faults (Davies 1968, pp. 270-271) cannot be explained without admitting some (less than 1 km) transcurrent movements. But the offsets produced by the majority of the faults are compatible with downfaulting towards the Esquiman Channel.

The longitudinal fault system of the Esquiman Channel Area probably continues through the Strait of Belle Isle, because north-east trending high-angle faults have been noted on Belle Isle by Williams and Stevens (1969, p. 1148). They think that except for one fault which appears to contain strike slip movements, the others are probably normal faults or southeast dipping high-angle reverse faults. Apart from the longitudinal faults minor faults of transverse trend are present in some areas (see Riley 1962, p. 46) but there is little information available regarding the nature of these faults.

Ages of High-angle Faults

The age question of the high-angle faults of the Esquiman Channel Area, cannot be answered with certainty as there are large gaps in the record and the available evidence itself cannot be interpreted unequivocally. The problem is quite similar to that of the Champlain-St. Lawrence Valleys. Any attempt to answer the age question, has to take the following evidence and points into account.

1. In the Belle Isle area, the emplacement of northeast trending diabase dikes and the eruption of flood

basalts (Light House Cove formation) indicate a tensional stress environment which may have been accompanied by crustal rifting (Strong and Williams 1972, p. 51). Rifting and crustal subsidence are also indicated by the graben-like development of the surface on which the shelf sediments on the west side of the Long Range massif was deposited (Clifford 1969). Except for the Bateau formation in the Belle Isle, the basal units of the shelf sequence appear to have been deposited soon after the eruption of flood basalts in the Belle Isle area. The basal units contain immature arenites and boulder conglomerates, presumably of local derivation, indicating high relief (block uplift?) of the adjacent Precambrian areas. According to the available K/Ar ages of the diabase dikes (Pringle et al. 1971) the above rifting event may have taken place about 800 m. y. ago.

2. The close spatial association of Baie-des-Moutons alkaline complex and of the associated dike swarm with high-angle faults along the mainland coast indicate that the magmatic events may be related in cause and time to fault movements in the area. The K/Ar age of Baie-des-Moutons Complex is about 568 m. y. (Doig and Barton Jr. 1968) and some of the associated dikes are younger.

3. The subsidence of the Early Ordovician carbonate shelf and the establishment of deeper water conditions in the early Middle Ordovician may have been achieved by movements on high-angle faults (Zen 1972).
4. Faults in the North Coastal Lowlands cut Middle Ordovician rocks. Fault movements seem to have taken place after the emplacement of klippen (Brückner 1966, p. 89; also see Tuke and Baird 1967, p. 9) although a case can also be made for Ordovician movements before the emplacement of klippen (Zen 1972, p. 26). High-angle faults parallel to those of the North Coastal Lowlands cut Devonian rocks of Port au Port peninsula and Carboniferous rocks further to the southwest.

The main conclusions that can be drawn from the above evidence and discussions are as follows:

1. High-angle faults in the area may have originated before the shelf sequence began to be deposited, roughly synchronously with the eruption of Light House Cove lavas, possibly about 800 m. y. ago. The Bateau Formation in Belle Isle appears to be older than the Light House Cove lavas, but may have been deposited within a relatively short period of time.
2. Later fault movements in the area, including the large post-Ordovician movements on faults of the North

Coastal Lowlands, may represent renewed movements on the older faults. The youngest rocks involved in fault movements of the area are Carboniferous.

Tectonic Significance

The Esquiman Channel in its present form appears largely to be a product of erosion. It is presumably cut into Paleozoic rocks. Its trough-like form is probably of glacial origin. However, the wider topographic low in which the channel is set, i.e. the Esquiman Channel Area, is bordered, in part at least, by high-angle faults, on which the Esquiman Channel Area has structurally downdropped with respect to the adjacent areas, suggesting that the faults may constitute a structural framework on which the topographic low has developed.

The faults along the coastal belts include large normal faults. In the southwestern part of the Strait of Belle Isle, vertical or inward dipping normal faults outline a downfaulted block with a graben-like cross section. As already discussed earlier, the emplacement of northeast trending diabase, dike swarms and the intrusion of Baie-des-Moutons complex may have been accompanied by tension fracturing. Also mentioned earlier was that the basement configuration beneath the shelf rocks along parts of the southeastern coastal belt suggest a graben-like development. The above criteria are interpreted as indicating that the Esquiman Channel Area is underlain by a downfaulted rift zone (Fig. 16), similar to the rift zone that characterise the platform along the Champlain-St. Lawrence Valleys.

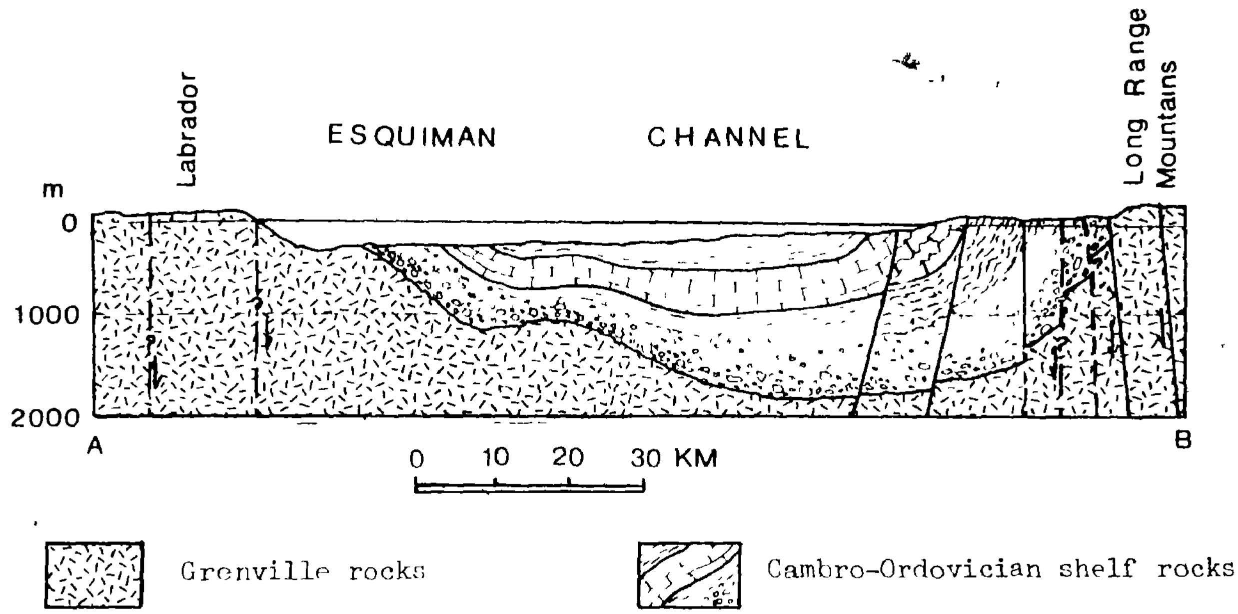

Fig. 17. Possible structural section across the Esquiman Channel Area. Basement configuration partly from Sheridan and Drake 1968. For location of the line of section see Fig. 16.

Along the southeastern coastal belt, however, northwest dipping normal faults occur in close association with southeast dipping high-angle reverse faults. It is nonetheless assumed that the high-angle reverse faults are restricted largely to a relatively narrow belt bordering the Appalachian foldbelt. Such an assumption appears reasonable in view of the analogous situation in the Champlain and St. Lawrence Valleys.

In view of the earlier discussions on ages of high-angle faulting, it appears that the rift zone postulated above, originated about 800 m.y. ago, during the very early stages of the Appalachian evolution. The southeast dipping high-angle reverse faults along the southeastern coastal belt were probably features imposed on the area by tectonic stresses related to orogeny in the adjacent geosyncline. There was renewed rifting and structural subsidence of the area in post-Ordovician times.

SUMMARY AND CONCLUSIONS

The geology of the Esquiman Channel Area is largely concealed. Therefore, interpretation of its possible structure has to be based mainly on evidence known from the coastal areas. This evidence, although fragmentary, indicates that the area is underlain by a downfaulted rift zone. This rift zone appears to have formed in the Hadrynian, during the very early stages of the Appalachian evolution, and seems to have controlled to a greater or lesser extent the subsequent tectonic evolution of the area. Diabase dikes, presumably emplaced during the initial rifting give K/Ar ages of about 800 m. y. The rift zone responded to subsequent tectonic stresses imposed on the area, including a phase or phases of post-Ordovician tensional stresses, during which renewed rifting occurred.

CHAPTER VII: MARGINAL SEGMENT:

THE INNER PART OF THE LAURENTIAN CHANNEL AND VICINITY

INTRODUCTION

The topographic low of the St. Lawrence Valley continues northeastwards beyond the Saguenay outlet and is almost completely occupied by the St. Lawrence Estuary. Just below the Saguenay outlet, the bottom of the estuary deepens rather abruptly (from about 35 to about 200 m), to form a trough-like depression known as the Laurentian Channel (Fig. 15). On the upstream side of Anticosti Island, the Laurentian Channel is joined by a minor branch channel: Chenal d'Anticosti. Approximately where they join, the main channel bends rather sharply clockwise through an angle of 45^{o} (from N63^{o}E to N108^{o}E) and extends on the south side of Anticosti Island. The branch channel runs on the north side of the Island. On the downstream side of Anticosti Island the Laurentian Channel branches again the branch being the Esquiman Channel. At this bifurcation too, the main channel bends clockwise through an angle of about 45^{o} and extends along Cabot Strait (between the Islands of Newfoundland and Cape Breton) and thence across the entire continental shelf. The part of the channel between the Saguenay outlet and where it is joined by the Esquiman Channel is the inner part of the Laurentian Channel. It is about 800 km long, more than the lengths of Champlain and St. Lawrence Valleys taken together. It links the topographic lows of the Champlain and St. Lawrence Valleys with the Esquiman Channel. The area discussed in the present chapter includes not only the channel itself but also the general area of the St. Lawrence Estuary and Gulf in which the channel is set, thus including Chenal d'Anticosti and Anticosti Island (Fig. 15).

The inner part of the Laurentian Channel is essentially a submarine trough. Downstream from the Saguenay outlet, it increases in width progressively and just about where it is joined by Chenal d'Anticosti attains a width of about 80 km. Depths of a little more than 200 m are common. In this part, the channel is set in a broader depression, for elevations on either side of it, rise within a short distance, to more than 100 m above sea level. Further downstream, the channel forms a nearly straight trough with a fairly uniform width of about 60 km. Depths of 300 to 400 m prevail with unconnected oval depressions as much as 45m below the surrounding bottom. These oval depressions are probably a product of glacial erosion (Shepard 1931).

GEOLOGICAL CHARACTERISTICS

Tectonic Setting

The tectonic setting of the area is similar to those of the Champlain and St. Lawrence Valleys and of the Esquiman Channel Area. Grenville crystalline rocks are exposed along north shore of the St. Lawrence Estuary and Appalachian rocks of Logan's Zone are exposed along the south shore. Therefore, the Appalachian boundary, Logan's Line, probably extends along this part of the channel (Geological Map of Canada 1969), but the feature is entirely hidden.

General Geology

In the section between the Saguenay outlet and Anticosti Island, the autochthonous (possibly partly parautochthonous) shelf rocks similar to those of the St. Lawrence Valley, probably occur between the Grenville rocks on the north shore and the presumed trace of Logan's Line, for in several localities along the north shore, small areas of Ordovician shelf rocks have been observed above the water level of the St. Lawrence Estuary (Faessler 1929, 1932, 1942). However, near the Saguenay outlet, there is no room for more than a 10 km strip of shelf rocks between the inferred trace of Logan's Line and the north shore of the St. Lawrence. The same is true further upriver as far as Quebec City and downriver for about 50 km or so. But, in the Sept Isles area where Ordovician shelf rocks are exposed along the north shore (Faessler 1942) there is room for about a 75 km strip of shelf rocks and at the down stream end of the channel the figure may be as high as 200 km.

The largest exposed area of shelf rocks is on Anticosti Island. There, the succession ranges from Early Ordovician to middle Silurian (Twenhofel 1928; Bolton 1972) and thickens southwards to a maximum of about 3300 m along the south shore of the island (Roliff 1968). As in the Champlain - St. Lawrence Valleys and in the Esquiman Channel Area, thick accumulations of arenite are not known to occur at the bottom of the shelf sequence in Anticosti Island. However, there are no sub-surface data from the top of the basement in the southern part of the island, and basal sandstones and conglomerates may be present in this area (see Roliff 1968, Fig. 8, section 5). Lower and Middle Ordovician rocks are mainly carbonates indicating the establishment of a carbonate shelf more or less synchronously with the establishment of similar shelves in the Champlain - St. Lawrence Valleys and in the Esquiman Channel Area. The carbonate shelves in these areas are most likely parts of a continuous shelf rather than separate shelves. Near the top of the Middle Ordovician, black shale (Macasty formation) appears and continues into the early Upper Ordovician. The appearance of black shale possibly indicating deeper water is also roughly synchronous with the appearance of flysch wedges in the Champlain - St. Lawrence Valleys and slightly later than the influx of flysch in the Esquiman Channel Area. However, quite unlike in the other areas, the stratigraphic record on Anticosti Island shows that a carbonate shelf was re-established in late Ordovician times and continued to exist at least into middle Silurian times. Because the Late Ordovician and Silurian record appears to be unbroken, one might expect to find, in the Anticosti succession, molasse derived from "Taconic Mountains" which are believed to have existed to the south, but none appears to be present. This is one of the puzzling aspects of Anticosti geology. Zen (1972, p. 69) cautions against ruling out the possibility that the sub-surface rocks in the southeast part of the island may contain Taconic structures. The same could be said of Taconic molasse wedges.

The onshore geology on the southwest side of the Gulf of St. Lawrence and in Madgdalen Islands indicate that downstream from Anticosti Island the channel may be cut, at least partly, in Permo-Carboniferous successor basin deposits (mainly terrestrial clastic beds with some marine formations, the latter including limestones and evaporite) which presumably overlie Early Paleozoic rocks in the area.

Tectonic Significance

Because the geology of the Laurentian Channel is largely hidden, views on its possible tectonic significance have to be founded mainly on arguments based on regional considerations and inductive reasoning. The inner part of the Channel, connects topographic lows (of the Marginal Segment) that are eroded on structures that appear essentially to be downfaulted rift zones. Therefore, the channel itself may be carved out along a similar structure. Such a supposition is compatible with the following regional considerations which suggest the tectonic unity of this 2000 km belt.

1. From one end of the belt to the other, its tectonic setting is along the edge of the craton. Throughout the evolution of the Appalachian geosyncline its paleotectonic role, appears to have been that of a fault-basin margin. Except for the puzzling differences observed in Anticosti Island the overall sedimentation characteristics also seem to have been similar along the belt. The early Middle Ordovician carbonate shelf in the area, for example, is remarkable for its lithic similarity from one end of the belt to the other (see Rodgers 1970, p. 150).

2. The St. Lawrence Valley high-angle faults and the Saguenay graben faults (see Plate I in pocket) appear to meet at an acute angle and seem to point in the direction of the

Laurentian Channel. It is unlikely that these two structures of regional extend stop after coming together; they probably continue as a single structure beneath the Laurentian Channel.

In addition, there is some field evidence in the area suggesting that a longitudinal high-angle fault system may be present along the channel. For instance, in the section of the coast in the Sept Isles area, Faessler (1942) has described a zone of faults and "fissures" parallel to the Laurentian Channel. The faults are probably of the normal type, because Faessler mentions that Ordovician shelf rocks have been downdropped into small grabens along these faults. Few faults were observed during geological mapping of Anticosti Island by Twenhofel (1928). But subsequent seismic reflection work on the island has revealed the presence of faulting, on a scale unsuspected before (Roliff 1968, p. 36). The faults are high-angle. According to structural sections (based on seismic work and drilling) compiled by Roliff, they are steep to near vertical normal faults (Fig. 18). The displacement pattern produced by them appears to be essentially one of step-faulting down from the shield side towards the foldbelt (Fig. 18), similar to the diplacement pattern in Champlain and St. Lawrence Valleys and possibly also in the Esquiman Channel Area.

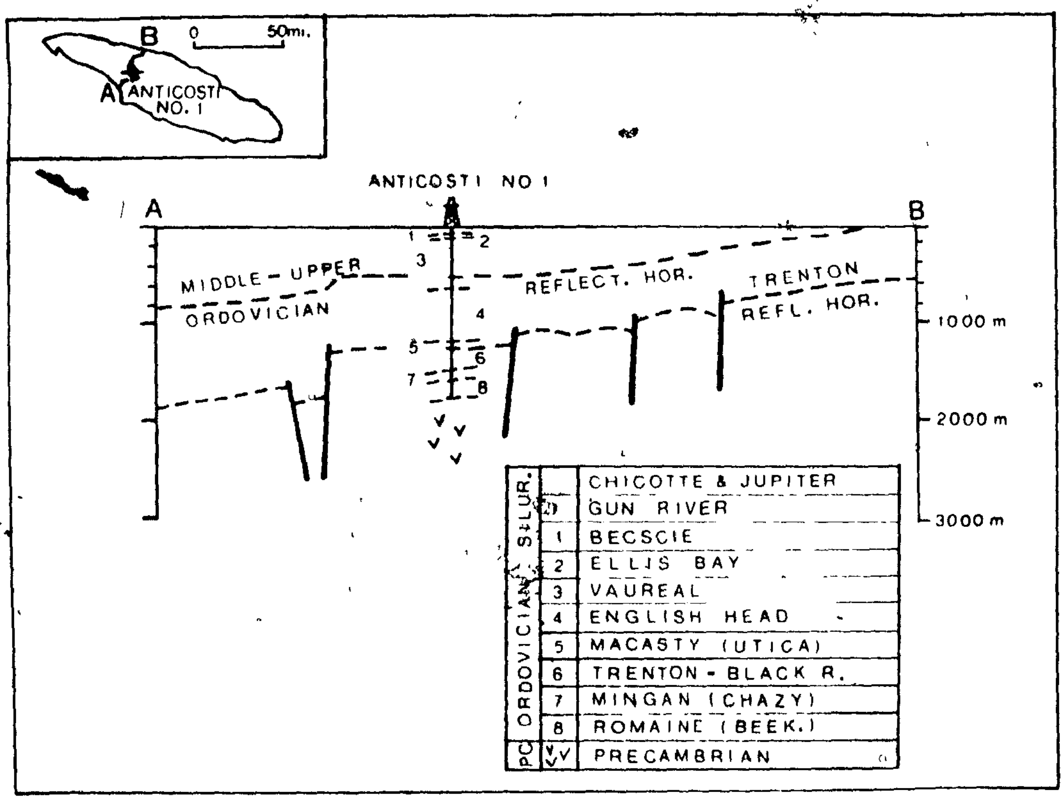

Fig. 18. Section across Anticosti Island, based on seismic reflection work carried out in connection with oil prospecting. After Roliff 1968.

CONCLUSIONS

Because the Inner part of the Laurentian Channel is largely hidden, nothing definitive can be said about its possible structure. However, from regional considerations and from evidence from Anticosti Island and the Sept Isles area, it appears that a rift zone similar to that of the other parts of the Marginal Segment may extend along the channel.

CHAPTER VIII: SHIELD SEGMENT A:

THE LOWER OTTAWA VALLEY

INTRODUCTION

Of the two branches into which the St. Lawrence Valley divides at its south end, the branch on the north side of the Adirondack Mountains is drained by the Ottawa and St. Lawrence Rivers and is referred to as the Lower Ottawa Valley (Fig. 19). A N-S line through the confluence of the Ottawa and St. Lawrence Rivers is taken as its eastern limit, and a line through Ottawa and Brockville is taken as its western limit. Thus, defined, the valley is about 125 km long and about 75 km wide. It has a slight arcuate trend (convex to the north) which on the average is about N75°E.

The Lower Ottawa Valley is a broad shallow topographic depression. Its plain-like floor is generally less than 100 m above sea level, and

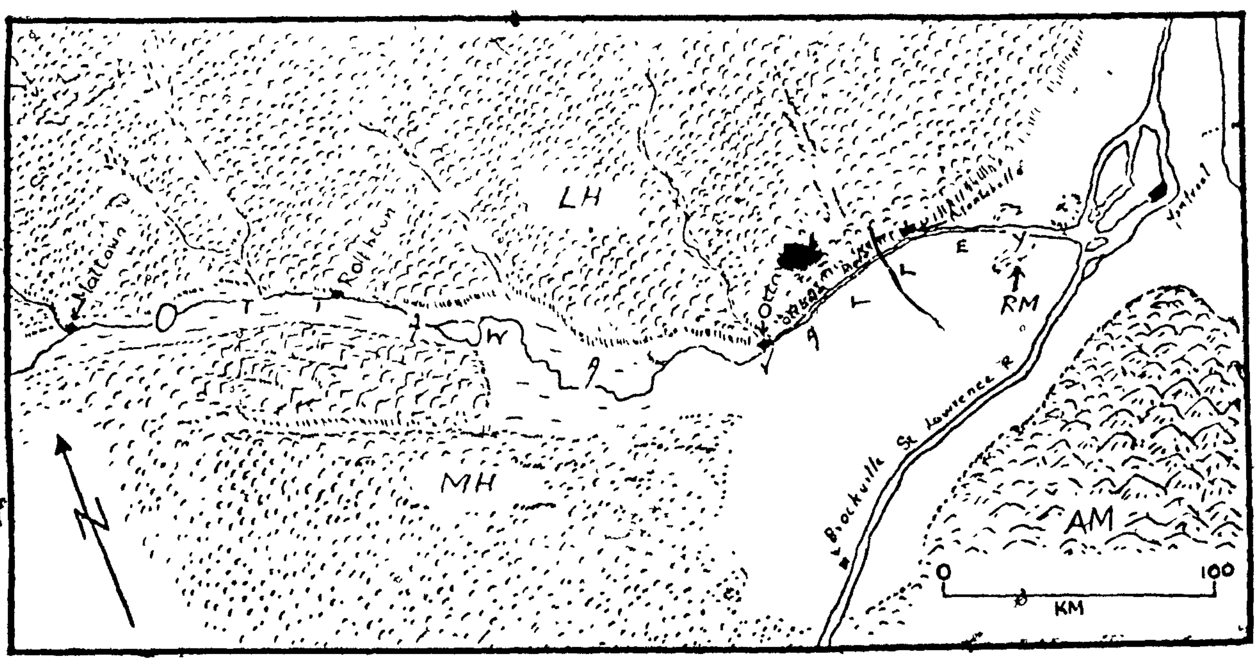

Fig.19. Physiographic sketch of the Lower and Upper Ottawa Valleys and vicinity. Key to abbreviations: AM-Adirondack Mountains; LH-Laurentian Highlands; RM-Rigaud Mountain.

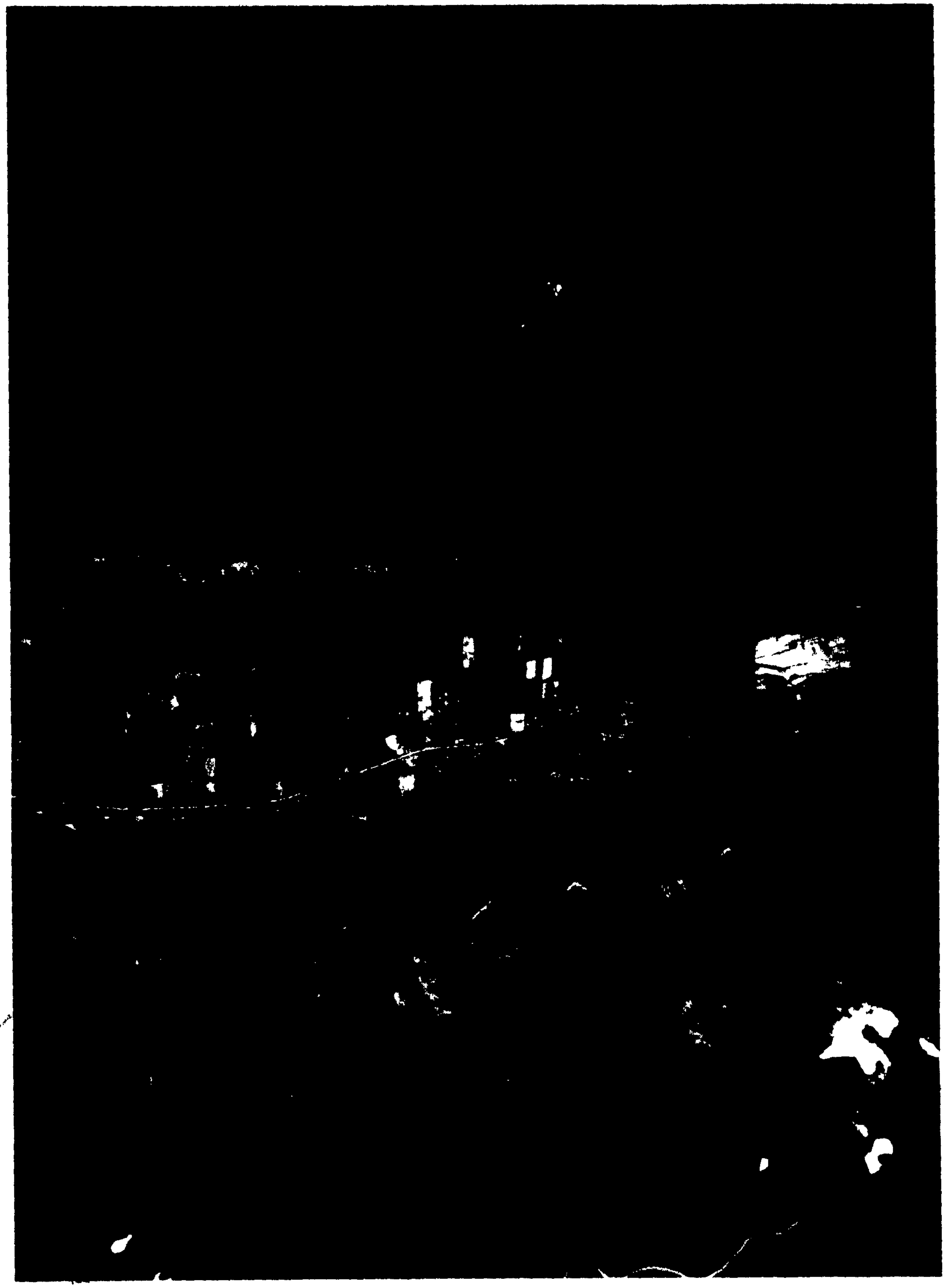

Fig.20. Vertical airphoto of a part of the Grenville scarp, about 10' 1 east of Montebello. Scale 1:50,000 approx. Note that the scarp transects the Grenville trends at large angles. National Airphoto Library, Ottawa, Photo No.A-12359-482.

the immediately adjacent areas of the Shield Region have altitudes in the range of 200 to 300 m. From the east end of the valley to about Montebello, its north boundary is an abrupt, linear scarp (Grenville scarp). From Montebello westwards the boundary is less sharply defined, but the rise to the general level of the Laurentian Highlands is rapid and takes place in about 10 km. On the south side, the boundary of the valley is rather indefinite, the rise towards the Adirondack Mountains being gradual.

GEOLOGICAL CHARACTERISTICS

General Geology

The Lower Ottawa Valley is in the southwestern part of the Grenville structural province. The dominant Grenville gneissic trends in the area are north-northeast. Because the valley extends approximately in an east-west direction, it is discordantly superimposed on the Grenville "grain" (Figure 20).

Much of the valley floor is carpeted by a sequence of Ordovician and possibly partly Cambrian platformal strata, similar to those of the St. Lawrence Valley but of lesser total thickness, the maximum being less than about 1000 m. The main deformational process that has affected these rocks is block-faulting. The faulting is mostly concentrated in the northern part of the valley (Fig. 21). In this part, the rocks are broken into numerous blocks which are tilted at varying angles. Elsewhere, the platformal rocks are flat or nearly so.

Several relatively small inliers of the Grenville basement emerge through the platform cover and generally form low hills. The larger inliers are in the Rigaud Mountain and St. André areas, at the east end of the valley (Fig. 21). On Rigaud Mountain, Grenville rocks occur only along its outer margin (except on the north side), the central part being made up of younger igneous rocks. These igneous rocks form a stock-like mass known as the Rigaud stock.

The Rigaud stock appears to be a ring complex with outer rings of syenitic rocks and a central core of granitic rocks (Greig 1968). The rocks are alkali-calcic in their petrochemical character, their alkali-lime index (see Peacock 1931) being about 53 percent (Greig 1968). About 30 km north-northwest of the Rigaud stock and located on the northern margin of the valley, is a similar intrusion, the Chatham-Grenville stock (Osborne 1934). Rocks from these two stocks date (K/Ar) around 450 ± 25 m. y. (Doig and Barton Jr. 1968, p. 1403).

Quaternary deposits cover much of the bedrock surface of the Lower Ottawa Valley. Their maximum thickness is a little less than 100 m. They consist of till, marine (Champlain) deposits, lacustrine and fluvial muds and sands.

High-Angle Faults

High-angle faults were first detected in the Lower Ottawa Valley by Ells (1900, pp. 99-120). However, only a few faults were known until Wilson's (1946) work, which revealed the numerous faults shown in Figure 21. Amongst these faults two main sets can be recognized. One set strikes approximately N75°E and is parallel to the valley. The other is oblique, its strike on the average being approximately northwest.

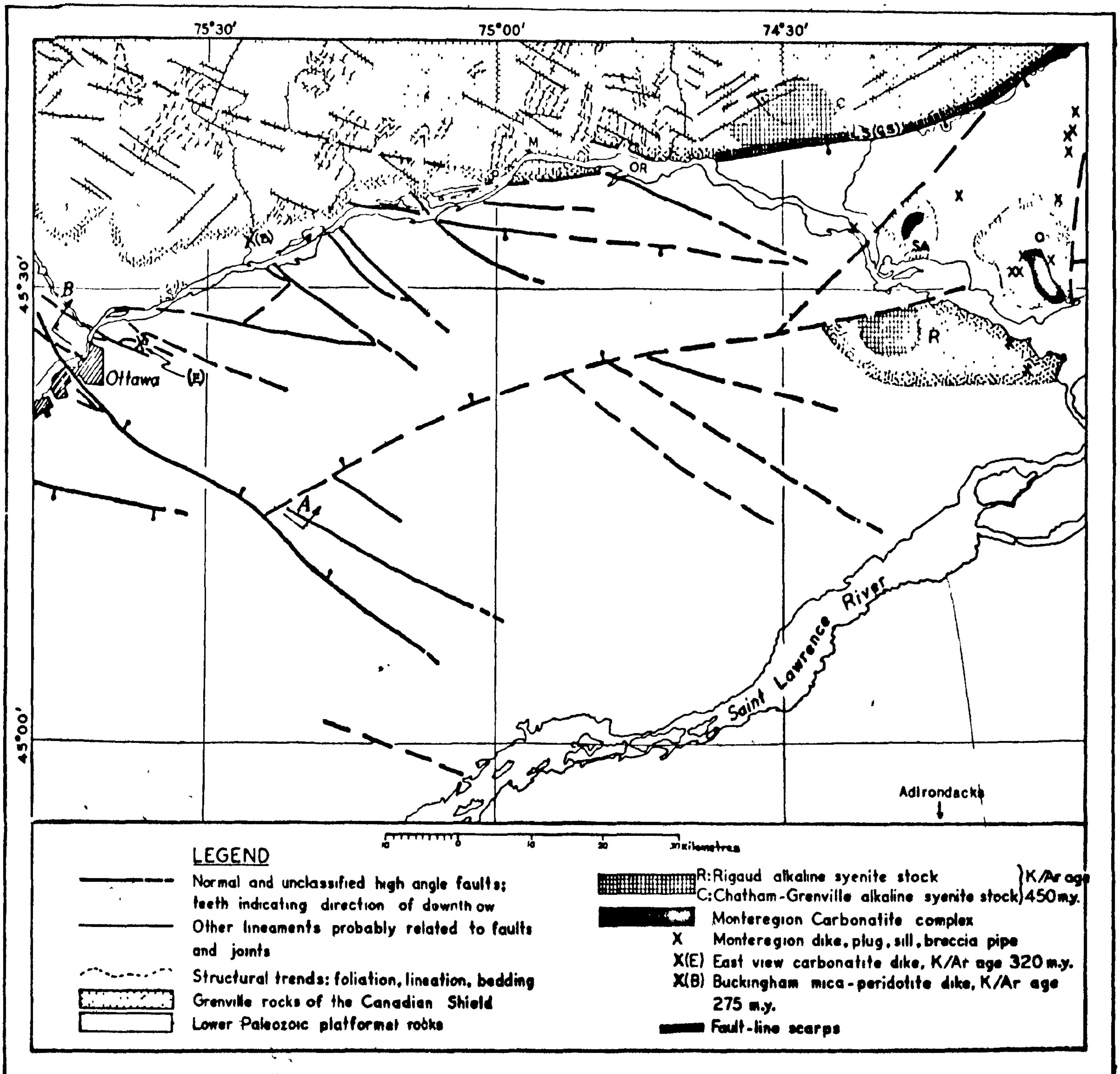

Fig.21. Map of the Lower Ottawa Valley area showing selected features. Key to abbreviations: B-Buckingham dikes; C-Chatham-Grenville stock; E-Eastview carbonatite; LS(GS)-Grenville scarp; M-Montebello; O-Oka complex; OR-Ottawa River; P-Papineauville; R-Rigaud; SA-St.André. Data from Wilson (1946) and other sources.

The fault traces themselves are concealed; most commonly they have been detected by the juxtaposition of rocks of different beds or formations. In a few places, however, their positions are indicated by down-dragged rocks (Wilson 1946). The faults are probably of the normal type because:

1. The displacement pattern produced by them is one of downfaulting; there is no evidence for large strike-slip movements.
2. Wilson (1946, p. 34) states that almost all the faults have their largest displacement about midway of their length. This is a common characteristic of normal faults (Hills 1965, p. 180).
3. Minor faults exposed on road cuts and other embankments (common in Ottawa area) are of the normal type.

Their throws in platformal rocks range from a few meters to a few hundred meters. The maximum throw estimated on a single fault is about 500 m (Wilson 1946, p. 34). The dip of the faults has not been observed, but because none of the wells in the Ottawa area intersect any faults, Wilson (1946) suggests that the faults are steep. The high salinity ($\sim$15,000 ppm of dissolved solids) of water in some of the many mineral springs (Elworthy 1918; Wilson 1946) in the Lower Ottawa Valley is probably caused by deep circulation of meteoric water along some of these faults.

Ages of High-angle Faulting

With regard to ages of high-angle faulting in the area, there are many uncertainities mainly because of large gaps in the stratigraphic record. The following evidence and their implications are pertinent to the age question.

1. The faults are post-Grenville and some of them cut the youngest Ordovician rocks (Upper Ordovician) in the area.
2. A well-defined, E-W trending diabase dike swarm, known as the Grenville dike swarm (Murthy 1971; Fahrig 1972, p. 576) cuts the Grenville rocks just north of the Lower Ottawa Valley and extends westwards along the Upper Ottawa Valley, Nipissing Depression to Lake Huron area, a total distance of about 650 km (See Fig. 54). In the Lower Ottawa Valley area, the dikes commonly trend E-W, approximately parallel to the longitudinal faults of the valley. The frequency of dikes increases nearing the valley margin and it is likely that dikes of this swarm are also present in the basement within the valley. The dikes are post-Grenville (Fahrig 1970) and are believed to be Hadrynian (Murthy 1971; Fahrig 1972, p. 576). Their absolute ages are uncertain however. K/Ar age determinations have yielded ages ranging from 400 m. y. to 974 m. y. (Wanless et al. 1967). The dikes are not

known to cut the Paleozoic platformal rocks in the area. The presence of diabase dike swarms in the crust is interpreted by most workers as indicating a tensional stress environment at the time of their emplacement (e.g. Fahrig and Wanless 1963). In the Lower Ottawa Valley area the indicated stresses are similar to those that formed the fault system (see p.116). The emplacement of these dikes, therefore, is the earliest post-Grenville event that can conceivably be taken as mutually connected with the faulting in the area.

3. The unusual thickness of the basal sandstone (Nepean) near the two main longitudinal faults (Wilson 1946, p. 33; Sobczak 1970, p. 161) suggests that these thick sand deposits accumulated in topographic depressions which were in some way related to faulting.

4. The longitudinal faults of the Lower Ottawa Valley extend eastwards and are continuous with the E-W trending faults in the southern part of the St. Lawrence Valley. In the latter area the faults are believed to contain Cretaceous movements because of their close space association with Monteregian igneous rocks (Kumarapeli 1970).

5. There are three dated minor igneous events (for localities of the related igneous bodies see Fig. 21) which may be related to fault movements in the area. They are the emplacement of (i) Rigaud and Chatham-Grenville stocks with K/Ar ages of about 450 m. y. along longitudinal faults which in turn are known to cut the Chatham-Grenville stock (ii) carbonatite dikes at Eastview just east of Ottawa with K/Ar ages of about 320 m. y. (Doig 1970, p. 23), (iii) mica peridotite dikes near Buckingham with K/Ar ages of 275 m. y. (Doig 1970).

6. The mild earthquake activity in the area (Smith 1967) may be related to seismic strain release on some of the faults.

Some of the above lines of evidence are admittedly equivocal. For example, the minor igneous events in the area need not have been accompanied by faulting. However, the following conclusions are compatible with the overall picture.

1. The initiation of faults and the emplacement of early dikes of the Grenville dike swarm may have taken place in the Hadrynian, as mutually connected events.

2. The faults since their initiation appear to have been reactivated several times. There is no reason to believe

that any or all of these reactivations were general; some may have been quite local. There is conclusive evidence of post-Ordovician movements on most of them. A major part of these movements may have taken place in the mid-Mesozoic, synchronously with the Monteregian igneous event. Some stress release may be taking place on these faults at the present time.

Structural Effects of High-angle Faulting

The faults in the northern half of the valley produce a well-defined asymmetrical graben (Fig. 22) about 25 km wide and involving the basement as well as the cover rocks. From its east end to about Papineauville, this graben is outlined on the north side by the Grenville fault which runs roughly along the base of the Grenville scarp. From Papineauville westwards, structural subsidence seems to have been achieved mainly by movements along oblique faults with flexures accommodating the subsidence to the southeast. The transverse as well as longitudinal sections of this graben are asymmetrical. The structure is deeper on the south and on the west sides. The gradual eastward rise of the basement is interrupted about midway under the graben by a slight reversal in inclination (Wilson 1946, p. 34). Thence, however, the gradual rise resumes and culminates along the Beauharnois axis where faulted inliers of the basement come to the surface in Rigaud and Oka Mountains and in St. André area.

The area between the graben described above and the St. Lawrence River is also downfaulted by play of movements on oblique faults.

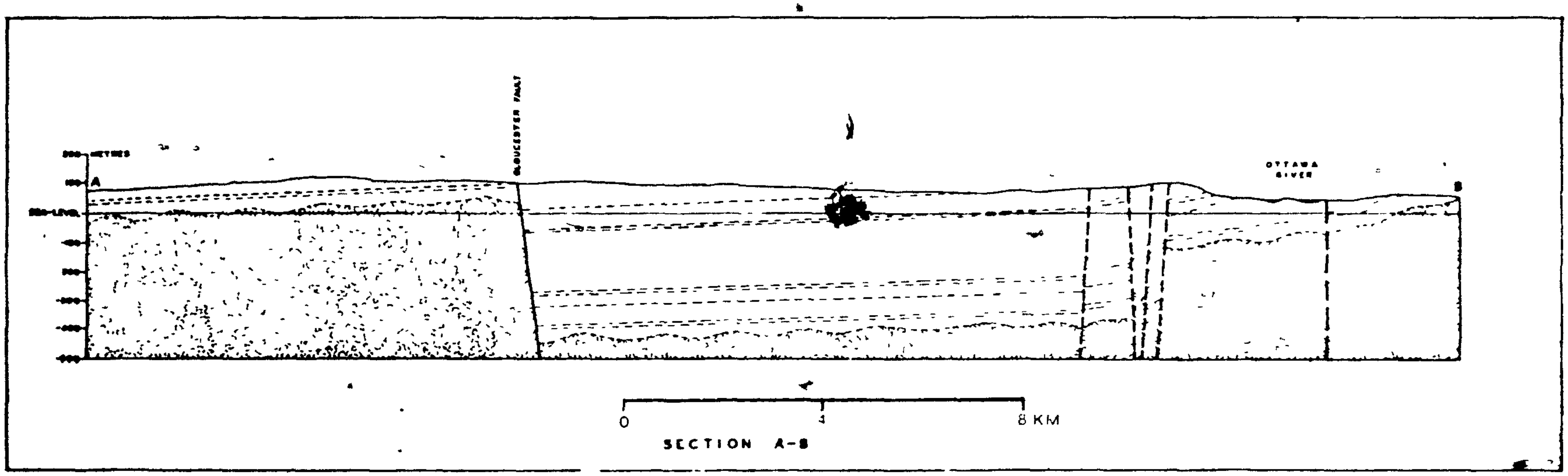

Fig. 22. Section across the Lower Ottawa Valley. After Wilson (1946). For location of section see Fig. 21. Unoriented dash pattern-Grenville rocks. Fine broken lines-bedding of platformal rocks. Thick broken and continuous lines-faults.

Longitudinal faults are not known in this area. The oblique faults seem to die out nearing the St. Lawrence River, possibly along a hinge zone. The overall structure along the valley appears to be a complex graben, about 60 km wide. The narrower graben in the north half of the valley is a sub-graben within the larger structure.

The stresses that produced this roughly E-W trending graben were probably horizontal tensional forces oriented approximately N-S. Apparently, there was partial release of stresses on northwest oriented oblique faults also. In the Grenville area to the north, one of the prevalent direction of airphoto lineaments is northwest and these lineaments, as discussed earlier, (see p.42) appear to reflect a regional fracture pattern. It is conceivable, therefore, that the oblique faults formed under the same stresses that produced the longitudinal faults but were controlled by a pre-existing northwest oriented fracture pattern.

In view of the earlier discussions on ages of faulting, the graben structure in the cover rocks is definitely post-Ordovician and may have formed largely in the mid-Mesozoic. But the basement structure is probably much older. It may have originated sometime in the Hadrynian, synchronously with the emplacement of the early dikes of the Grenville dike swarm and undergone later reactivations including some that led to the formation of the graben in the cover rocks.

SUMMARY AND CONCLUSIONS

The part of the Lower Ottawa Valley, north of the St. Lawrence River, is block-faulted and the overall structure of this block-faulted area appears to be a complex graben. The graben structure in the Cambro-Ordovician cover rocks formed in post-Ordovician times and possibly to a large extent in the mid-Mesozoic, synchronously with the early Cretaceous Monteregian igneous event. The post-Ordovician structure appears to have formed by reactivation of an older structure which also may have been a rift zone. The ancestral structure probably originated in the Hadrynian, synchronously with the emplacement of the Grenville dike swarm.

The grabens along the Lower Ottawa and Champlain Valleys appear to be branches of the rift zone along the St. Lawrence Valley, the acute angle ($\sim 70^{\circ}$) between the two branches being occupied by the horst bock of the Adirondack massif.

CHAPTER IX: SHIELD SEGMENT A: THE UPPER OTTAWA VALLEY

INTRODUCTION

In the vicinity of Ottawa, the northern part of the Lower Ottawa Valley bends clockwise through an angle of about 45° and continues westwards as the Upper Ottawa Valley as far as Mattawa, a distance of about 275 km (Fig. 19). Between Ottawa and Rolphton the trend of the valley is approximately N60°W, but a rather sharp change in direction to N80°W takes place around Rolphton. Thence westwards, the trend swings gradually to almost E-W near Mattawa.

Between Ottawa and Pembroke the valley is a nearly symmetrical trench about 55 km wide;on both sides it is bounded by scarps about 300 m high and flanked by highlands (Madawaska Highlands on the south side). West of Pembroke the appearance of more and more highlands within the valley, causes it to divide into two branches. The south branch is quite narrow (less than 10 km) and follows the basins of Golden and Round Lakes and the valleys of the Bonnechere and Petawawa Rivers, becoming increasingly narrow and ill-defined westwards. The north branch is wider but it also undergoes a progressive reduction in width and depth westwards, until in the vicinity of Bisset it is only about 15 km wide. In this area it assumes a markedly asymmetrical cross profile, being deepest along the course of the Ottawa River at the north margin of the valley. The width and form of the valley undergo little change between Bisset and Mattawa.

GEOLOGICAL CHARACTERISTICS

Tectonic Setting

The tectonic setting of the Upper Ottawa Valley is similar to that of the Lower Ottawa Valley. Both valleys are in the southeastern part of the Grenville Province which in the Paleozoic era was a part of the St. Lawrence Platform. Although the platform cover forms a more or less continuous carpet in the Lower Ottawa Valley, it has vanished from much of the Upper Ottawa Valley (Fig. 23). The Grenville trends in the general area are variable, but the valley extends with little regard to these trends.

Brief Description of General Geology

In the Upper Ottawa Valley bedrock exposures are common only in areas of higher elevations. In areas of lower elevations, bedrock is usually masked by a veneer of Quaternary deposits, consisting mainly of glacial deposits, marine (Champlain) clays and sands.

The platform cover that carpets the Lower Ottawa Valley extends northwestward into the Upper Ottawa Valley as far as Arnprior (Fig. 23). Further northwest, as far as about Pembroke several outliers of the platform cover are preserved as small downfaulted blocks (Fig. 23). Bedrock in the rest of the Upper Ottawa Valley is made up of Grenville crystalline rocks and some post-Grenville igneous bodies of minor areal extent, including dikes of the Grenville dike swarm.

The oblique fault pattern of the Lower Ottawa Valley appears to continue westwards to form a longitudinal fault system in the Upper

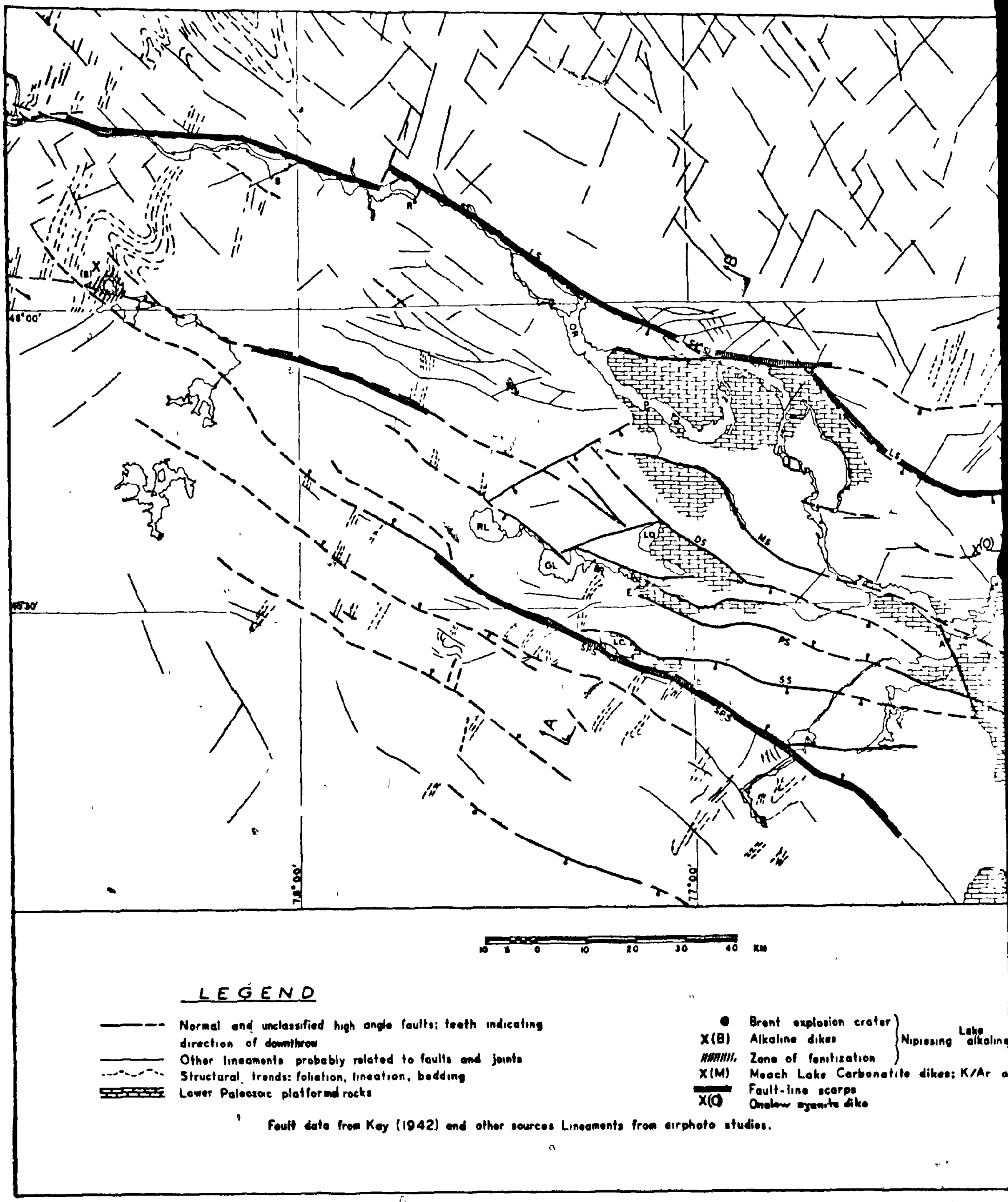

Fig.23. Map of the Upper Ottawa Valley showing selected features. Key to abbreviations: A-Arnp Crater; B-Bisset; BR-Bonnechere River; D-Doré scarp; E-Eganville; GL-Golden Lake; LC-Lake Clea LS-Laurentian scarp; LS(CS)-Coulonge scarp; LS(ES)-Eardley scarp; (M)-Meach Lake carbonatite d MS-Muskrat scarp; (O)-Onslow dike; OR-Ottawa River; P-Pembroke; PS-Pakenham scarp; Q-Quyon; R- Round Lake; SS-Shamrock scarp; SPS-St.Patrick scarp.

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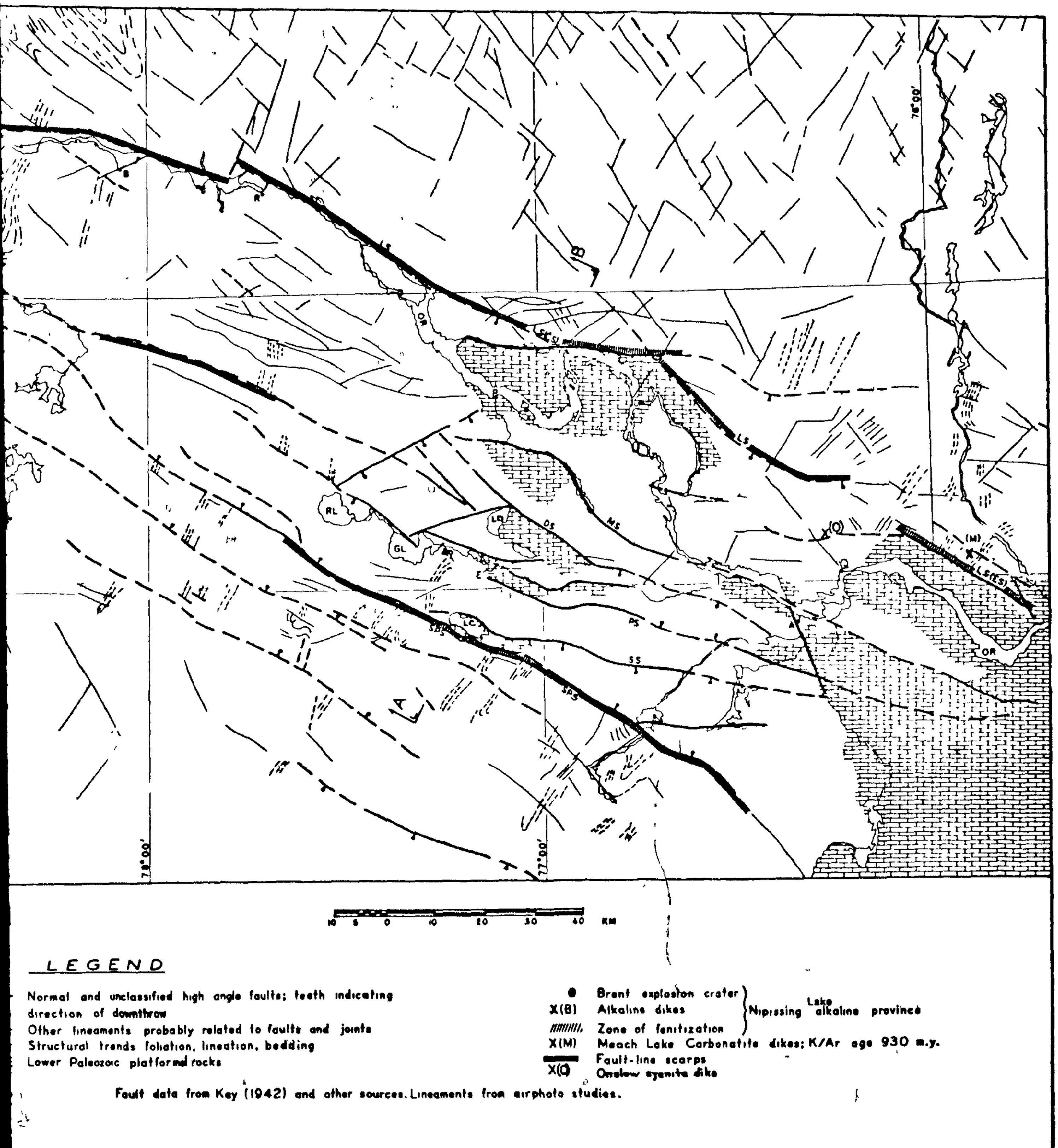

the Upper Ottawa Valley showing selected features. Key to abbreviations: A-Arnprior; (B)-Brent
t; BR-Bonnechere River; D-Doré scarp; E-Eganville; GL-Golden Lake; LC-Lake Clear; LD-Lake Doré;
scarp; LS(CS)-Coulonge scarp; LS(ES)-Eardley scarp; (M)-Meach Lake carbonatite dikes; MA-Mattawa;
rp; (O)-Onslow dike; OR-Ottawa River; P-Pembroke; PS-Pakenham scarp; Q-Quyon; R-Rolphton; RL-
-Shamrock scarp; SPS-St.Patrick scarp.

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Ottawa Valley area. These faults and a few transverse faults constitute the main post-Grenville structural elements in the area. The longitudinal faults dictate the valley trend which as mentioned earlier, consistently deviates from the Grenville trends in the general area (Fig. 23).

High-Angle Faulting

High-angle faulting in the Upper Ottawa Valley was recognized by early workers (Murray 1857, p. 95; Ells 1902, p. 14 J; Wilson 1924, p. 58) but it was Kay (1942) who for the first time, recognized the unusual intensity of faulting in the area (Fig. 23). Major faults are nowhere clearly exposed, but their presence is inferred from the following evidence.

1. Abrupt sub-~~parallel linear~~ scarps of crystalline rocks 100 to 300 m high and tens of kilometers long occur along the valley margins and also within the valley. The trend of Grenville rocks consistently deviates from that of the scarps. Therefore, the scarps are best explained as physiographic expressions of faultlines. The most prominent scarps are between Ottawa and Pembroke, where the Eardley and Coulonge scarps form the north wall and the St. Patrick scarp the south wall of the valley. These scarps are remarkable for their linearity and abruptness (Figs. 46 & 49) and have an amplitude of about 300 m. The Eardley and Coulonge scarps succeed one another in

en echelon fashion probably reflecting a similar fault pattern. Also between Ottawa and Pembroke, five cuesta-like longitudinal ridges divide the valley floor into relatively narrow (less than 10 km) strips. These ridges rise to a maximum height of about 100 m above the general level of the valley floor. Four of the ridges present south-facing scarps (the Muskrat, Doré, Eganville and Shamrock scarps), and the fifth (Pakenham) faces north. From Pembroke to Mattawa the Coulonge scarp continues as the north wall of the valley. Its height diminishes progressively westward, until it is no more than 100 m in the vicinity of Mattawa. On the south side of the valley, the St. Patrick scarp continues only for a short distance west of Round Lake before becoming lost.

2. Stratigraphy and structure of the several Paleozoic outliers along the bases of linear scarps show that the outliers are preserved in downthrown fault blocks.

Kay (1942) states that with few exceptions:

> "The Paleozoic rocks commonly dip towards the scarps, with their youngest representatives adjoining them... Approaching the scarps dips tend to steepen, precluding the possibility that the sediments are in synclinal basins; belts of outcrop are truncated by the pre-Cambrian outcrop; and drag is rather exceptional. The sediments in the outliers retain consistent primary characters irrespective of their proximity to the scarps; secondary dolomitization is common near the scarps, however. There is striking uniformity in the thin members such as Chaumont in separated outliers, a condition strongly opposed to an interpretation that they were originally laid in elongate separate basins."

3. The crystalline rocks of the linear scarps are characterized by vertical to steeply dipping slickensided surfaces and occasionally by brecciation and mineralization (Kay 1942, p. 623).

The faults appear to be steep to near vertical normal faults because slickensided surfaces prevalently dip towards the downthrown blocks at high angles. There is no evidence of large strike-slip movements. Minor strike slip components present locally on slickensided surfaces are probably due to adjustments between fault blocks. As mentioned earlier, the Paleozoic rocks commonly dip towards the linear scarps and approaching the scarps dips tend to steepen. Because the faults appear to be normal faults such "reverse drag" was probably caused by lack of support below the hanging wall when blocks moved apart under tensional forces.

Judging from the amplitude of the scarps, the boundary faults of the valley appear to have the largest vertical displacements. A minimum vertical displacement of about 300 m is indicated.

The directions of faults in the area conform with the "regional fracture pattern" as inferred from airphoto lineaments (Fig. 23). In this pattern northwest and northeast trending fractures are prevalent and the longitudinal and transverse faults are nearly parallel to these directions. As was discussed earlier (see p.40), the fracture pattern is probably older than the faults and the directions of faulting were controlled by those of the fractures (also see Kay 1942, p. 623).

It appears that the tensional stresses which produced the faults were so directed (probably nearly N-S) that stress release took place principally along the northwest trending fractures (which later evolved into longitudinal faults) but only to a minor extent on the northeast trending fractures.

Ages of High-angle Faults

The problem of assigning ages to fault-movements in the area is similar to that of previously described sections of the St. Lawrence Valley system. Pertinent lines of evidence are given below; they either give wide ranges of ages or are subject to various interpretations.

1. The faults are post-Grenville and some cut Middle Ordovician rocks. According to Kay (1942, p. 622), the primary character of sedimentary rocks in the Ordovician outliers persist without change irrespective of their proximity to the linear scarps and the thin members of Ordovician formations show striking uniformity in separated outliers. He suggests that the evidence is opposed to an interpretation that the post-Middle Ordovician movements are renewed movements on older faults. Actually what the evidence does indicate is that if older faults did exist, they did not have sufficient physiographic expression to cause changes in sedimentation.
2. The earliest post-Grenville record of a N-S directed tensional stress regime, similar to that required to

explain the high-angle fault pattern in the area, is the emplacement of the Grenville dike swarm (Fig. 54) which as mentioned earlier, is believed to have been emplaced in the Hadrynian. The density of dikes in the Upper Ottawa Valley area appears to be greater than in the Lower Ottawa Valley area.

3. Other post Grenville igneous and related events that may be mutually connected with fault movements in the area are as follows.

   a) The emplacement of the Onslow syenite dike (for location see Fig. 23) which is about 25 km long and has a K/Ar age in the range of 592 to 640 m.y. (Doig and Barton Jr. 1968),

   b) Intrusion of alkaline dikes (trachyte and ocellar monchiquite) in the western part of the valley with K/Ar ages of 558 to 576 m. y. (Currie and Shafiqullah 1967; Shafiqullah et al. 1968; Currie 1971a). These latter dikes are closely associated with the Brent explosion crater (Millman et al. 1960). Similar dikes become more common westwards and in Nipissing Depression they are associated with four alkaline-carbonatite complexes also with K/Ar ages of about 565 m. y.. Currie(1970) has included all these igneous bodies in the Lake Nipissing alkaline province (to be described later, see p.138). He also thinks that the Brent crater itself is a product of explosional igneous activity related to Lake Nipissing alkaline magmatism (Currie 1971a) Another view is that the crater is the result of a meteorite impact (Beals et al. 1963; Dence 1968, pp. 171-176). Accordingly, its origin is the subject of a contemporary controversy. A discussion of this controversy is deferred until later (p.242), but for the present it may be mentioned that Currie's (1971a) volcanic explosion hypothesis is favoured by (i) the close association of the Brent crater with one of the principal faults of the Upper Ottawa Valley (ii) its geographical setting within the Nipissing alkaline province (iii) occurrence near the crater of alkaline dikes with petrography, petrochemistry and absolute ages identical to those of the products of Lake Nipissing alkaline magmatism.

c) Deposition of post-Ordovician Pb (Zn)-Ag-barite (calcite; fluorite) veins which occur in Grenville carbonate host-rocks within the Upper Ottawa Valley as well as in the area of Madawaska Highlands to the south (Sangster 1970, p. 34) which is also severely block-faulted by the same general fault system (Fig. 23). It is likely that these veins are genetically related to the faulting in the area. Barite-calcite-fluorite veins in many block faulted ore districts are believed to be related in a physico-chemical sense to alkaline-carbonatite magmas in a manner analogous to the relationship between granitic magmas and quartz, pegmatite and aplite veins (Kuellmer et al. 1966, p. 353). Because the most extensive episode of post-Ordovician alkaline-carbonatite magmatism in the general region is the Cretaceous Monteregian igneous event, it is possible that the veins represent a manifestion of the Monteregian alkaline magmatism in the area.

4. The faults of the Upper Ottawa Valley appear to be continuous with those of the Lower Ottawa Valley (and also with faults of the Nipissing and Timiskaming Depressions, to be described later, see Chapter X and XI) and the chronology of faulting in this entire belt may be broadly similar, although in detail it may differ from one part to the other.

5. The indications of seismic strain release on the St. Lawrence Valley and Lower Ottawa Valley faults, applies equally well to the faults of the Upper Ottawa Valley area.

The following conclusions appear to be consistent with the above lines of evidence.

1. The faulting initiated at the same time as in the Lower Ottawa Valley, probably related in time and cause with the emplacement of the Grenville dike swarm, in the Hadrynian. The lack of stratigraphic evidence for the existence of pre-Middle Ordovician faults in the area is not a serious objection to the above conclusion.
2. The faults since their initiation appear to have been reactivated, at least once after the Middle Ordovician. Minor stress release may be taking place on these faults at the present time. The linear scarps in the area seem to have been produced largely by post-Middle Ordovician movements. A major part of these movements may have taken place during the mid-Mesozoic more or less synchronously with the Monteregian magmatism in St. Lawrence Valley area.

Structural Effects of Normal Faulting

The high-angle faults shown in Figure 23, outline a graben structure along the Upper Ottawa Valley (Fig. 3 of Kindle and Burling 1915; Kay 1942) - the Ottawa-Bonnechere graben of Kay - and break up the Madawaska Highlands area into a series of tilted horst-blocks. This graben structure is actually the continuation of the graben along the Lower Ottawa Valley, and the two structures together are generally referred to as the Ottawa graben (Wilson 1959, p. 316). Between Ottawa and Pembroke where the valley is a nearly vertical-walled trench, the graben structure is well-defined and only slightly asymmetrical, the structural subsidence along the north boundary being somewhat greater (Fig. 24). The valley floor in this area is broken by longitudinal faults into several blocks which are tilted. The cuesta-like ridges within the valley appear to represent the uptilted edges of these blocks. In the area, south of Pembroke, the longitudinal fault blocks in turn are broken by cross-faults to produce rhombic patterns.

Westward from Pembroke, the north boundary fault zone seems to assume more and more the role of key fault of the graben with the south boundary fault-zone possibly assuming the role of a hinge fault. In this area, the graben appears to show much greater asymmetry with greater structural subsidence again on the north side. The morphotectonic relations in this area are probably as shown schematically in Figure 25. According to this scheme, the asymmetrical (topographically) north branch of the valley is located largely over the dip slope (or the back slope) of the tilted graben block; the highlands between the two branches of the valley are located over the uptilted edge of this block; the south branch of the valley lies along the south boundary fault-zone and may actually be located over a narrow fault trough produced by play of antithetic faulting (consequent to hinge faulting and fissure formation). The preservation pattern of Paleozoic outliers in the Upper Ottawa Valley indicates that the structural subsidence increases progressively from northwest to south-east along the structure. The structural subsidence along the Lower Ottawa Valley structure on the other hand increases from east to west. Therefore, the Ottawa graben as a whole is cradle-shaped.

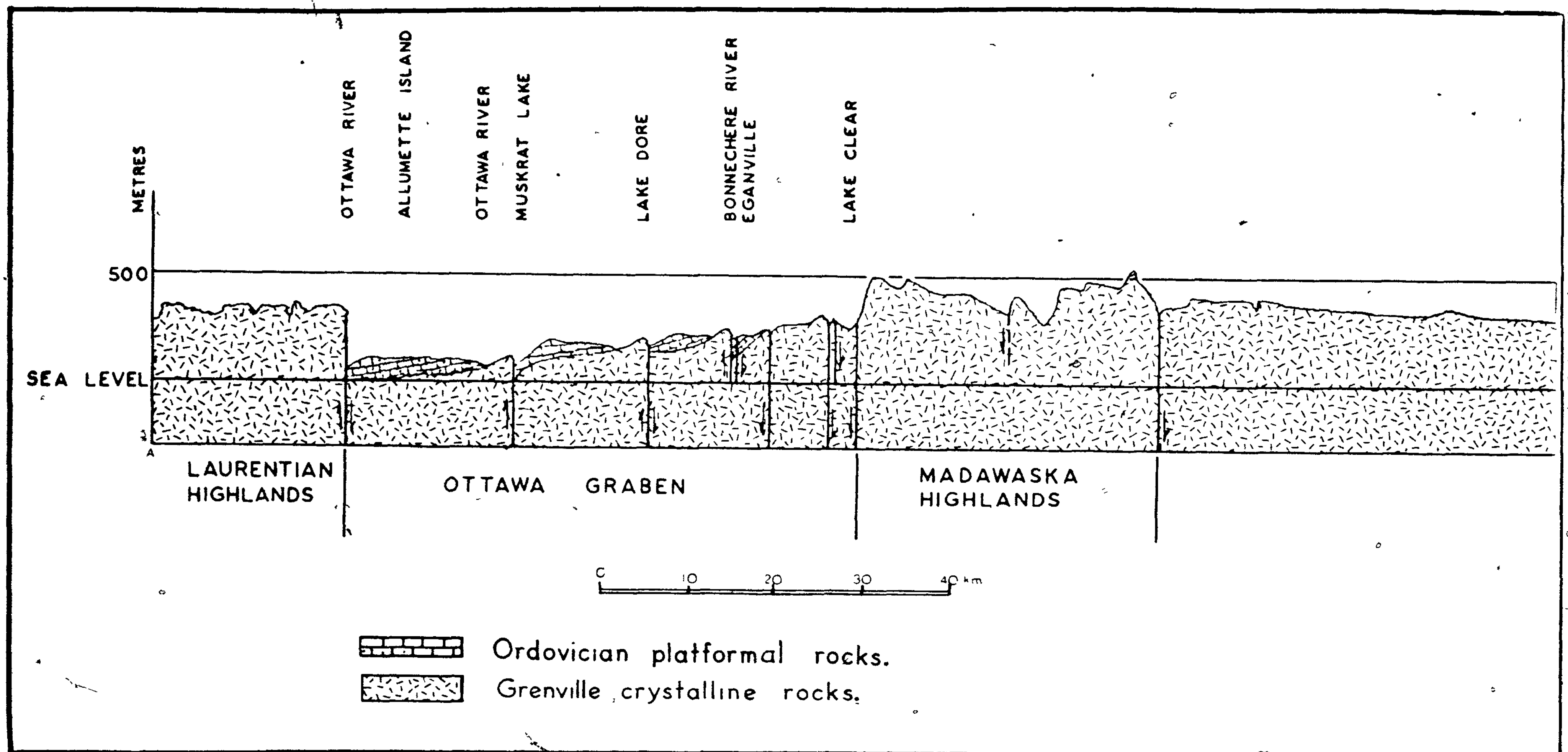

Fig. 24. Structure section across the Upper Ottawa Valley. After Kay (1942). For section location see Fig. 23.

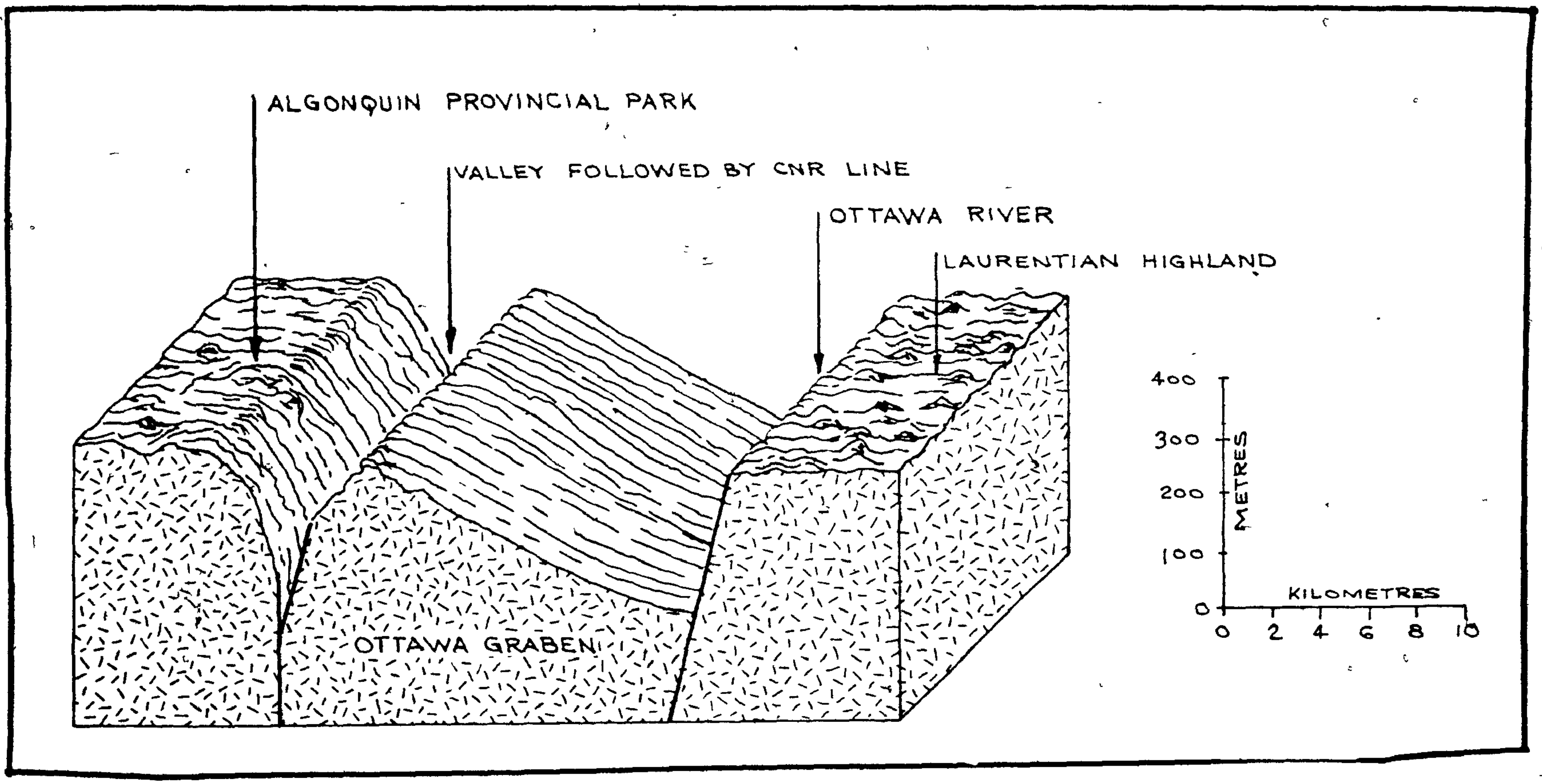

Fig. 25. Diagram illustrating morphotectonic relations of the western part of Upper Ottawa Valley.

In view of the earlier discussions on ages of fault movements in the Lower and Upper Ottawa Valleys the following conclusions can be made regarding the age and development of the Ottawa graben.

1. The Ottawa graben is a post-Grenville structure which appears to have formed as a result of N-S directed horizontal tensional forces but partly controlled by a prevalently northeast and northwest trending "regional fracture pattern".
2. The present structure includes post-Ordovician structural subsidence.
3. The earliest post-Grenville record of a N-S directed horizontal tensional stress regime, similar to the one required to explain the formation of the graben is related to the emplacement of the Grenville dike swarm in the Hadrynian. The initiation of this dike swarm and of the graben may have been mutually connected events.
4. The post-Ordovician graben subsidence may have taken place, largely in the mid-Mesozoic, synchronously with the Monteregian igneous activity. Also, several other minor episodes of alkaline - carbonatite activity, apparently of late Precambrian and Paleozoic ages, occur along the graben structure and may be related to re-activations of the structure, perhaps of local extent.

PHYSIOGRAPHY AND STRUCTURE

In the Upper Ottawa Valley area, most if not all, of the known faults were inferred initially on physiographic expression (linear, scarps) and later supported with geological evidence (e.g. see Kay 1942). This excellent correlation between geological and physiographic evidence for faulting, seems sufficient justification to warrant confidence in purely physiographic evidence for faulting in the area when geological evidence is not available. The exhumation of the basement topography in the area probably took place relatively recently (Ambrose 1964) and it appears that sufficient time has not elapsed to cause any large scale effacement by erosion of basement topography related to faulting.

SUMMARY AND CONCLUSIONS

The Upper Ottawa Valley is underlain by a complex graben - the Ottawa-Bonnechere graben of Kay (1942). This structure together with the graben along the Lower Ottawa Valley (see p.114) form the Ottawa graben which is about 400 km long and 30 to 65 km wide.

The Ottawa graben is discordantly superimposed on Grenville structural trends (bedding, lineation, foliation) but the directions of graben faults are thought to be largely controlled by a regional fracture pattern which is prevalently oriented northwest and northeast. The graben appears to be cradle-shaped longitudinally, the deepest part being near Ottawa. There is conclusive evidence that structural subsidence along the graben took place in post-Ordovician times. It is thought that a large part of this structural subsidence took place in the mid-Mesozoic, synchronously with the alkaline-carbonatite activity (Monteregian) at the east end of the graben. There are indications, however, that the post-Ordovician movements took place along an older structure of tensional origin which also may have been a graben. It is proposed that the origin of the ancestral structure was related to the emplacement of the Grenville dike swarm in the Hadrynian. The graben and the dike swarm are closely related in space.

CHAPTER X: SHIELD SEGMENT A:

THE NIPISSING DEPRESSION

INTRODUCTION

The topographic low of the Upper Ottawa Valley continues westwards past Mattawa to include the Lake Nipissing basin and this continuation is referred to here as the Nipissing Depression (Fig. 26). West of Lake Nipissing, the depression can be traced through the French River valley system into the North Channel of Lake Huron for a total distance of over 200 km.

Between Mattawa and Bonfield, the morphology of the depression is much the same as that of the adjacent parts of the Upper Ottawa Valley (Fig. 25); it is a composite topographic low about 25 km wide, consisting of two closely spaced parallel branches separated by a strip of highlands. The south branch is generally less than 5 km wide. Its floor lies about 150 m below the adjacent highlands and is partly occupied by a series of water courses and lakes. The Canadian National Railway line follows this low ground. The north branch has a highly asymmetrical cross-profile and is bounded on the north side by a south facing scarp which is the westward continuation of the Coulonge scarp. This scarp itself is no more than 50 m high but to the north of it elevations rise another 200 m or so within a short distance.

Just west of a N-S line through Bonfield the strip of highlands

Fig. 26. Map of the Nipissing and Timiskaming Depression areas. Black: lakes and other water courses. Orange: the depressions as outlined by the 300 m (generalized) contour. From National Topographic Series maps (1:500,000) 31NW and 41NE.

between the north and south topographic lows dies out and the depression ceases to be composite. Thence, it continues westwards as a shallow topographic depression, poorly defined except on the north side. In the area of lake Nipissing, the depression widens considerably (Fig. 26). West of the lake, further widening of the depression is accompanied by the tendency of its boundaries to be indistinct, and the depression as a whole to become indefinite, except for the narrow depression extending westwards through the French River system.

GEOLOGICAL CHARACTERISTICS

General

The Nipissing Depression, like the Ottawa Valley, is largely set in the Grenville province. Grenville rocks in the area make up a plutonic igneous-metamorphic complex with variable trends but the depression, for the most part, is superimposed discordantly on these trends.

The unity of the Nipissing Depression with the Upper Ottawa Valley is not restricted to their morphology but applies to a whole range of post-Grenville geological events.

1. The Nipissing Depression lies along a zone of severe block faulting (Lumbers 1971, pp. 65-68) which appears to be the westward continuation of the Ottawa Valley high-angle faulting.

2. The Grenville dike swarm continues into the Nipissing Depression area (see Fig. 54) and conforms well with the trend of the depression.
3. The alkaline dikes of the western part of the Upper Ottawa Valley develop into a regional dike system in the Nipissing Depression area (Lumbers 1971, p. 54) and are associated with at least four alkaline-carbonatite centres.
4. Middle Ordovician platformal rocks (sandstones, limestones, shales) occur as small outliers in the Nipissing Depression also, (Lumbers 1971, pp. 54-56) indicating that the Grenville rocks of this area too were once buried under Paleozoic strata but are now almost completely exhumed.

The bedrock surface, especially in the low laying areas of the depression floor, is usually buried under a veneer of Quaternary deposits composed mainly of till and post-glacial lacustrine deposits.

Alkaline-Carbonatite Intrusives: Lake Nipissing Alkaline Province

The main alkaline intrusive centres are central complexes of mafic alkaline rocks and carbonatite, four of which are known (Fig. 27). They are the Iron Island, Manitou Island (Heinrich 1966, pp. 386-389), Callandar Bay (Currie and Ferguson 1971, pp. 498-517; Ferguson and Currie 1972) and Burritt Island (Lumbers 1971) complexes. The Brent crater would be a fifth complex located in the same general area if Currie's endogenic view (Currie 1971a) of its origin is correct.

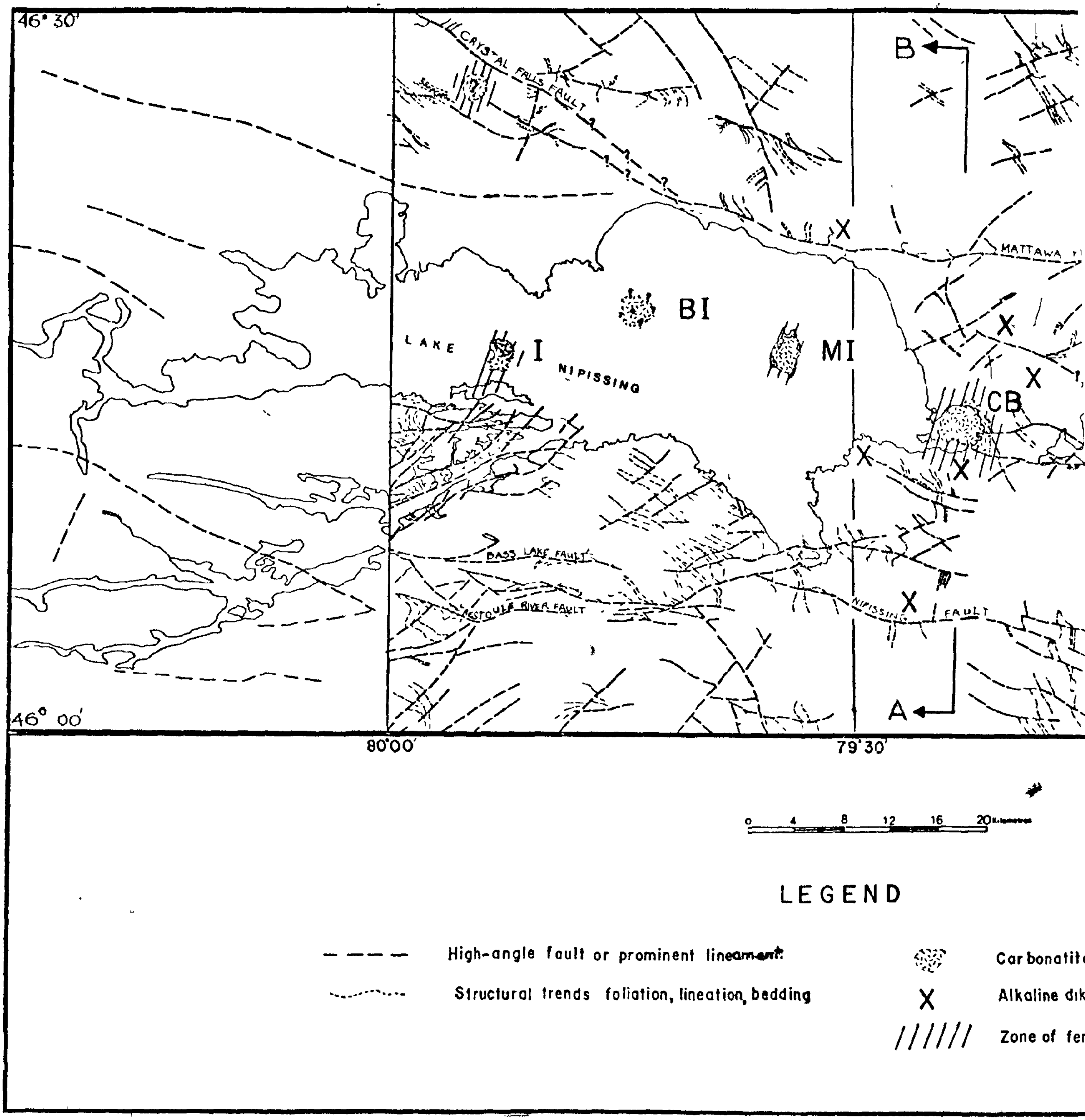

Fig.27. Map of the Nipissing Depression area showing selected features. Key to abbreviations: B-Bonfield; BI-Burritt Island; CB-Callander Bay; I-Iron Island; MA-Mattawa; MI-Manitou Island; OR-Ottawa River. Data from Lumbers 1971 and other sources.

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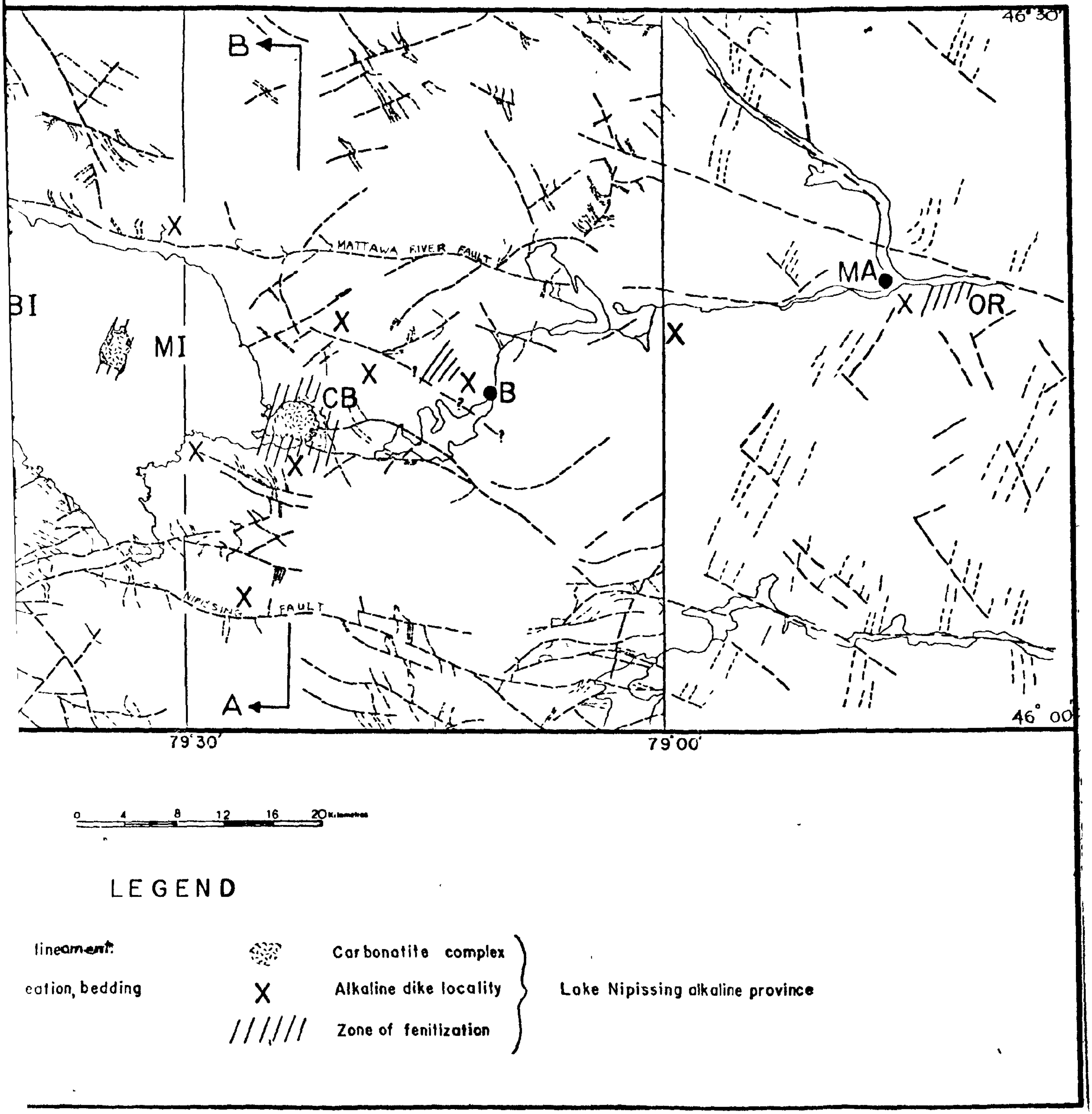

res. Key to abbreviations: B-Bonfield;
·lanitou Island; OR–Ottawa River.

Lumbers (1971, p. 47) suggests that another complex may be present approximately at Lat. 46°26 1/2' and Long. 79°48 1/2' (Fig. 27). Numerous dikes of related rocks (mainly lamprophyres) and patches of fenitization are present around the central complexes. Lamprophyre dikes similar to those associated with alkaline-carbonatite complexes are present elsewhere in the depression area, and also as mentioned earlier in the adjacent parts of the Upper Ottawa Valley, and form a regionally developed dike system. Currie (1970) has included all these igneous bodies in a petrographic province which he calls the Lake Nipissing alkaline province (Currie 1970, p. 411).

The Lake Nipissing alkaline province, as such, has been recognized only recently (Currie 1970). It is likely that both its geographical extent and its igneous body distribution are only imperfectly known. Even the known central complexes are poorly exposed. Natural exposures of dikes are also rare mainly because of their recessive weathering. Currie (1970) points out that virtually all the exposures of dikes that have been observed are in recently blasted road cuts.

The Manitou Island complex is the best known, largely because it has been intesively prospected for niobium ore deposits. The complex is a roughly elliptical ring structure of alkaline rocks, carbonatite and fenite, about 3 km long (north-south) and 2 1/2 km across (Rowe 1958, pp. 44-65; Heinrich 1966, pp. 386-389). It intrudes Grenville rocks (hornblende-granite-gneiss) and is overlain in places by Ordovician strata.

The dike rocks of the Lake Nipissing alkaline province are mostly lamprophyres that are chemically similar to monchiquite. Thin dikes of pale pink carbonatite and potash-rich chocolate-brown trachyte also occur (Currie and Ferguson 1970, p. 526). Cross-cutting relationships indicate that all three types of dikes are roughly the same age (Currie and Ferguson 1970, p. 526).

Biotite from carbonatite of the central complexes yielded a mean K/Ar age of about 565 m. y. (Lowden <u>et al</u>. 1963; Gittins <u>et al</u>. 1967; Shafiqullah <u>et al</u>. 1968) indicating that they formed around the beginning of the Cambrian period. Some of the trachyte dike rocks from around the Brent crater have yielded younger ages of about 310 to 450 m. y. (Shafiqullah <u>et al</u>. 1968; Hartung 1968). Currie (1971a p. 496) suggests that these lower ages may actually be due to argon loss from potash feldspar which was the main source of K^{40} from trachyte. Alternatively, all dikes of the area need not be of the same age.

<u>High-angle Faulting</u>

The use of physiographic evidence to dilineate faults in the Upper Ottawa Valley was already discussed (p.133). Lumbers (1971, p. 65) has shown that in the Nipissing Depression too, prominent lineaments (scarps, linear valleys) coincide with faults. He found the following features in rocks exposed on or near the lineaments:

1. displacement of geological contacts
2. major discordancies in foliation trends

3. zones of mylonitization, brecciation and shearing
4. locally intense hematization
5. quartz vein networks.

The high-angle faults of the Nipissing Depression and Upper Ottawa Valley, although described separately in this thesis should be regarded as parts of a continuous fault system. The fault pattern of the former area is probably as shown in Fig. 27.

The faults can be grouped into two sets according to their dominant strike directions:

1. a major set trending west-northwest; the strike directions of this set actually range from west to northwest and is nearly parallel to the trend of the depression.
2. a lesser set trending northeast, the strike directions of which varies from north-northeast to east-northeast.

Presumed fault traces are nowhere clearly exposed and fault planes have not been observed. The character of faults, therefore, has to be inferred on indirect evidence. Lumbers (1971, p. 65) points out that the associated mylonitization zones, slickensided surfaces and breccia zones dip near vertically, although locally as low as 60°, in a manner suggesting that the faults are high-angle normal faults.

Although some of the faults strike nearly in an E-W direction and conform with the trend of the depression, the majority of faults in the area tend to conform to two prevalent strike directions, oriented approximately northeast and west-northwest. This tendency is interpreted as reflecting a control from a pre-existing "regional fracture pattern" (see p. 42), also oriented prevalently northeast and west-northwest. If the faults are normal faults, as they appear to be, then the fault pattern could have been produced by a horizontal tensional stress field oriented approximately N-S so that the main stress release took place along west-northwest oriented fractures in the area, and minor stress release and block adjustments occurred along northeast oriented fractures. That the west-northwest fractures opened up into fissures is indicated by the fact that some faults of the west-northwest set are intruded by mafic dikes. The northeast set on the other hand is not associated with dikes (Lumbers 1971, p. 67).

Ages of High-angle Faults

The faults are definitely post-Grenville but apart from this the chronology of faulting in the area is poorly known. Some insight into the age question may, however, be gained from the following evidence.

1. The Grenville dike swarm extends through the area and nearly conforms with the west-northwest trending longitudinal set of faults. At least some of the dikes occur along faults (Lumbers 1971, p. 46). These characteristics indicate that the faulting and dike formation may be chronologically related. As mentioned earlier the diabase dike swarm appears to have been emplaced in the Hadrynian.

2. A few dikes are sheared and brecciated by faulting (Lumbers 1971, p. 47). Therefore, some fault movements have occurred after the emplacement of the dikes.
3. Most of the northeast faults are diplaced by west-northwest faults although the reverse relationship has also been noted in one locality (Lumbers 1971, p. 67). However, it is likely that both sets of faults formed during the same general period of faulting.
4. The Lake Nipissing alkaline magmatism (K/Ar age 565 m.y.) in the area may have been accompanied by faulting. In fact, faults of the northeast system appear to be mostly concentrated in a broad zone trending northeast to include the area in which the alkaline complexes are located. It is possible that the complexes are localized at intersections of faults of the northeast and west-northwest sets.

According to the above evidence, it appears that faults probably originated in the Hadrynian, synchronously with the emplacement of the Grenville dike swarm and have since been reactivated, at least once around the beginning of the Cambrian. Although there is conclusive evidence for post-Ordovician movements in the Ottawa Valley area, there is no clear evidence for such movements in the Nipissing Depression. However because the faults in the two areas appear to be parts of a continuous fault system, it is likely that post-Ordivician movements occurred in the Nipissing Depression also; the preservation of Ordovician outliers in both areas may be in downfaulted blocks.

Structural Effects of High-angle Faulting

The structural effects of high-angle faulting in the Nipissing Depression area cannot be evaluated with certainty from local evidence alone. The available evidence together with regional considerations, however, indicate that the overall structure along the depression is essentially a graben (Fig. 28).

The Coulonge scarp which in the Upper Ottawa Valley area forms the north boundary scarp of the Ottawa graben, can be traced west of Mattawa as a continuous south-facing scarp along the north side of the Nipissing Depression. Slickensided surfaces accompanied occasionally by brecciation and mineralization, appear to be characteristic of the crystalline rocks of this scarp, indicating that the scarp in this area too, is related to a fault zone (Mattawa River Fault, Crystal Falls Fault, see Fig. 27). The physiographic relations and the evidence from slickensided surfaces suggest that the indicated fault zone, like the Coulonge Fault, has its south side downthrown. Westwards from Mattawa the scarp diminishes in height, possibly reflecting a similar change in the throw of the fault zone. On the south side of the depression, a prominent zone of <u>en echelon</u> faults (Bass Lake, Restoule River and Nipissing faults; see Fig. 27) appears to be present (Lumbers 1971, p. 67). This fault zone can be traced east-southeast until it merges with the south boundary fault zone of the Ottawa graben. In the Nipissing Depression area, the displacement pattern on this fault zone is not known, although in the case ~~of~~ the Nipissing fault, Lumbers (1971, p. 66) thinks that the preservation pattern of Paleozoic rocks on the northside of the fault suggests that its north side is downthrown. Thus, despite uncertainities, the available evidence indicates that the overall structural effect of high-angle faulting in the area is structural subsidence. This structurally depressed zone is a continuation of the Ottawa graben is suggested by: (i) the continuity of the morphototectonic features of the Ottawa graben into the Nipissing Depression (ii) the continuity, from one area to the other, of high-angle faults, the Grenville dike swarm and of the products of the Lake Nipissing alkaline province. In the Nipissing Depression area, the structure will be referred to as the Nipissing graben (Fig. 28). It is not known how far west of Lake Nipissing the structure continues but the following features are pertinent to the question.

1. Except for the relatively narrow topographic low along the French River, the main topographic low and the associated lineaments die out westwards and become indistinct features about 40 km west of the lake, suggesting that the structure itself begins to die out in this general area.
2. West of the lake, the depression widens. Also the fault system as well as the Grenville dike swarm fans out.

These features suggest splaying out of the structure. Splaying out characteristically occurs where graben structures die out (de Sitter 1964, p. 124).

3. The Grenville dike swarm which is spatially related to the Ottawa-Nipissing grabens is not known to extend west of the North Channel of Lake Huron (See Fahrig 1970, p. 132).

The above features indicate that the Nipissing Depression fault system terminates somewhere in the area of the North Channel of Lake Huron. This is contrary to the view expressed earlier by Kumarapeli and Saull (1966a) that the Nipissing Depression faults continue westwards through the Lake Superior area and further southwest into the continent.

Lumbers (1971, p. 69) states that the Crystal Falls Fault can be traced westwards, as a lineament to within a few kilometers of the Grenville Front and that it is on strike with the Onaping Lineament (see Church 1972, p. 355) and that the Nipissing-Bass Lake-Resolute River fault zone can be traced through French River and across the Grenville Front to merge with the westerly striking Murray Fault system (see Church 1972, p. 353) which extends along the north shore of Lake Huron. The Murray Fault system and the Onaping lineaments (Church 1972) are probably much older structures than the Nipissing Depression fault system. The apparent continuity of the younger fault system with the older faults is thought of by the writer as due to terminal control of the former by the latter.

In view of the earlier discussions on ages of high-angle faults, the following conclusions can be made on the age and development of the Nipissing graben.

1. The Nipissing graben is a post-Grenville structure which developed as a part of the Ottawa graben, probably due to N-S oriented tensional stresses, but controlled by a prevalently west-northwest and northeast oriented regional fracture pattern. The earliest known post-Grenvillian event indicating a N-S directed tensional stress regime in the area is the initiation of the Grenville dike swarm, which may have been mutually related in time and cause to the initiation of the graben.

2. Although there is no clear evidence for post-Ordovician graben subsidence in the area, regional considerations suggest that the post-Ordovician graben formation along the Ottawa Valley may have extended into the Nipissing Depression. Tension fracturing may also have taken place in the area, around the beginning of the Cambrian, synchronously with the Lake Nipissing alkaline magmatism.

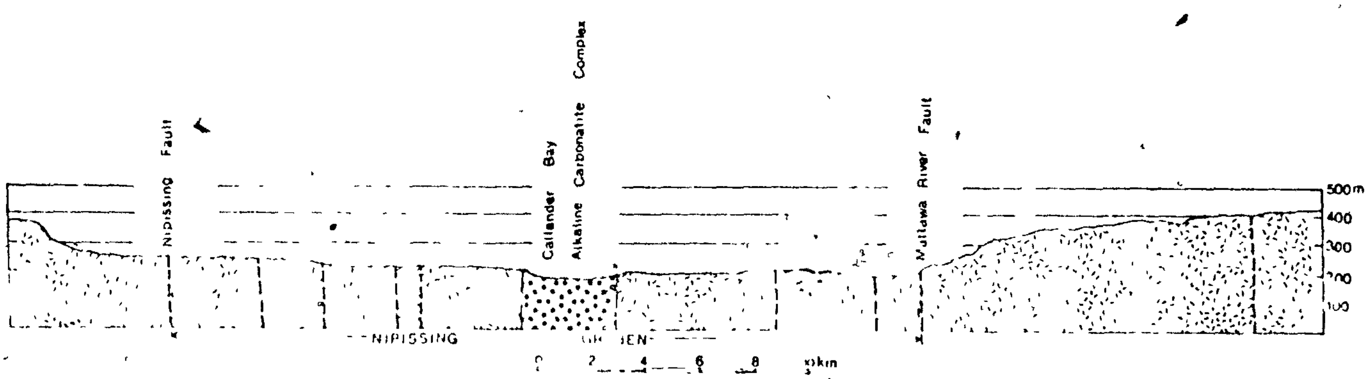

Fig.28. Section across the Nipissing graben. For location of the line of section see A-B in Fig.2. Faults assumed to be vertical.

SUMMARY AND CONCLUSIONS

The Nipissing Depression is underlain by a zone of intense block faulting which also appears to be a zone of downfaulting. Although this faulted zone and the Ottawa graben are described separately in this thesis, they should be regarded as parts of a single graben structure. For the part along the Nipissing Depression, the name Nipissing graben is proposed. The Nipissing graben faults appear to terminate somewhere in the area of the North Channel of Lake Huron.

The products of two post-Grenville igneous events: (i) the Grenville dike swarm of Hadrynian age (ii) alkaline-carbonatite intrusions of the Lake Nipissing alkaline province with K/Ar ages of about 565 m.y., are closely associated with the Nipissing graben. It is thought that the graben initiated synchronously with the emplacement of the Grenville dike swarm and that the graben faults went through a period of reactivation, synchronously with the Lake Nipissing alkaline magmatism. Renewed graben subsidence may have occurred along the structure in post-Ordovician times, synchronously with the post-Ordovician graben subsidence along the Ottawa Valley.

CHAPTER XI: SHIELD SEGMENT A:

THE TIMISKAMING DEPRESSION

INTRODUCTION

Upriver from the head of the Ottawa Valley, the Ottawa River lies in a rocky gorge extending approximately N30^{O}W, which further upriver expands to form a wide triangular depression containing Lake Timiskaming and traceable northwards to about the Hudson Bay watershed, a total distance of over 225 km (Fig. 26). To this topographic feature, the term Timiskaming Depression is applied in this work.

The rocky gorge in which the Ottawa River lies is about 1 1/2 km wide and has a slightly zig-zag course. Its steep to nearly vertical walls rise 125 to 175 m above the general level of the river. The Lake Timiskaming basin is about 20 km wide at its northern tip. The west side of the lake basin is bounded by an abrupt, linear fault-line scarp (Hume 1925, p. 47) which in some places rises about 100 m above the lake level. The east margin of the lake basin is quite irregular and along this side the slope towards the lake is gradual. The lake water is reported to be about 125 m deep (see Barlow 1897, p. 231 I), therefore, the maximum depth of the depression below the surrounding country is about 300 m. North of the lake, the depression continues to widen to form a rather ill-defined, topographic low of composite character consisting of several narrow, northeast and northwest trending linear depressions separated by patches and strips of relatively high ground (Fig. 26).

Parts of the depression, particularly the area northeast of Lake Timiskaming, are partly filled with stratified clay that was deposited in glacial lake Barlow-Ojibway (Grant and Hobson 1964; Lee 1965, pp. 18-19; Lovell and Caine 1970, p.6). The thickness of clay generally varies from 0 to about 100 m. In one locality, however, a thickness of as much as 200 m has been estimated (see Hobson and Lee 1967).

GEOLOGICAL CHARACTERISTICS

General

The southern part of the Timiskaming Depression is in the Grenville Province and its northern part is in the Superior Province. Throughout its length, the depression is superimposed discordantly on both Grenville and Superior trends; it extends across the Grenville Front almost at right angles (Fig. 29). A group of three closely spaced Paleozoic outliers composed mainly of carbonates and shales occur as downfaulted blocks in the depression (Fig. 29), indicating that the shelf environment of the St. Lawrence Platform extended into this area. Unlike the platformal sequences in Ottawa and St. Lawrence Valleys, strata as young as middle Silurian are preserved in one of the outliers (Hume 1925, Ollerenshaw and McQueen 1960). The oldest rocks present are Middle Ordovician. In the context of the present work, the most significant geological characteristic of the area is the existence of regional high-angle faults (Fig. 29). The larger faults are parallel to the general trend of the depression and there is sufficient space

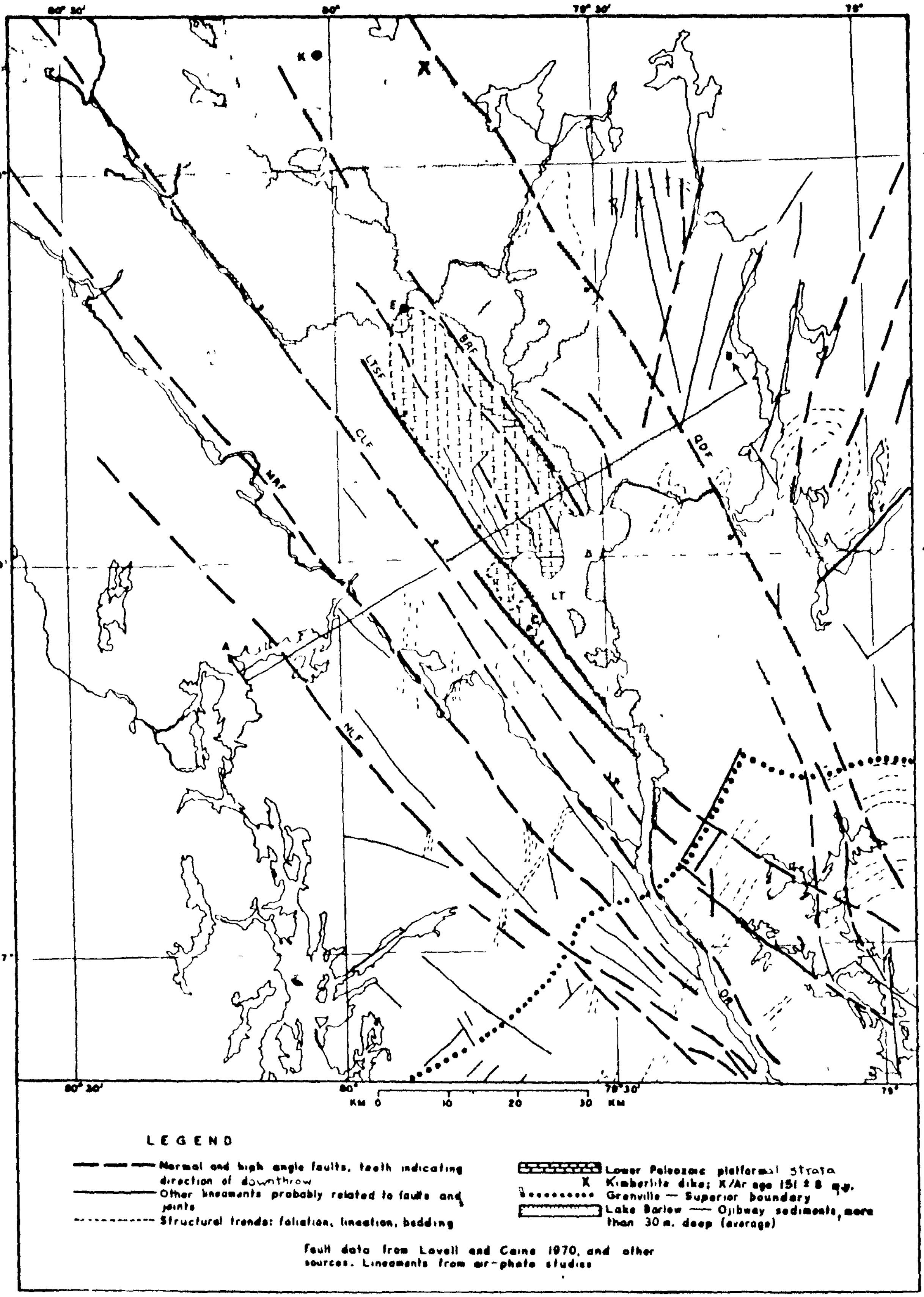

Fig. 29. Map of the Timiskaming Depression Area showing selected features and some locality names. Key to abbreviations: BRF-Blanche River fault; CLF-Cross Lake fault; C-Cobalt; K-Kirkland Lake; LTSF -Lake Timiskaming west shore fault; MRF-Montreal River fault; NLF-Net Lake fault; QDF-Quinze Dam fault.

correlation between the faulted belt and the depression to warrant the view that they are related features.

High-angle Faults

Faults in the area have strong physiographic expression, commonly in the form of narrow linear depressions, tens of kilometers long, occupied by water courses and filled in varying degrees by glacial debris. Such lineaments form a systematic pattern in the general area consisting of a strongly-developed northwest set and a less well-developed northeast set. They are such conspicuous features of the physiography of the area that they attracted the attention of early workers who debated their origin (Bell 1894, p. 364; Miller 1905, p. 36; Hobbs 1905, p. 19; Pirrson 1910, pp. 23-32; Wilson 1918, p. 39; also see Quirke 1936, pp. 267-288). Over the years, increased geological observations have led to the acceptance of faulting and/or jointing as the primary cause of their origin (Chagnon 1965, p. 229).

The nature of faults in the area is poorly known, except in a few instances where they cut Paleozoic rocks or have been intersected in underground mine workings (e.g. see Ninacs 1967, p. 153). One of the earliest known and better documented faults in the area is the one related to the scarp along the west margin of Lake Timiskaming (Hume 1925, p. 47). The Lake Timiskaming West Shore fault (fig. 29) as it is referred to, appears to be a normal fault. It strikes approximately N40°W, dips about 50°NE, and can be traced for a distance of about 50 km. Of the three Paleozoic outliers, the largest one is preserved on the downthrow side of this fault, and the vertical displacement of the downthrown block (of platformal rocks) is a little over 300 m (see Hume 1925, p. 48). The other two smaller outliers are preserved in a step-like downfaulted block between the Lake Timiskaming West Shore fault and the parallel McKenzie fault (Thomson 1964). Parallel to the above faults, and to the west of them, is the Cross Lake fault (Figs. 29 & 30), extending through Cross Lake and along the Englehart River valley. This appears to be one of the longest faults in the area; the lineament associated with it can be traced for a distance of over 250 km. In two localities the fault has been intersected by underground mine workings (Lovel and Cain 1970, p. 4; Ninacs 1967, p. 153). The observed dip is 65°NE, movement is dip slip with northeast side downthrown. This fault is presumed to cut a composite granite intrusion known as the Round Lake batholith. Gibb and van Boeckel (1970) found that the levels of the Bouguer gravity field over the supposed faulted halves of the batholith are different. They showed that the anomaly can be explained using two mutually exclusive models: one involving normal faulting of a granitic batholith of uniform density and the other based on density variations corresponding to an observed facies change within the pluton. Based solely on the first model, Gibb and van Boeckel (1970) estimated that the northeast half of the pluton has dropped about 3 km. Southwest of the Cross Lake fault, other parallel high-angle faults of regional extent have been inferred largely on physiographic evidence. Examples are the Montreal River fault (Ginn <u>et al.</u>

Fig.30. Vertical airphoto lineament of the Cross Lake Fault, Timiskaming Depression. National Airphoto Library, Ottawa, Photo No. A-13126-125. Scale-1:50,000 approx.

1964) and Net Lake fault (Thomson and Savage 1965). The nature of these faults and the movements involved are not known, but they, together with the Lake Timiskaming West Shore fault and the Cross Lake fault, appear to form a single fault system. On the east side of Lake Timiskaming there appears to be at least one fault of regional extent - the Quinze Dam fault (Fig. 29) which is nearly parallel (strike N30°W) to the faults on the west side of the lake. Little is known about the nature of this fault except that its physiographic expression is compatible with a downthrow to the southwest. Thus, apart from the Lake Timiskaming West Shore fault and the Cross Lake fault, the nature of other northwest trending faults is poorly known, but the close space association and parallelism of the faults indicate that they are parts of a single fault system. Based on characteristics of the better known faults the following generalizations can be made on the fault system.

1. The faults are probably of the normal type.

2. The faults on the west side of Lake Timiskaming dip NE.

The northwest trending faults described above fan out northwestwards. This tendency for the strain to be distributed over an increasingly wider area probably represents a splay out pattern suggesting that the faults die out northwestwards. The associated lineaments extend well beyond the northern limit of the depression, for another 100 km or so before they finally become lost. Southeastwards, some of the lineaments can be traced to the Upper Ottawa Valley. For example the lineaments of

Quinze Dam and Net Lake faults can be traced to join the Upper Ottawa Valley, suggesting that the faults in the two areas may also be connected. If this be the case, it follows that the Ottawa graben faults bifurcate and extend along the Nipissing and Timiskaming Depressions. Such a bifurcation is also consistent with the following considerations:

1. Because the Nipissing Depression faults and Timiskaming Depression faults appear to die out westwards and northwestwards respectively, a bifurcation of the Ottawa graben as postulated would fit in as part of a splay out pattern.
2. The Ottawa graben, where it appears to branch, bends 30° counterclockwise as though to accommodate a bifurcation and if not for this bend, the graben trend would bisect the 60-degree angle between the trends of the Nipissing and Timiskaming Depressions.

There is a much less prominently developed northeastern set of faults, best known from the Cobalt silver mining district (Jambor 1971, p. 23) located on the west side of Lake Timiskaming (Fig. 29). Some of the $N20^{\circ}E$ striking lineaments northeast of Lake Timiskaming (see Fig. 26) also may reflect faults of this set. The directions of the two sets of faults were probably controlled by a pre-existing regional fracture pattern (see p.42).

Ages of High-angle Faulting

The Lake Timiskaming West Shore fault cuts middle Silurian rocks and therefore, contains post-Middle Silurian movements. Close to the Quinze Dam Fault, kimberlite dikes occur in two localities (Lee 1968; Satterly 1948, p. 13) about 50 km apart. Material from one of these occurrences (for location see Fig. 29) has yielded a K/Ar age of 151 ± 8 m. y. (Lee and Lawrence 1968, p. 1). It is likely that the emplacement of these dikes and fault movements in the area occurred as mutually connected events. Therefore, at least some of the post-Silurian movements may have taken place in the mid-Mesozoic. Evidence of post-Pleistocene fault movements have been observed by Miller (1913). Minor stress release may be taking place on these faults at the present time, because a large earthquake (M - 6.2) that took place in the area in 1935, appears to be located on these faults (Smith 1966a).

There are indications that at least some of the faults in the area existed in the Precambrian. For instance, in the Cobalt District, the northeast trending faults appear to be pre-ore (Thomson 1967, p. 137) and the ore is believed to be genetically related to the Nipissing diabase whose K/Ar age is about 2095 ± 105 m. y. (Lowden et al. 1963). Furthermore, the unusual thickness of Proterozoic sediments in an area northwest of Lake Timiskaming may be an indication that subsidence occurred along and near the Cross Lake and Lake Timiskaming faults early in the Proterozoic (Jambor 1971, p. 23).

The chronology of fault movements in the area can also be considered from a regional point of view. If the Ottawa Valley faults and the faults along the Nipissing and Timiskaming Depressions are parts of a single bifurcating fault system, then like the faults of the two former areas, the Timiskaming Depression faults also may have originated in the Hadrynian, synchronously with the emplacement of the Grenville dike swarm (see p.111). Although this dike swarm is not known to extend along the Timiskaming depression some of its dikes in extending past the southern part of the depression do tend to orient themselves northwest (see Fig. 54). But as discussed above, some of the Timiskaming depression faults, such as the Cross Lake fault, may have existed long before the Hadrynian. This is not in conflict with the hypothesis that the faulting in the three areas is related in time and cause however, because the indications for older faulting are present only in the part of the fault system that extends into the Superior Province; it is conceivable that older faults of favourable orientation already existed in this area when the Hadrynian faulting became operative.

Structural Effects of High-angle Faulting

Since Wilson (1903) first speculated that the Lake Timiskaming is the physiographic expression of a graben, increased geological knowledge of the area has led to the general acceptance of this view (Wilson 1959, p. 316; Ollerenshaw and MacQueen 1960; Lovell and Caine 1970). The name Timiskaming graben is proposed for this structure.

The Timiskaming graben appears to be a highly asymmetrical structure bounded on the southwest side by a zone of step faults formed by the northeast dipping Cross Lake fault, the McKenzie fault (Fig. 31) The Montreal River and Net Lake faults and the Lake Timiskaming West Shore fault may also belong to this step fault system. The precise

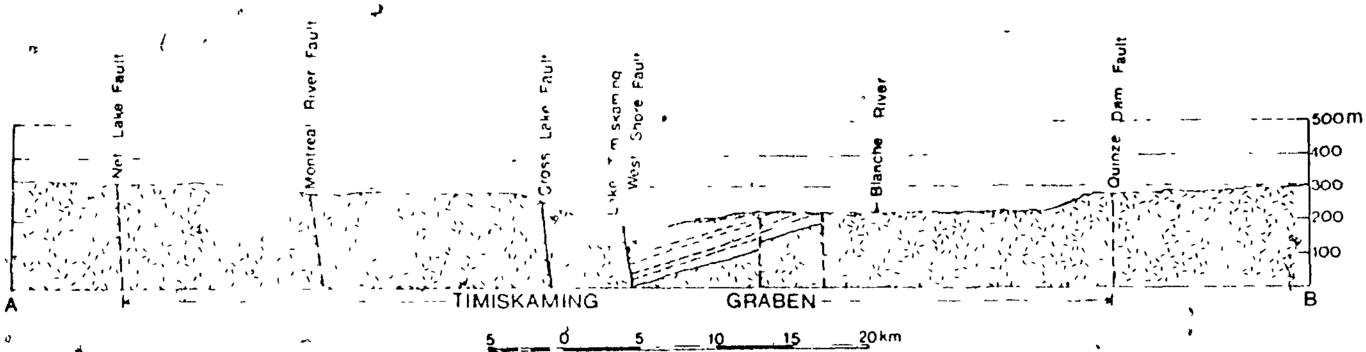

Fig.31. Section across the Timiskaming graben. For location of the line of section see A-B in Fig. 29. Ornamentation: unoriented-Precambrian rocks; oriented-Early Paleozoic(Middle Ordovician to Middle Silurian) platformal rocks; stippling-unconsolidated Quaternary sediments.

rôle of the Quinze Dam fault is not known. However, because this fault appears to be downthrown to the southwest, a dip in the same direction would be consistent for a normal fault (see Lovell and Caine 1970, Fig. 2).

As mentioned earlier, there are indications that the Ottawa graben faults and the Timiskaming graben faults are connected but the actual connections are poorly known. However, because the two grabens are structures of regional extent, their close spatial arrangement in a branching dispostion indicates that they are parts of a single structure.

SUMMARY AND CONCLUSIONS

The Timiskaming Depression overlies the axial zone of a complex graben, for which the name Timiskaming graben is proposed. The structure can be traced for a distance of about 200 km and possibly more. Northwestwards its faults fan out, suggesting splaying out and termination of the structure.

There can be little doubt that the Timiskaming graben is a branch of the Ottawa graben, the other branch being the Nipissing graben. Thus, it appears that the Ottawa graben, after extending for a distance of about 400 km along the Ottawa Valley bifurcates. Each of the two branches extends for over 200 km and appears to terminate by splaying out of their structures. This branching graben system which underlies the Shield Segment A, probably originated in the Hadrynian, synchronously with the emplacement of the Grenville dike swarm and underwest subsequent reactivations of its faults. Along a large part the Shield Segment A, there is conclusive evidence for post-Ordovician graben subsidence which may have taken place in the mid-Mesozoic synchronously with the emplacement of the Monteregian intrusives (at the east end of the graben system) and the late Jurassic-early Cretaceous kimberlite dikes along the Timiskaming graben.

CHAPTER XII: SHIELD SEGMENT B:

THE SAGUENAY-LAC ST. JEAN DEPRESSION

INTRODUCTION

From the lower end of the St. Lawrence Valley, two nearly a parallel canyons, spaced 6 to 10 km apart, extent N70°W across the Laurentian Highlands. One of these is occupied by the Saguenay River and the other by the Moulin-à-Baude and Ste. Marguerite Rivers. About 100 km from the St. Lawrence River, the canyons widen and eventually coalesce, to form a distinct topographic low, which continues west-northwestwards to include the Lac St. Jean basin. This topographic low together with the two canyons and the intervening strip of country is referred to in this thesis as the Saguenay-Lac St. Jean Depression

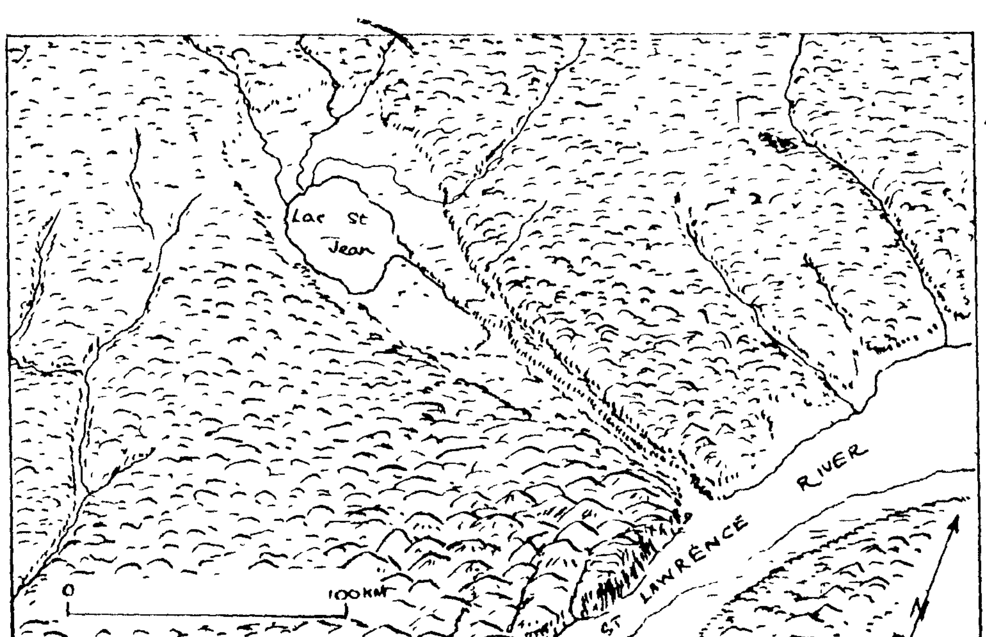

Fig. 32. Physiographic sketch of the Saguenay-Lac St. Jean Depression and vicinity.

(Fig. 32). The entire feature is horn-shaped in outline, the narrow end being at the St. Lawrence River. The Laurentian Highlands in the general area has elevations ranging from 300 to about 1000 m above sea level. If the 300 m topographic contour is taken as delimiting the depression, then its width is about 6 km at the St. Lawrence River, about 55 km just west of Lac St. Jean and about 100 km where it looses its identity in the Shield Region; its total length is about 300 km.

The Saguenay canyon is a spectacular topographic feature. Its steep to vertical walls (Fig. 33) rise as high as 450 m above the level of the river. In the river, depths of over 225 m have been recorded (see Dresser and Dennis 1944, p. 196) so that in places, the canyon walls are over 650 m high. The Ste. Marguerite-Moulin-à-Baude canyon (see Fig. 34) is partly filled with glacial debris; its actual depth (in bedrock) is not known. In many places, however, this canyon is as wide as the Saguenay canyon.

West of where the canyons coalesce, the depression lies between north and south facing scarps, about 100 m high. The north facing scarp is only slightly dissected whereas the south facing scarp is highly dissected. Westwards both scarps increase in amplitude, until in the area of Lac St. Jean the more prominent north facing scarp attains a height of as much as 200 m. Further westwards the scarps decrease in amplitude and finally become lost. The Lac St. Jean basin is a shallow feature; the lake level is about 100 m above sea level and water depth is less than 60 m.

Fig. 33. Oblique airphoto of the Saguenay canyon, south wall. National Airphoto Library, Ottawa, Photo No. A-1672-98.

Fig.34. Vertical airphoto lineament of Ste. Marguerite "trench", Saguenay- Lac St.Jean Depression. Scale 1:50,000 approx. The river in the picture is Ste. Marguerite. National Air Photo Library, Ottawa, Photo No.A-13941-23.

GEOLOGICAL CHARACTERISTICS

The Saguenay Lac St. Jean Depression lies entirely within the Grenville Province. Grenville crystalline rocks form the bedrock under much of the depression. They make up the walls of the canyons and the faces of the bounding scarps of the depression. In two separate areas within the depression, however, outliers of Paleozoic platformal strata overlie Grenville rocks (Fig. 35). The rocks are similar to parts of the platformal succession of the St. Lawrence Valley and consist of nearly flatlying Middle and Upper Ordovician limestones and shales (Dresser 1916), probably less than 100 m thick.

The depression extends with little regard to Grenville gneissic trends which in places make large angles with the depression trend. The linear crystalline scarps associated with the depression have traditionally been interpreted as fault-line scarps and the depression itself has been regarded as the physiographic expression of a graben - the Saguenay graben (e.g. see Kay 1942, p. 641). The recent discovery of a carbonatite complex - the St. Honoré complex (Vallée and Dubuc 1970, pp. 346-356) - within the depression, adds weight to the graben interpretation.

The St. Honoré Carbonatite Complex

The carbonatite complex was discovered in 1967 and is the only such complex known in the Saguenay-Lac St. Jean Depression. It is located in the area of the Paleozoic outlier north of the Saguenay

River (Fig. 35), and is overlain by Paleozoic (Middle Ordovician) rocks. Material from the intrusion has yielded a mean K/Ar age of 564 ± 4 m.y. (Doig and Barton Jr. 1968, p. 1403), an age almost identical to that of Lake Nipissing carbonatite complexes.

Vàllée and Dubuc (1970) state that the carbonatite body is roughly kidney-shaped with dimensions of 5 km by 3 km approximately. It is made up of dolomitic, sideritic and calcitic carbonatites (with minor urtite bands) containing niobium deposits of economic importance.

The St. Honoré complex is in contact with three other intrusive bodies, two of dioritic and one of syenitic composition. Also, about 8 km northwest of the complex is another circular complex - the Shipshaw complex - with a possible ring structure consisting of syenitic, granitic and gabbroic bands around a gabbroic core (Vallée and Dubuc 1970, p. 348). Apart from the close space association, it is not known whether the latter intrusions are related to the carbonatite.

In 1966, Kumarapeli and Saull emphasized (see Kumarapeli and Saull 1966a, p. 656) the possible use of the structures along St. Lawrence Valley system as a regional guide to prospect for niobium-bearing carbonatite bodies. The subsequent discovery of the St. Honoré Carbonatite complex adds more weight to this view. It is possible that further search may

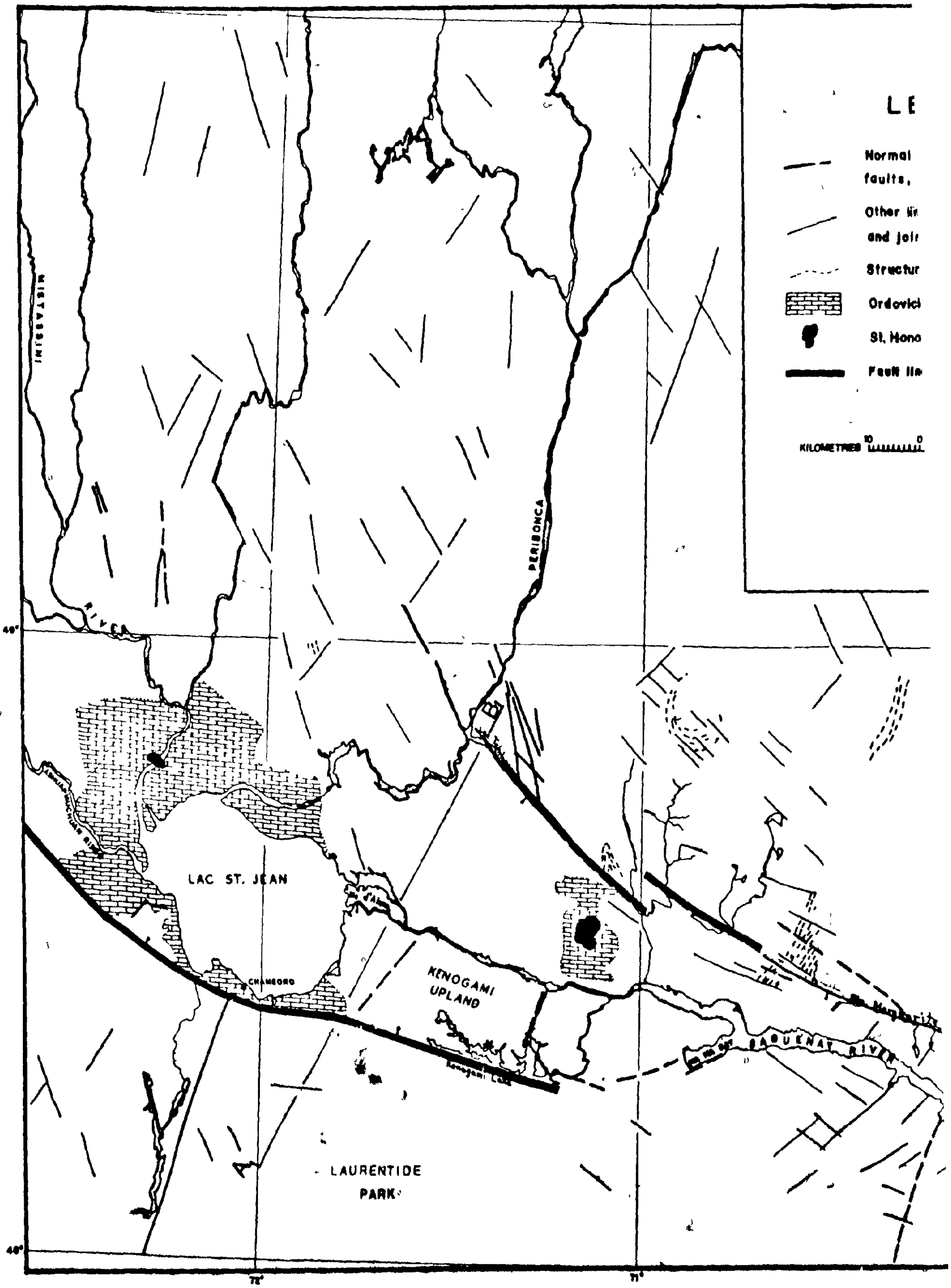

Fig.35. Map of the Saguenay-Lac St.Jean Depression area showing selected features. Fau
Clibbon and Bergeron(1963); Philpotts 1965. Lineaments from airphoto studies.

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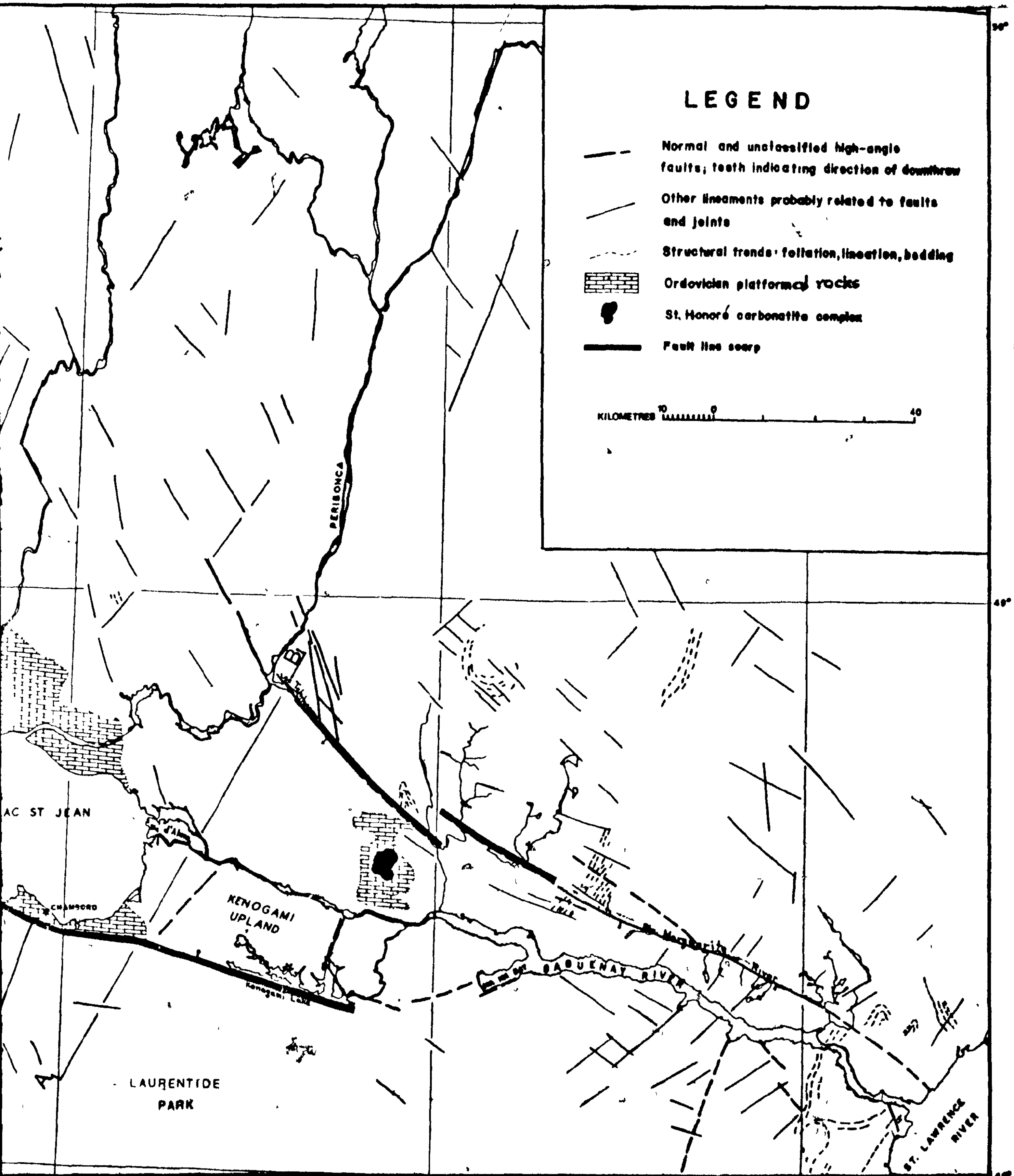

he Saguenay-Lac St.Jean Depression area showing selected features. Fault data from Dresser (1916); eron(1963); Philpotts 1965. Lineaments from airphoto studies.

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lead to other discoveries especially in the Upper Saguenay and Lac St. Jean areas and may eventually reveal in these areas an alkaline-carbonatite province similar to the Lake Nipissing alkaline province.

High-angle Faulting

Although high-angle faults of regional extent have been accepted as the primary spatial determinant of the Saguenay-Lac St. Jean depression, the presumed faults are nowhere known to be exposed. The faults are inferred from the following lines of evidence.

1. Abrupt sub-parallel linear scarps up to 200 m high or vertical-walled canyons up to 650 m deep occur along the depression margins. The trends of these lineaments transect Grenville trends, commonly at large angles. Although the canyons may have been initiated along faults or joints and subsequently overdeepened by glacial erosion, the systematically lower elevations of the depression floor, as compared with the elevations of the adjacent areas cannot be attributed to differences in erodibility of rocks (see Clibbon and Bergeron 1963, pp. 4-7) but is consistent with downfaulting along the depression. Therefore, faulting appears to explain adequately the linear scarps, the systematically lower elevation of the depression floor and possibly also the vertical-walled canyons.

2. Dresser (1916) states that on the south side of Lac St. Jean, near Chambord railway junction, Paleozoic rocks almost straddle the south boundary fault-line scarp and the Paleozoic formations there show drag (with dips ranging from 24 to 30^{o}N) typical for a normal fault with north side downthrown.

3. Numerous minor faults, presumably related to the larger faults reflected by topography, cut the Paleozoic formations. These minor faults, in all well-documented instances, are known to be of the normal type (Dresser 1916) indicating that the larger faults in the area are also normal faults.

4. The presence of at least the one carbonatite complex and the preservation of Paleozoic outliers are compatible with deep faulting and crustal subsidence along the depression.

The most persistent regional faults of the area appear to be those related to the boundary scarps of the depression. The faults are assumed to follow closely the bases of the scarps (Fig. 35). Eastward, therefore, these fault zones would extend into the canyons. They may continue along the canyons to join up with the Marginal Segment fault system. Such a connection is also indicated by the presence of the uplifted mass of the Laurentide Parc massif between the Saguenay-Lac St. Jean Depression and the St. Lawrence Valley. Westward the boundary faults zones of the Saguenay-Lac St. Jean Depression curve to assume a northwest orientation probably indicating a directional control from the northwest oriented component of the regional fracture pattern (see p.42) of the area. The lineaments (scarps, water courses) along the north boundary of the depression show a conspicuous en echelon arrangement indicating that the related fault zone may also have a similar arrangement.

Apart from the longitudinal faults, a less prominent set of transverse faults is also present in the depression area. These latter faults strike from north to northeast and are probably controlled by the northeast component of the regional fracture pattern (see p.42) of the area (Fig. 35). Where they cut Paleozoic rocks, the faults are well authenticated (Dresser 1916, p. 37). But other transverse faults are probably reflected by low, dissected scarps that extend across the depression. An example is the north-south trending scarp, 60 to 120 m high and located about 40 km west of Lac St. Jean (Clibbon and Bergeron 1962, p. 2). Compared with the longitudinal faults, the transverse faults appear to be shorter and to have smaller displacements.

The faults are believed to be mostly of the normal type because:

1. The observable minor faults in the area are normal faults. At Chambord, the drag associated with the south boundary fault indicates that in this locality at least, the fault is of the normal type.

2. The general elevation difference between the depression and adjacent highlands is consistent with downfaulting controlled by gravity faults.

3. Emplacement of carbonatite complexes such as the St. Honoré complex, commonly takes place in close space association with tension-fractured zones.

Throws in Grenville rocks are probably not determinable but throws in Paleozoic rocks have been estimated in some places. A throw (in Paleozoic rocks) of about 150 m, with northside down, has been estimated on the south boundary fault (Clibbon and Bergeron 1963, p. 12). Philpotts (1965, p. 7) has reported strike-slip movements of as much of 1800 m on what appears to be the easternmost part of the north boundary fault zone.

Ages of High-angle Faulting

The chronology of fault movements in the area is poorly known. The following are the pertinent lines of evidence.

1. Faults that cut the platformal rocks in the area have substantial post-Ordovician movements on them.

2. The intrusion of the St. Honoré carbonatite complex indicates that a tension-fractured zone existed in the area around the beginning of the Cambrian period.

Structure

That the physiographic and geological characteristics of the Saguenay-Lac St. Jean Depression are best explained as reflecting a graben-like zone of downfaulting was recognized by early workers (e.g. Wilson 1903; Dresser 1916, p. 46). Over the years, the structure underlying the depression has come to be accepted as a graben, which as mentioned earlier, is referred to as the Saguenay graben (see Kay 1942, p. 641). The boundary fault zones of the depression are obviously the key faults of this graben (Fig. 36); the transverse faults can be ascribed to block adjustments which must have accompanied structural subsidence of the main graben zone.

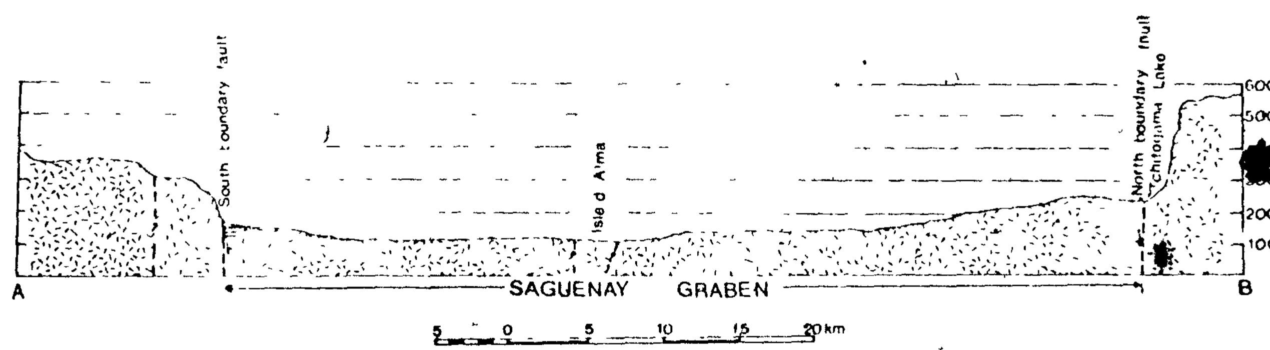

Fig.36. Section across the Saguenay graben. For location of the line of section see Fig.35. Ornamentation: unoriented-Grenville crystalline rocks; oriented-Early Paleozoic(Middle and Upper Ordovician) platformal rocks; stippling-unconsolidated Quaternary sediments.

Although the reality of the Saguenay graben can hardly be doubted, details of its structure are poorly known. There are indications that its structure may actually be a complex graben. For instance, a roughly rectangular area of the graben floor, lying between the Saguenay River and Kénogami Lake (See Kénogami Upland in Fig. 35) has elevations systematically higher (about 75 m) than those of the adjacent parts of the graben floor. The higher elevations cannot be explained in terms of differences in erodibility of bedrock in the area, but are adequately explainable in terms of a horst-block (Clibbon and Bergeron 1963, p. 13).

As mentioned earlier, the possible connections between the Saguenay graben faults and the Marginal Segment fault system are poorly known. However, because the Saguenay graben and the Marginal Segment faults are tensional structures of regional extent, their close spatial association in an intersection disposition suggests that they are probably connected.

Westwards from Lac St. Jean, the boundary faults of the Saguenay graben fan out. At least, the north boundary fault, appears to split into several faults. These features are thought to be parts of a splay-out pattern related to termination of the graben structure.

SUMMARY AND CONCLUSIONS

Since Wilson (1903) speculated that the Saguenay-Lac St. Jean Depression is a rift valley, increased geological knowledge of the area has led to general acceptance of the rift valley concept and the related graben has been referred to as the Saguenay graben. Although the possible connections of the Saguenay graben faults and the Marginal Segment faults are poorly known, there can be little doubt that the two fault system are connected, the acute angle between them being occupied by the Parc des Laurentides massif which appears to be a block mountain.

Downfaulting, along the graben occurred in post-Ordovician times, but the existence of an earlier tensional structure is also indicated by alkaline-carbonatite activity around the beginning of the Cambrian period.

CHAPTER XIII

THE OUTER PART OF THE LAURENTIAN CHANNEL

INTRODUCTION

As mentioned in Chapter VII, the Laurentian Channel branches on the downstream side of Anticosti Island and at this bifurcation the main channel bends clockwise through an angle of about 45^{o}; its continuation thence, is referred to in this thesis as the outer part of the Laurentian Channel (Fig. 15). This part of the channel is the most impressive part of the St. Lawrence Valley system. It is nearly straight submarine trough, about 500 km long and about 100 km wide and for the most part has a U-shaped cross-profile. Depths of over 400 m are common, with a few elongate depressions descending to depths of over 500 m. The deepest depression is in the western part of Cabot Strait and has a maximum depth of 535 m. On each side of the Strait, land rises to an elevation of about 500 m within a short distance from the shorelines. Therefore, the total amplitude of the topographic low in this area is about 1 km.

As a segment of the St. Lawrence Valley system, the outer part of the Laurentian Channel is unique. It is the only part that extends across the Appalachian foldbelt. For the most part, the channel appears to be in Permo-Carboniferous successor basin deposits and Mesozoic and Tertiary platformal strata (Fig. 37). Acadian and older Appalachian structures seem to emerge through cover rocks only

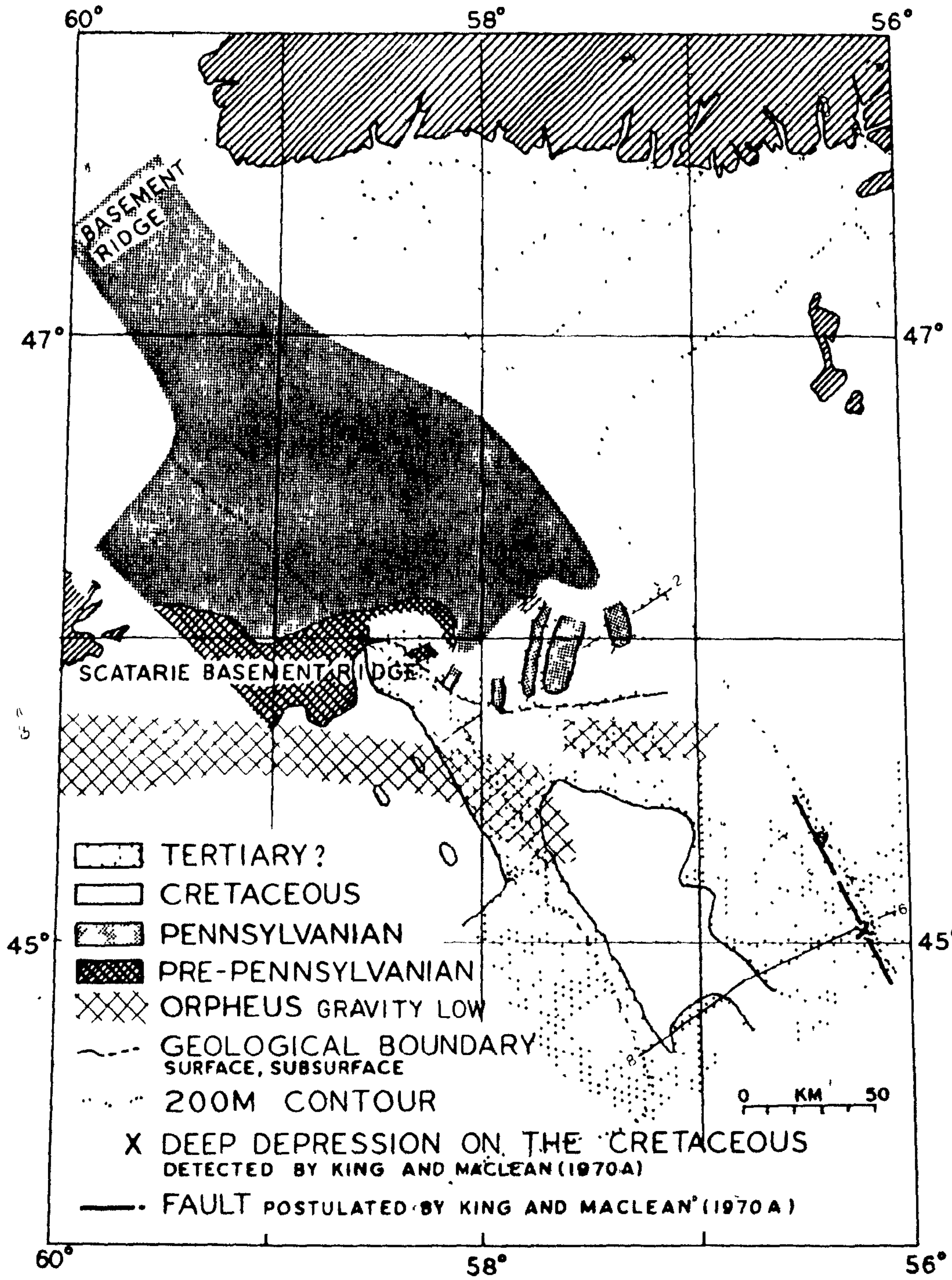

Fig. 37. Geology of the outer part of the Laurentian Channel, after King and MacLean (1970a).

where two fault-controlled structural highs cross the channel area. (Fig. 37). One of them is near the entrance to Cabot Strait (Sheridan and Drake, 1968). The other, known as the Scatarie ridge (Fig. 37) is directly east of Cape Breton Island.

The outer part of the Laurentian Channel attracted the attention of early workers (e.g. Spencer 1890). Over the past 80 years or so four different hypotheses have been advanced to explain its origin. These hypotheses are briefly outlined and commented on in Appendix V. The proponents of these hypothesis advocate two basically different views of its origin.

1. The channel overlies a structure of regional extent (e.g. Gregory 1929).
2. It is an erosional feature with little or no structural control (e.g. Shepard 1931).

For simplicity these two views will be referred to hereafter as "structural" and "erosional" hypotheses respectively.

In 1966, Kumarapeli and Saull (1966a) advocated a structural hypothesis for the origin of the channel. They favoured the view that the underlying structure is a rift zone and that this rift zone is a branch of a large rift system, coextensive with the St. Lawrence Rift system. This hypothesis has been criticized by several workers (Sheridan and Drake 1968; King and MacLean 1970a; Keen 1972) who support an "erosional" origin for the channel. The criticisms are based largely on interpretations of geophysical results. In the discussions to follow, the writer will attempt to show that the geophysical data, as known at the present time, are not fatal to the idea of a structural origin, and that a structural origin has certain attractions.

GEOPHYSICAL SURVEYS

Gravity Surveys

Gravity surveys have been carried out by Loncarevic (1965), Goodacre and Nyland (1966), Loncarevic and Ewing (1967), Goodacre et al. (1969), Stephens et al. (1971) and Watts (1972). A Bouguer gravity map of the area is shown in Figure 38. Some of the gravity trends show no deviation of their trends as they cross the channel margin. However, several of the larger gravity anomalies deviate from their strong linear trends as they cross the margins of the channel. The Orpheus gravity anomaly ("A" in Fig. 38; Loncarevic and Ewing 1967), for example, is disrupted at the southeast margin of the channel. Also, the trend of anomaly marked "B" in Figure 38 changes at the southwest margin and terminates at the northeast margin (Stephens et al. 1971). These disturbances of the gravity trends indicate that Appalachian structures, in crossing the channel, undergo some change which might be brought about by a dislocation.

Magnetic Surveys

Magnetic surveys over the outer part of the channel have been carried out by Bower (1962), Loncarevic and Ewing (1967), Hood (1967) and Watts (1972). A magnetic map of the area is shown in Figure 39

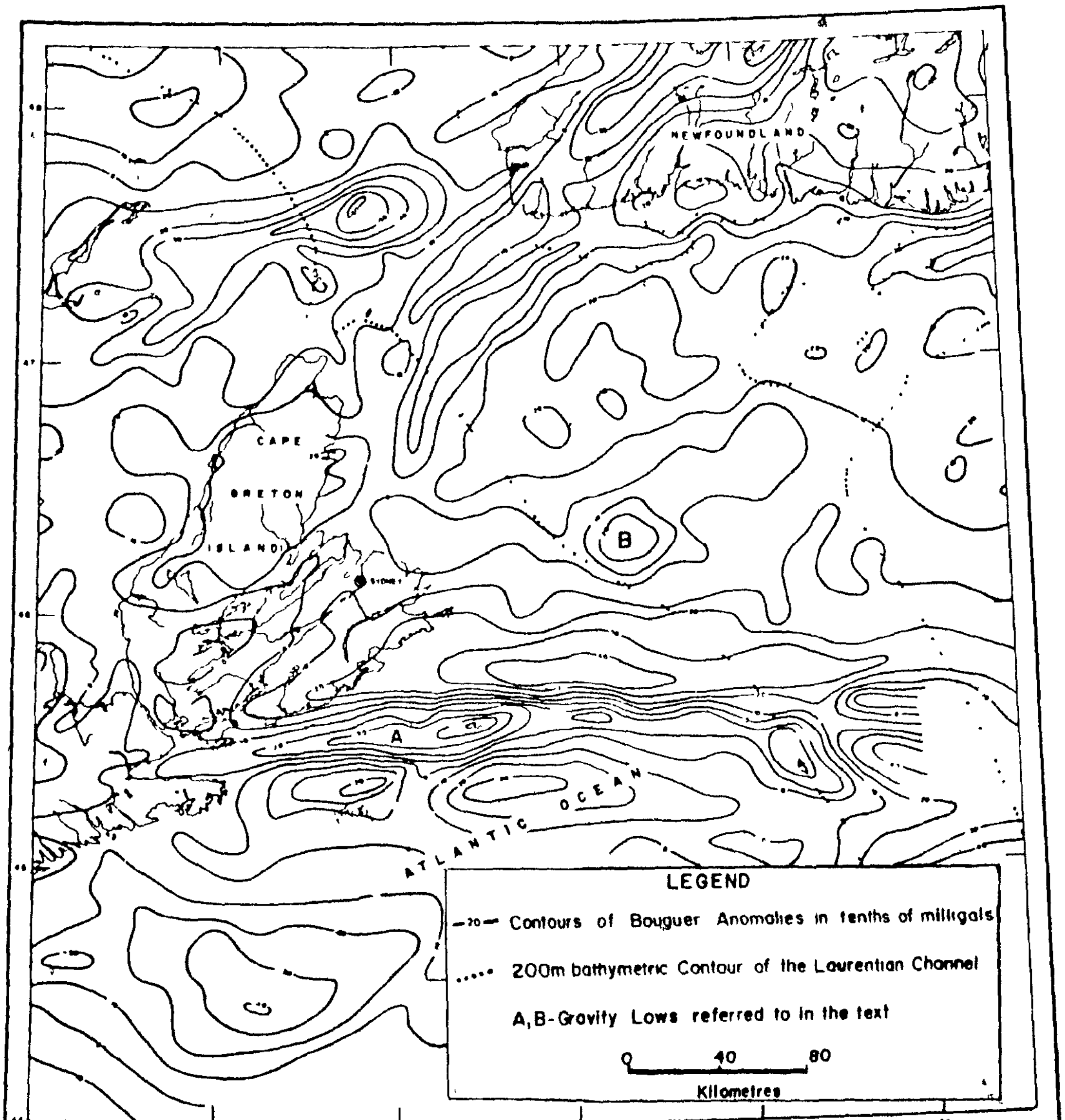

Fig.38. Bouguer gravity anomaly map of the Laurentian Channel (outer part) area. Gravity data from Bouguer anomaly map, Halifax-Burgeo, Nova Scotia-Newfoundland, Earth Physics Branch, Can. Dept of Energy Mines and Resources.

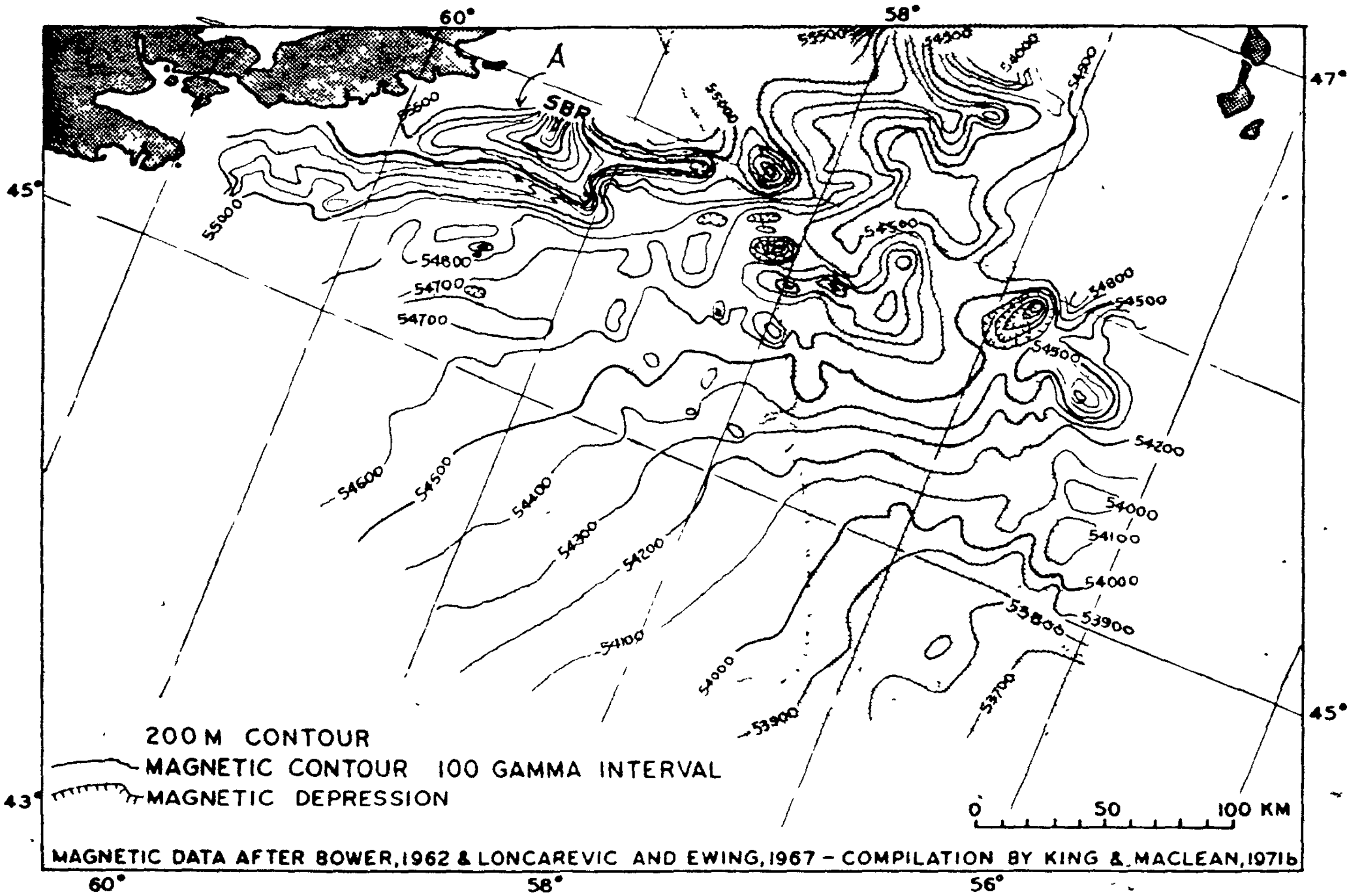

Fig. [illegible]. Total intensity magnetic anomaly map of the outer part of the Laurentian Channel and vicinity.

It is seen that the magnetic trends (total field) related to the Appalachian structures are distorted as they cross the channel margin. For example, the intense magnetic high ("A" in Fig. 39), presumably related to the pre-Carboniferous meta-volcanics and meta-sediments of the Scatarie basement ridge, becomes much less pronounced in crossing the southwest wall of the channel. These distortions of the magnetic field cannot be attributed to channel topography because the channel walls are made up of relatively non-magnetic post-Carboniferous rocks (King and MacLean 1970a, Fig. 6). Therefore, the most viable interpretation of the distortions of the magnetic field is that the normal Appalachian trends are in some way disturbed by a transverse trend along the channel.

Seismic Refraction Surveys

Seismic refraction surveys across the outer part of the channel have been carried out by Press and Beckmann (1954), and by Sheridan and Drake (1968). The former workers concluded that their results support the presence of a graben-like structure along this part of the channel, thus supporting Gregory's (1929) hypothesis. The key fault of this structure was interpreted to be along the northeast margin of the channel. Sheridan and Drake (1968) on the other hand, found no evidence in their results to support the existence of a large fault as suggested by Press and Beckmann(1954). Furthermore, they argued that Press and Beckmann's (1954) results could be revised and reinterpreted to show no large downfaulting along the northeast margin of the channel. Sheridan and Drake (1968, p. 357) mention, however, that the basement configuration indicated by seismic evidence along two refraction profiles (their profile 141 and the reinterpretation of profile 18 of Press and Beckmann be related to a high-angle fault (Carboniferous or younger) along the southeast side of the channel.

Sheridan and Drake's (1968) work casts doubt on the reality of the structural section, across the channel as interpreted by Press and Beckmann (1954) but it is not clear whether Sheridan and Drake's seismic refraction results are opposed to the hypothesis of a regional structure along the channel. On the contrary, the fault suspected by them (along the southeast side of the channel) may be a part of such a structure. The lack of clear cut evidence in the refraction results, together with the gravity-magnetic picture as known in 1968, led Sheridan and Drake (1968, p. 357) to conclude that:

> "there is little existing evidence to support the presence of a significant southeasterly trending structure through Cabot Strait and the Laurentian Channel".

The present gravity-magnetic picture however, renders the above conlusion much less viable.

Seismic Reflection surveys

Shallow (depth penetration ~ 200 m) seismic reflection surveys of the area have been carried out by King and MacLean (1970b) These include two continuous cross-profiles (for locations see 1-2 and 6-8, Fig. 37) the results of which have been the basis for two geological sections across the channel (King and MacLean 1970a, Fig. 6). Along both sections, the pre-Tertiary erosion surfaces on Cretaceous and Pennsylvanian strata are characterised by deep valleys which are filled with Tertiary deposits. In section 1-2 (for location see Fig. 37), valleys are cut into Pennsylvanian rocks to depths varying up to 125 m. In section 6-8 (for location see Fig. 37) one of the valleys is at least 800 m deep (actual depth not known) and it is apparently cut into Cretaceous strata and filled with Tertiary sediments. Some of these valleys might be carved along fault lines but the seismic results do not provide an unequivocal answer, mainly because of the shallow depth penetration of the survey method. King and MacLean (1970a) interpreted the paucity of conclusive evidence for faulting in their seismic records, in favour of an "erosional" origin (similar to the one advocated by Shepard in 1931) for the outer part of the channel. They supported their interpretation with the same geophysical (gravity-magnetic) arguments as used earlier by Sheridan and Drake (1968). Because the present gravity-magnetic data can be interpreted to support a "structural" origin for the channel (Stephen et al. 1971) it is not clear whether King and MacLean's (1970a) seismic reflection results are a challenge to a "structural" hypothesis.

King and MacLean (1970b) however, did postulate a fault along the outermost part of the channel (see Fig 37), mainly to account for a zone of mild deformation that they detected in the Cretaceous strata of the area. They placed this fault along the northeast margin of the channel (see Fig. 37). Thus placed, it coincides with a part of the major fault suggested by Press and Beckmann (1954). Although the fault is assumed to be parallel to the trend of the channel, King and MacLean (1970b) grouped it with the eastward extension of the Fundy graben structure through Chedabucto Bay. They identified the continuation of this fault with what the writer believes (see Kumarapeli and Saull 1966a, p. 653) to be the oceanic extension of the Laurentian Channel structure. This suggested oceanic continuation extends along the south margin of the Grand Banks and along the Southeast Newfoundland Ridge to meet the mid-Atlantic Ridge in the area of the Azores (Fig. 40). Along this suggested continuation, a fracture zone has been postulated by several workers (e.g. Le Pichon 1968; Watson and Johnson 1970) and has been called the Newfoundland Fracture Zone by Auzende et al. (1970). It appears reasonable to assume that the eastward extension of the Fundy graben structure from Chedabucto Bay across the shelf is along the Orpheus gravity low "A" in Figure 38. The anomaly probably reflects the low density sedimentary fill in the fault trough (Lancarevic and Ewing 1967). However, as mentioned earlier, the Orpheus anomaly deviates from its strong linear trend at the southeast margin of the Laurentian Channel. Within the channel the gravity anomaly assumes a pincer-like form*, perhaps suggesting that the structure goes into a splay before dying out. The same conclusion can also be reached when the Fundy graben is looked at from a regional

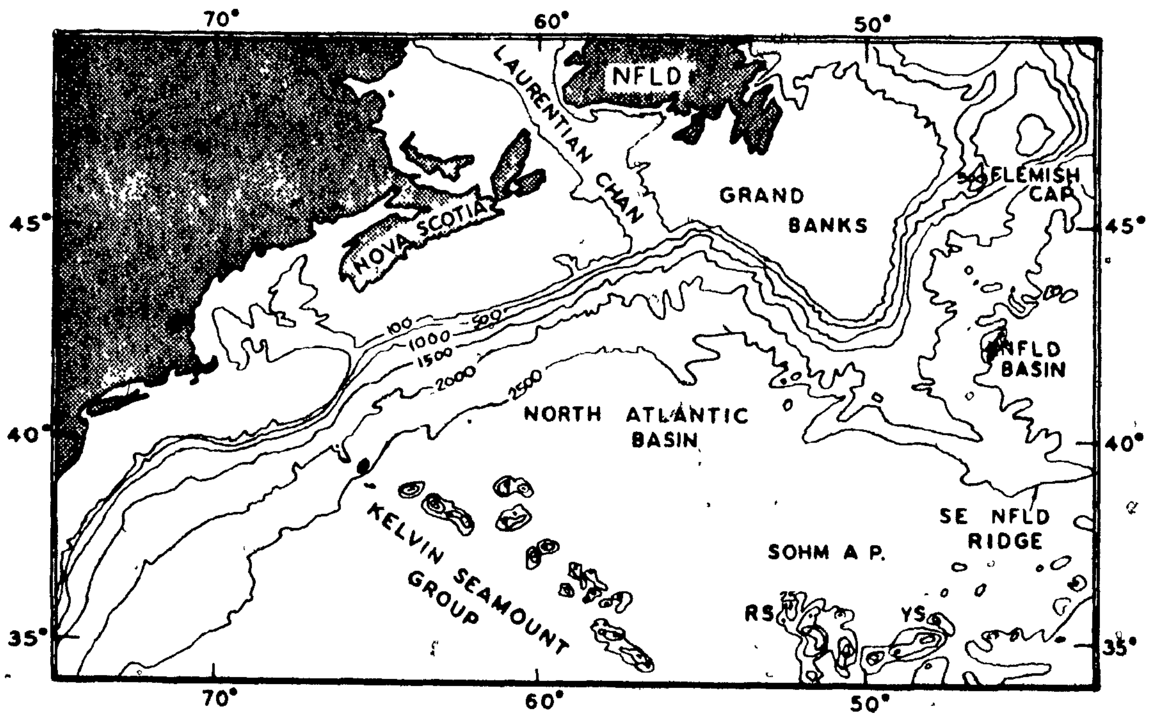

Fig.40. Map showing the near collinearity of the Laurentian Channel, the continental margin off-set in the Grand Banks area, and the Southeast Newfoundland Ridge.

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*The pincer-like form is clearly seen in the gravity map of Lancarevic and Ewing (1967). In the map of Stephens et al. (1971) reproduced in Figure 38 the cut off of gravity contours (in the westernmost part of the Orpheus anomaly) is due to lack of field data.

point of view. The graben structure divides into two branches at the northeast end of the Bay of Fundy (in a manner similar to the branching of the Ottawa graben into the Timiskaming and Nipissing grabens). This branching may be the actual beginning of a splay out pattern related to an impending termination of the structure.

It is however possible and quite likely that the Fundy graben faults are refracted on meeting a structure along the Laurentian Channel, but it is unlikely that they, after bending through 70° (to conform with the channel trend), would continue for any great distance let alone as far as the mid-Atlantic Ridge. Thus, to the writer, the assumption of King and MacLean (1970b) that the Newfoundland Fracture Zone represents the continuation of Fundy graben faults through Chedabucto Bay, appears ad hoc and artificial, whereas to identify the former structure with the extension of the Laurentian Channel trends seem more natural (see Fig. 40).

In summary, the existing geophysical evidence does not appear to be incompatible with the hypothesis of a structure, transverse to the Appalachians, along the present site of the channel. Such a hypothesis is attractive in the context of the following tectonic schemes.

1. The hypothesis of an anticlockwise rotation of the Newfoundland arc incorporates the concept that transcurrent faulting has taken place probably along the present site of the channel.

2. The branching pattern of the Marginal Segment rift zone at the south end of the St. Lawrence Valley is very similar to the branching pattern at the ends of the Rhine graben or the Eritrean graben (Fig. 42). In each of these latter grabens two similar bifurcations occur at either end. If the structures of the Shield and Marginal Segments do combine to form a large graben system, then symmetry considerations necessitate, in addition to the Esquiman Channel structure, another branch structure somewhere along the site of the outer part of the Laurentian Channel. It was partly on this basis that a branch structure of the St. Lawrence Rift system was postulated along the outer part of the Laurentian Channel (Kumarapeli and Saull 1966a).

Those who favour the hypothesis of a rotation of Newfoundland, suggest that as much as 300 km of pre-Carboniferous transcurrent movements may have taken place along the site of the channel (King 1951, p. 91). Although some of the geophysical anomalies (gravity, magnetic) are distorted as they cross the channel margins (see Figs. 38, 39) the overall geophysical picture does not support the hypothesis of substantial lateral offsets of the pre-Carboniferous structures. For example, the gravity and magnetic anomalies, presumably related to the two basement ridges (Fig. 37) show little or no lateral offsets as they cross the margins of the channel.

The evolution of a structure along the channel will be discussed later (see Chapter XV) within the framework of the evolution of the St. Lawrence Rift system. According to this scheme, the structure is envisaged as a relict fracture zone, inherited from a rift (graben?) which cut across the site of the Appalachian geosyncline during its early development. Such a structure could conceivably produce the distortions of some of the gravity and magnetic anomalies as they cross the channel. Post-Triassic movements on it may have disrupted the Fundy graben structure and its sedimentary fill, causing the distortions of the Orpheus gravity low at the southwest margin of the channel. The postulated fracture zone is compatible with the following features of the region.

1. The fracture zone is parallel to one of the dominant directions of faulting in Nova Scotia and in the adjacent shelf (Cameron 1956).
2. It is parallel to a magnetic trend of regional extent (reflecting a deep fault?) through Prince Edward Island (Battacharyya and Raychaudhuri 1967).
3. The postulated fracture zone and the Newfoundland fracture zone are nearly collinear (Fig. 40) suggesting that they may be parts of a continuous structure.

SUMMARY AND CONCLUSIONS

The outer part of the Laurentian Channel is topographically the most impressive part of the St. Lawrence Valley System. The question whether it is underlain by a regional structure (transverse to the Appalachian trend) or not, has been disputed for over 40 years and is still an issue of dispute. Much of the available evidence is equivocal on the question, although the latest gravity-magnetic data seem to indicate the presence of a structure. The hypothesis of a structure is geologically attractive in the contexts regional geology and symmetry considerations of the high-angle fault pattern of the St. Lawrence Valley system. The postulated structure is assumed to be a relict fracture zone inherited from a rift (graben?) which cut across the site of the Appalachian geosyncline during its early development. This fracture zone may be in an unhealed state, to a greater or lesser degree, at the present time.

CHAPTER XIV: THE CONCEPT OF A ST. LAWRENCE RIFT SYSTEM

GENERAL STATEMENT

The geological characteristics of the various topographic depressions that make up the St. Lawrence Valley system are described in Chapter IV through to XIII. In these descriptions the main emphases were on (a) the structures that underlie or are presumed to underlie the depressions (b) magmatic products that appear to be related to the structures. The main characteristics of these structures including their ages are summarized in Table I. Also included in Table I are the tectonic interpretations given to the structures by various workers. The following generalizations can be made regarding these structures.

1. The Shield Segments are underlain by grabens. In Segment A there is a main graben (the Ottawa graben) and two minor branch grabens (the Nipissing and Timiskaming grabens). Together they extend for a total length of over 600 km. The Segment B graben (the Saguenay graben) is about 300 km long.
2. A considerable part of the Marginal Segment is water covered and the geology of these areas is poorly known. Generalizing from areas of better known geology, it appears that two structural belts can be recognized. One is largely restricted to the narrow strip of platform

TABLE 1 ST. LAWRENCE VALLEY SYSTEM: STRUCTURE AND TECTONIC INTERPRETATION

MAJOR SUB-DIVISION	SECTION	REFERENCE PAGES OF THIS PAPER	APPROXIMATE LENGTH IN KM	RANGE IN WIDTH IN KM	STRUCTURE
MARGINAL SEGMENT	ST. LAWRENCE VALLEY	44-67	450	25-100	Block faulted and downbowed by play of movements on high-angle faults consisting mainly of a major set of longitudinal faults and two minor sets of oblique faults. The faults appear to be mostly of the normal type, except for one known reverse fault and possibly more on the east side of the valley. On this side the high-angle faults closely associated with east-dipping thrust faults.
	CHAMPLAIN VALLEY	68-80	175	10-70	Block faulted and structurally depressed by play of movements on high-angle faults which can be grouped into a major set of longitudinal faults and a minor set of transverse faults. Faults on the western part of the valley probably are normal faults. In the eastern part, nearing the Appalachians, normal, high-angle reverse and east dipping thrust faults occur in close association.
	ESQUIMAN CHANNEL AREA	81-97	500	40-250	Evidence from coastal belts indicate that the area is structurally depressed, mainly by movements on a set of longitudinal high-angle faults. On the Newfoundland side, the coastal belt cut by steeply northwest dipping to vertical normal faults and steeply southeast dipping to vertical high-angle reverse faults. The mainland coastal belt contains normal faults (downdropped to the southeast) and other longitudinal high-angle faults of uncertain type. Some of the latter faults appear to have transcurrent movements of less than 1 km.
	INNER PART OF THE LAURENTIAN CHANNEL	98-104	800	25-250	Structure uncertain. The area for the most part water covered. Large high-angle faults with south side downthrown known from Anticosti Island. Evidence of normal faulting parallel to the channel in the Sept Isles area. These pieces of evidence, considered together with the physiographic and tectonic setting of the Channel, lend some justification to think that the St Lawrence Valley and Esquiman Channel area faults may be continuous through the channel.
SHIELD SEGMENT A	LOWER OTTAWA VALLEY	105-117	125	75	Block-faulted and structurally downdropped by play of movements on two sets of high-angle faults: a major set of longitudinal faults and a set of oblique faults. Faults appear to be of the normal type.
	UPPER OTTAWA VALLEY	118-134	275	30-60	Block-faulted and structurally downdropped by play of movements on two sets of high-angle faults: a major set of longitudinal faults and a minor set of transverse faults. Faults appear to be of the normal type.
	NIPISSING DEPRESSION	135-149	>200	25->100	Severely block-faulted and possibly structurally downdropped. Appears to be the continuation of the Upper Ottawa Valley structure. Faults can be grouped into a longitudinal set and a transverse set. Some of the faults appear to extend as far as the North Channel of Lake Huron but they possibly terminate in this area.
	TIMISKAMING DEPRESSION	150-161	> 225	2->100	Structurally downdropped by play of movements on high-angle faults consisting of a major set of longitudinal faults which fan out northwestwards and a minor set of transverse faults. Faults probably of the normal type. The longitudinal faults may be connected with the Ottawa Valley faults. They appear to extend for a distance of over 100 km beyond the north tip of Lake Timiskaming.
SHIELD SEGMENT B	SAGUENAY-LAC ST. JEAN DEPRESSION	162-175	> 300	10 >100	Structurally downdropped by play of movements on high-angle faults consisting of a major set of longitudinal faults which curve and fan out northwestwards and a minor set of transverse faults. Faults probably of the normal type. The longitudinal faults are possibly connected with the St. Lawrence Valley faults. They appear to terminate about 75 km northwest of Lac St. Jean.
C	OUTER PART OF THE LAURENTIAN CHANNEL	176-189	500	95	Evidence equivocal as to whether the Channel is underlain by a structure or not. Hypothesis of a structure geologically attractive.

STRUCTURE	AGES OF HIGH ANGLE FAULTING	TECTONIC INTERPRETATIONS
. .-bowed by play of movements on .-sisting mainly of a major set of - and two minor sets of oblique faults. .o be mostly of the normal type, ex- reverse fault and possibly more on .e valley. On this side the high- .y associated with east-dipping thrust	Faults in the two areas form a continuous fault system. Evidence for post-Ordovician movements conclusive. What appears to be ancestral faults may have existed in the Cambrian or earlier. The post-Ordovician movements may have taken place in the mid-Mesozoic, synchronously with the Monteregian igneous activity.	A rift zone - Marginal Segment Rift Zone - which in some areas has a graben-like cross section. Source: This paper Also Kumarapeli and Saull 1966a.
.tructurally depressed by play of .gle faults which can be grouped into . tudinal faults and a minor set of Faults on the western part of the e normal faults. In the eastern .palachians, normal, high-angle .ping thrust faults occur in close		
.al belts indicate that the area is .ed, mainly by movements on a set .-angle faults. On the Newfound- .tal belt cut by steeply northwest .ormal faults and steeply south- .tical high-angle reverse faults. .belt contains normal faults - southeast) and other longitudinal .f uncertain type. Some of the .ar to have transcurrent movements	Evidence for post-Ordovician movements conclusive. Some faults cut rocks as young as Carboniferous. Possible an cestral faults appear to have originated about 800 m.y. ago synchronously with the emplacement of diabase dike swarms, and eruption of flood basalts in the Strait of Belle Isle area.	
n The area for the most part .ge high-angle faults with south .an from Anticosti Island. Evi- .iting parallel to the channel in . These pieces of evidence, con- .h the physiographic and tectonic .el, lend some justification to Lawrence Valley and Esquiman . may be continuous through the	Faults on Anticosti Island cut middle Silurian rocks and those in Sept Isles area cut middle Ordovician - rocks.	
structurally downdropped by play of .ts of high-angle faults: a major . faults and a set of oblique faults. . of the normal type.	Faults in the three areas form a continuous fault system. May have originated in the Hadrynian, synchronously with the emplacement of the Grenville dike swarm. Conclusive evidence for post-Ordovician (mid-Mesozoic?) movements in Lower and Upper Ottawa Valleys. Several Paleozoic episodes of alkaline and alkaline-carbonatite magmatism, including one major episode in Lake Nipissing area, indicate reactivations of faults, possibly of local extent.	Grabens, commonly asymmetrical and branching in the case of Segment A grabens. Sources: Wilson 1903 Kindle and Burling 1915 Kay 1942 Wilson 1959 Lovell and Caine 1972 Also this paper and Kumarapeli and Saull 1966a.
structurally downdropped by play of .ts of high-angle faults: a major . faults and a minor set of trans- .ts appear to be of the normal.		
.ted and possibly structurally down- .s to be the continuation of the . structure Faults can be grouped . set and a transverse set. Some .r to extend as far as the North .on but they possibly terminate in		
.ed by play of movements on high- .sting of a major set of longitudinal .t northwestwards and a minor set .ts. Faults probably of the normal .nal faults may be connected with faults. They appear to extend for - 100 km beyond the north tip of	Conclusive evidence for post-middle Silurian movements on one major fault of the longitudinal fault system, but these post-middle Silurian faults may have taken place on the other faults as well. The age of one of two kimberlite dike occurrences associated with the fault system suggest that some at least of the post-Silurian movements may have taken place in mid-Mesozoic. Ancestral faults appear to have existed in the Precambrian.	
.ed by play of movements on high- .sting of a major set of longitudinal . and fan out northwestwards and a .verse faults. Faults probably of .e longitudinal faults are possibly . St. Lawrence Valley faults. They .e about 75 km northwest of Lac St.	Evidence for post-Ordovician movements conclusive. Earlier movements indicated by the close space association of faults with one known alkaline-carbonatite complex dated at about 565 m.y.	
. as to whether the Channel is under- .e or not. Hypothesis of a structure .ctive.		Might consist of a fracture zone connecting the Marginal Segment rift zone with the Newfoundland Fracture Zone. Source: This paper and Kumarapeli and Saull 1966a.

and consists mainly of a belt downfaulted on longitudinal normal faults. Cross sections of this belt vary. In the Champlain Valley area the cross section appears to be that of a complex graben; in the St. Lawrence valley and on Anticosti Island a series of longitudinal crustal blocks downdropped away from the shield margin; and in the southwest part of the Strait of Belle Isle, a graben. This downfaulted belt, containing mainly dislocations of tensional origin, will be hereafter referred as the Marginal Segment Rift Zone. The feature is about ~2000 km long and has a sigmoid shape. The other structural belt of the Marginal Segment lies immediately to the south and east of the Marginal Segment Rift Zone and in some areas straddles the north and west boundaries of Logan's Zone. In this belt normal faults, high-angle reverse faults and low-angle thrust faults occur in close association. The age relations of the different types of faults are poorly known. The high-angle reverse faults and the thrust faults dip towards the Appalachian foldbelt and appear to be structural elements created as part of orogeny in the Appalachian geosyncline.

3. In many parts of the Shield and Marginal Segments there is conclusive evidence for post-Ordovician normal faulting

and substantial structural subsidence (including post-Silurian movements in the Timiskaming Depression and on Anticosti Island, and post-Carboniferous movements in parts of the Esquiman Channel Area). There are also numerous indications that the post-Ordovician tensional structures were superimposed on older structures, presumably also of tensional origin. Therefore, with regard to ages, it is convenient to divide the St. Lawrence Valley system structures into two age groups (i) post-Ordovician structures (ii) ancestral structures.

4. Evidence is equivocal as to whether the outer part of the Laurentian Channel is underlain by a structure. The hypothesis that it is underlain by a fracture zone (of uncertain type), linking the Marginal Rift Segment Zone with the Newfoundland Fracture Zone, is attractive in the contexts of regional geology and symmetry considerations of the St. Lawrence Valley system structures.

The Shield Segment A grabens and the Marginal Segment Rift Zone are connected. A connection between the Shield Segment B (Saguenay) graben and the Marginal Segment Rift Zone is very probable. Now, if we assume that the Shield Segment grabens and the Marginal Segment Rift Zone are parts of a single structure an interesting pattern emerges. The pattern (Fig. 41)closely resembles the branching pattern of a large continental rift system such as the Rhine graben. The similarity of patterns becomes complete if we include in the St. Lawrence pattern a branch structure along the outer part of the Laurentian Channel. For this pattern, which at least in part is underlain by rift structures, the name St. Lawrence Rift system is proposed (Kumarapeli and Saull 1966a). By analogy with large continental rift systems (see Fig. 42), the trunk rift zone of the St. Lawrence system appears to be the Marginal Segment Rift Zone along the St. Lawrence Valley and along the inner part of the Laurentian Channel (Fig. 41). In the following sections, the various characteristics of the St. Lawrence Rift system are compared with those of the three classical rift systems in East African, the Rhine and the Baikal regions.

If a comparison of the St. Lawrence Rift system with other large continental rifts is to be seen in correct perspective the following points must be taken into account. The St. Lawrence Rift system is an inactive structure except possibly for mild seismic strain release. The present day tectonic stresses in the St. Lawrence Region appear to be horizontal compressional rather than tensional (Voight 1969). There is no evidence that the rift system was active (in the sense of the Cenozoic activity of the classical rift systems) in the Cenozoic; the latest episode of magmatism along it took place in the Cretaceous (Monteregian magmatism). In contrast, the three classical rift systems of the world are active systems characterized by Cenozoic tectonism and volcanism (Beloussov 1969).

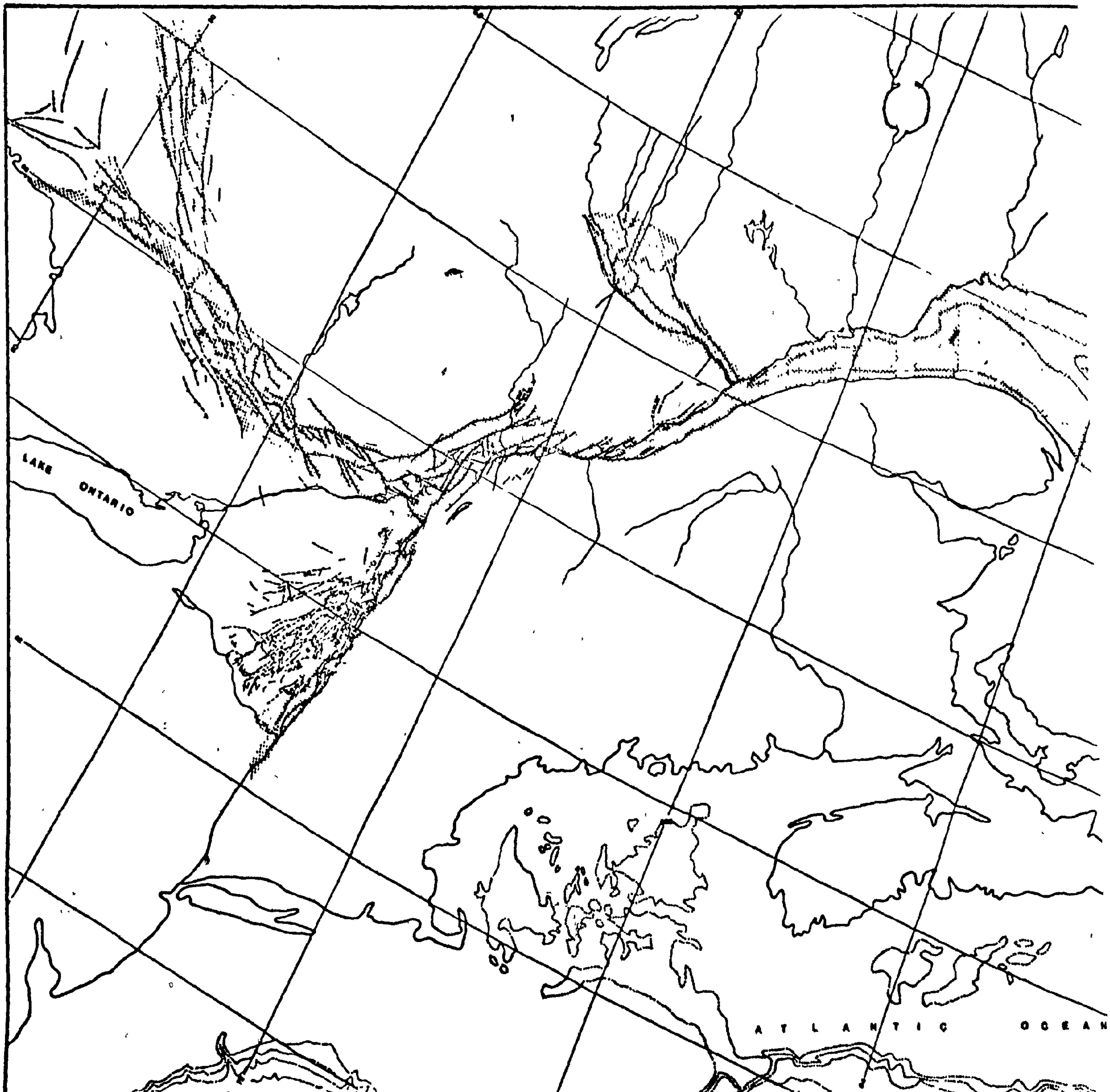

Fig.41. The rift zones of the St.Lawrence system.

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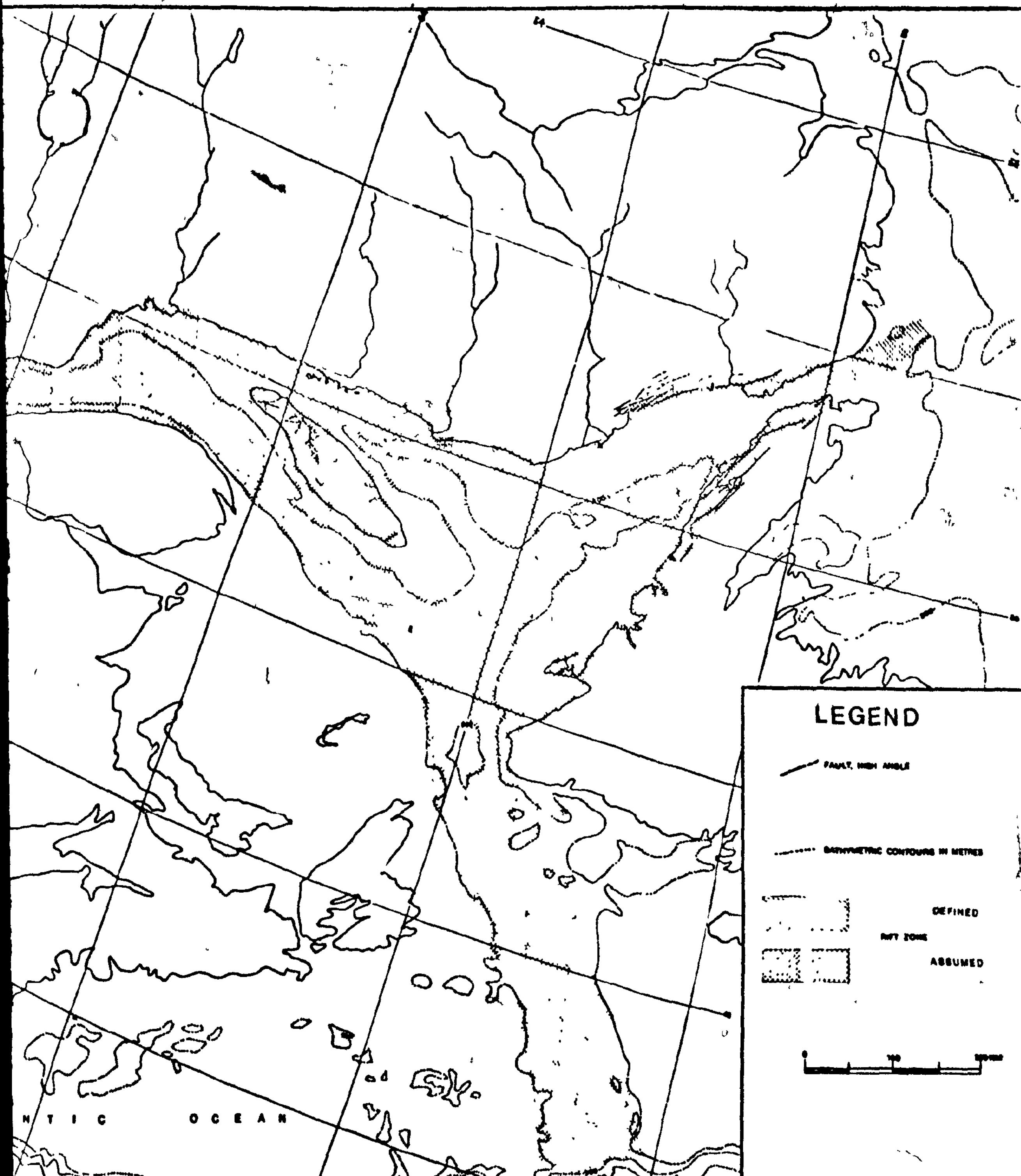
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POSSIBLE RELATION TO A CRUSTAL UPWARP

Grabens are generally located in crestal regions of upwarps (Cloos 1939). For example, small but well-defined grabens occur along crests of some anticlines and in oval uplifts related to diapiric intrusions of salt. Graben structures also occur along axes of upwarped fold mountain belts (e.g. in the Appalachians; Bain 1957) and along the crestal regions of mid-Oceanic rises (Heezen 1960). Similarly, large continental rift systems also traverse crustal swells or upwarps, many thousands of square kilometers in area. For instance, the northern part of the East African Rift system (sometimes called the Eritrean graben system) is set in a vast crustal swell about 2000 km wide (Fig. 42B). It exposes a crystalline Precambrian core which is surrounded by Mesozoic and younger sedimentary rocks. The central part of the upwarp forms a plateau at an altitude of 2 to 3 km. The altitude in the peripheral areas is about 1 km. The maximum structural uplift of the Precambrian basement is estimated at 5 km (Beloussov 1969). The upwarp is believed to have formed in late Eocene, just before the grabens began to develop (Mohr 1961 in Beloussov 1969). Similar upwarps also form the habitats of the East African Rift system proper, and of the Rhine and Baikal Rift system (Beloussov 1962, pp. 583-584). Largely as a result of this distinctive setting, the rims of many grabens are higher than the adjacent highlands. However, some grabens especially those at the ends of rift systems (e.g. Gulf of Suez, Gulf of Aquaba, Roer and

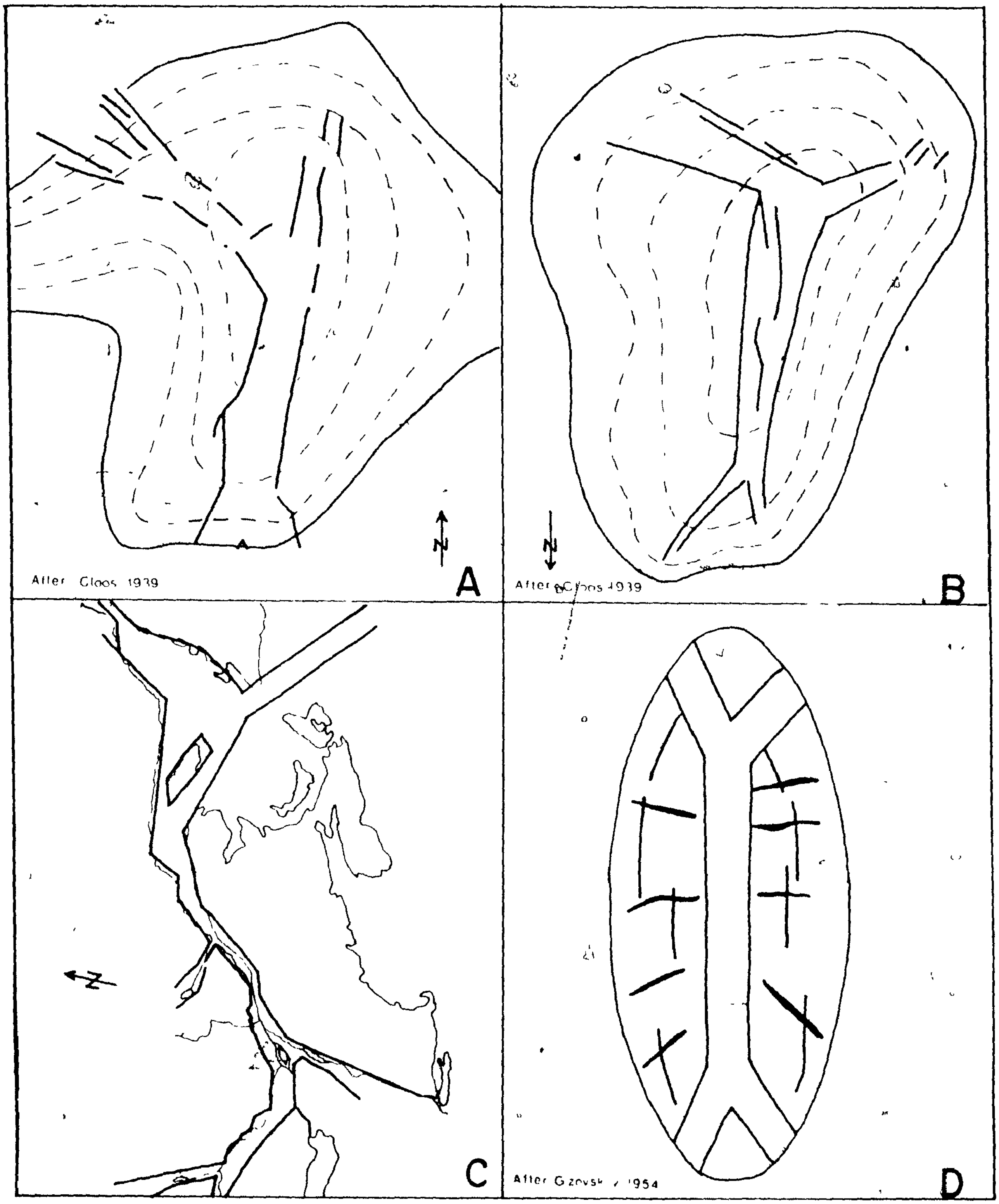

Fig.42. Diagram illustrating the similarity of graben patterns (A)Rhine graben (B)Eritrean graben (C)St.Lawrence Rift system (D)Fracture patterns in an oval uplift experimentally produced by M.V.Gzovskiy. In (A) and (B) are shown schematic structural contours of the Eritrean and Rhenish upwarps. The Eritrean graben is shown at nearly one-tenth and the St.Lawrence Rift system nearly one-sixth the scale of the Rhine graben.

Ruhr Valley grabens) do not have elevated rims (Freund 1966, p. 330).

A crustal swell similar to those that form the habitats of the classical rift systems cannot be discerned in the St. Lawrence Region. If such a feature did exist during the early development of the rift system, and has since been lost by erosion, then by analogy with the situation in other areas, the crustal swell may have been elliptical in outline with the trunk rift zone along its long axis and about a half of the crustal swell would have been on the Shield Region and the rest on what is now the Appalachian Region. There are, in fact, some features of the Shield Region that are compatible with the concept of a deeply eroded crustal swell as speculated above.

1. The Shield Region has undergone exceptionally deep erosion. Grenville rocks that formed in a catazonal environment, possibly at depths of 15 to 20 km (Wynne-Edwards 1972, p. 322), are now exposed. In fact, the Grenville Province (which is almost co-extensive with the Shield Region) despite being the youngest structural province of the Canadian Shield is the most deeply eroded. It is suggested that this unusually deep erosion was probably linked with crustal upwarping in the St. Lawrence Region.
2. The deep drainage lines (of the Shield Region) which appear to be fracture controlled (see p.42) form a rather distinctive pattern resembling the lateral fracture pattern that develops on an oval uplift (compare Fig. 42D with Fig. 43).

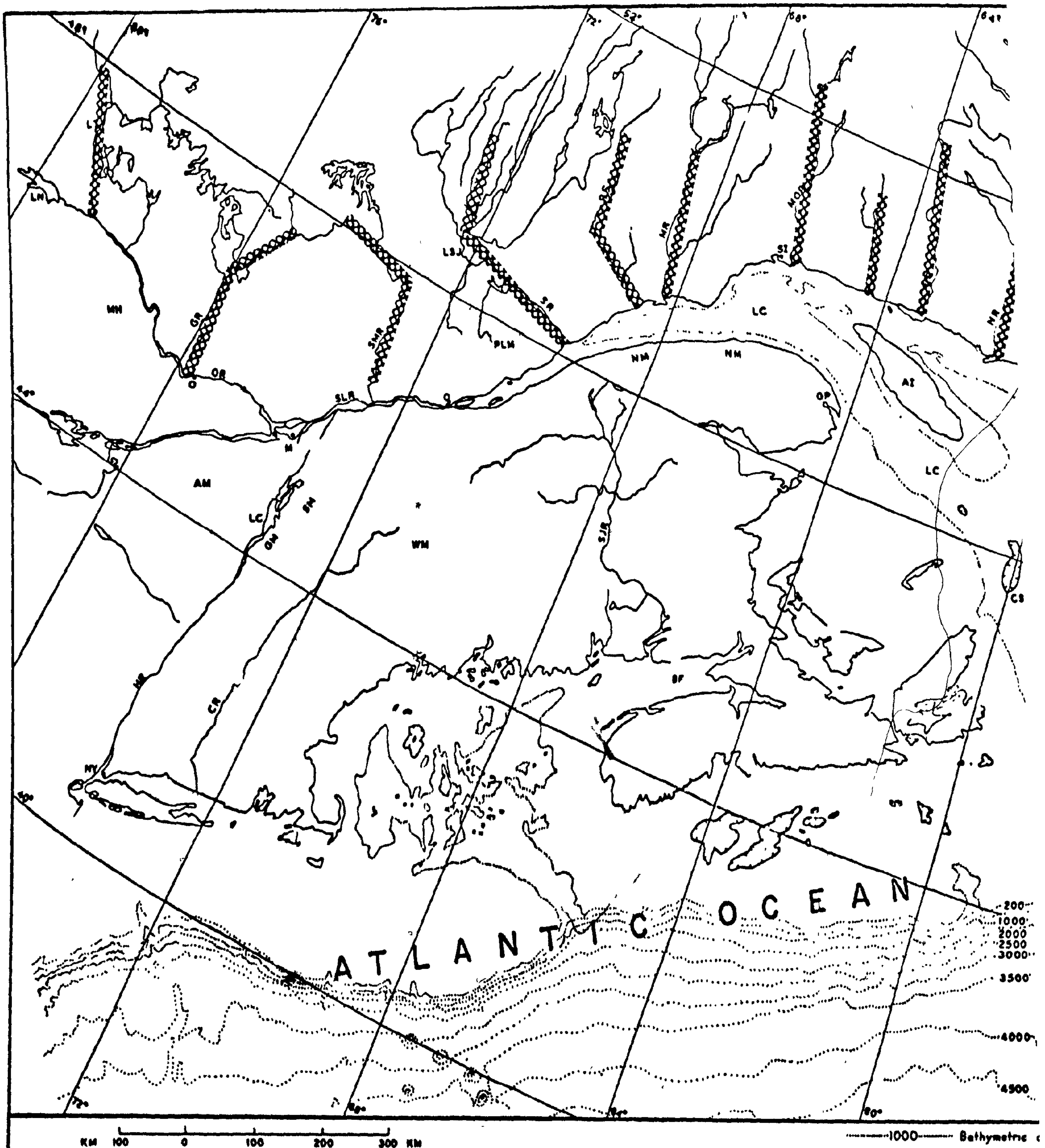

Fig.43. The pattern of deep valleys of the Shield Region

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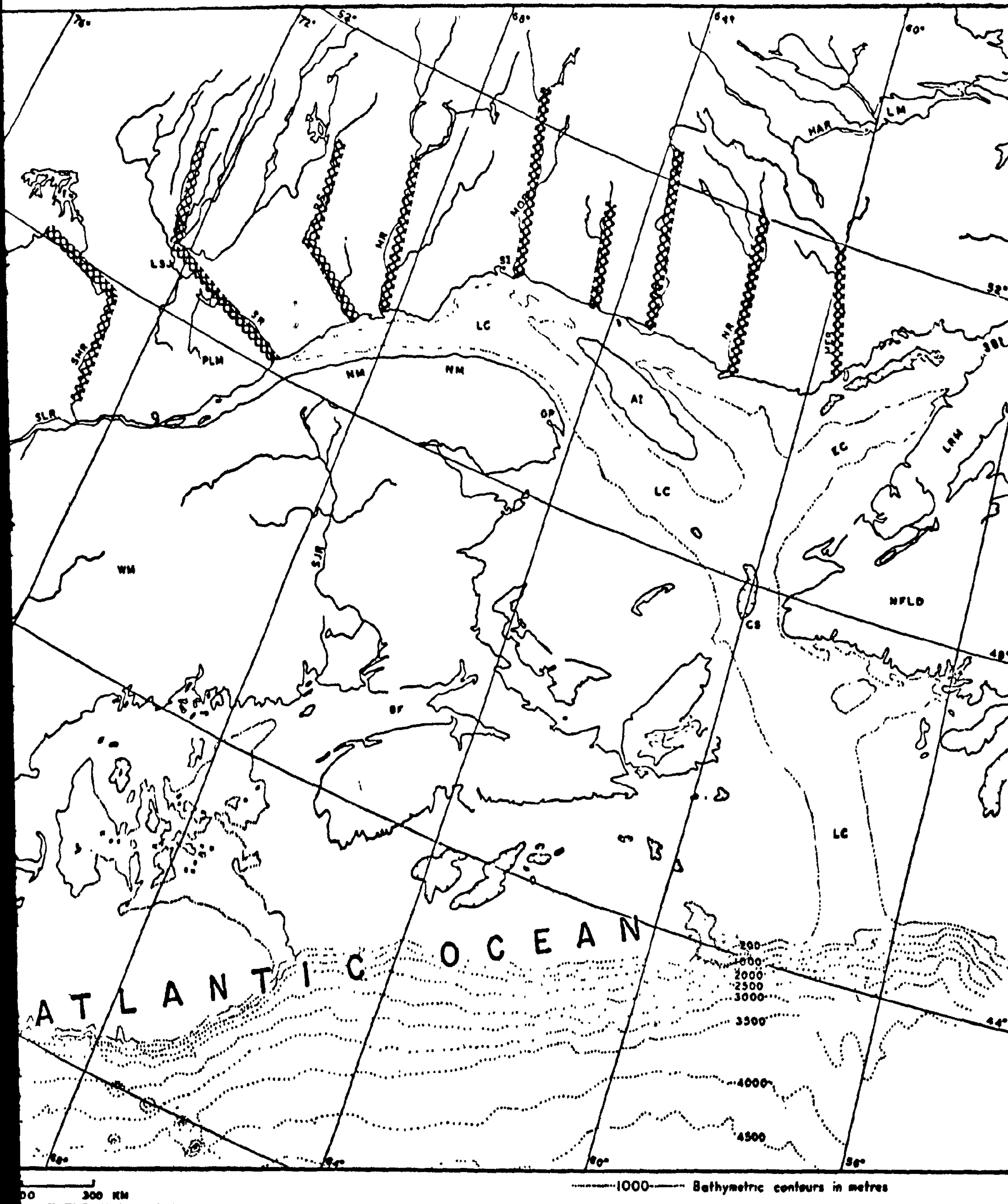


...eys of the Shield Region

3. In the Shield Region, the main rivers are deeply entrenched and have their sources to the north of the present height of land but flow southwards through canyons and gorges, against the regional slope for considerable distances (Appendix II). These features can be satisfactorily explained by assuming that the streams originated as vigorous consequent streams on lateral fracture zones of the north scarp of the trunk rift valley and they later extended (also along lateral fracture zones) ahead of the height of land by headward erosion.

What appear to be the resurrected parts of the eroded crustal swell, are the block mountains in the Parc des Laurentides, Adirondacks and Madawaska Highlands areas. It is interesting to note that the uplifted blocks in these areas are tilted away from their bordering depressions giving the depressions elevated edges.

Apparently, the deep valleys were established and structure adjusted in pre-Ordovician times (Ambrose 1964) and the deep erosion of the Grenville Province to the catazone was complete about 800 m. y. ago (Wynne Edwards, 1972, p. 322). Therefore, if a crustal swell did exist, it may have been related to the ancestral structures (see p.194) rather than to post-Ordovician rifting. It is possible that the presumably skeletal resurrections of the crustal swell in the Parc des Laurentides, Adirondacks and Madawaska Highlands areas are related to post-Ordovician rifting.

THE PATTERN OF THE ST. LAWRENCE VALLEY SYSTEM

The classical rift systems produce rather distinctive patterns on continental plates. These patterns are remarkably similar, demonstrating perhaps, the similarity of mechanical factors related to their formation. This similarity seems to be virtually independent of the scale of the patterns. For example, Figure 42 shows the resemblance between the Eritrean and Rhine grabens, although the former is about ten times the size of the latter. The patterns are similar to the fracture patterns experimentally produced by M.V. Gzovskiy (in Beloussov 1962) on a model of an oval uplift (Fig. 42D). The longitudinal fractures are the first to form, followed by lateral fractures. Where the uplift plunges, the longitudinal fracture system splits into two branches. As seen from Figure 42C, the pattern of the St. Lawrence Rift system is similar to the pattern produced by Gzovskiy. The longitudinal fractures are represented by the Marginal Segment Rift Zone along the St. Lawrence Valley and the inner part of the Laurentian channel;the lateral fractures by the Saguenay graben and the possible fractures along deep drainage lines of the Shield Region (Fig. 43); the main bifurcations of longitudinal fractures are represented by the splitting of the Marginal Segment Rift Zone at the south end of the St. Lawrence Valley and east of Anticosti Island.

The branching pattern of the St. Lawrence Valley fault system with the Ottawa and Champlain Valley faults resembles closely the fracture pattern produced by Cloos (1939) by swelling and oval shaped

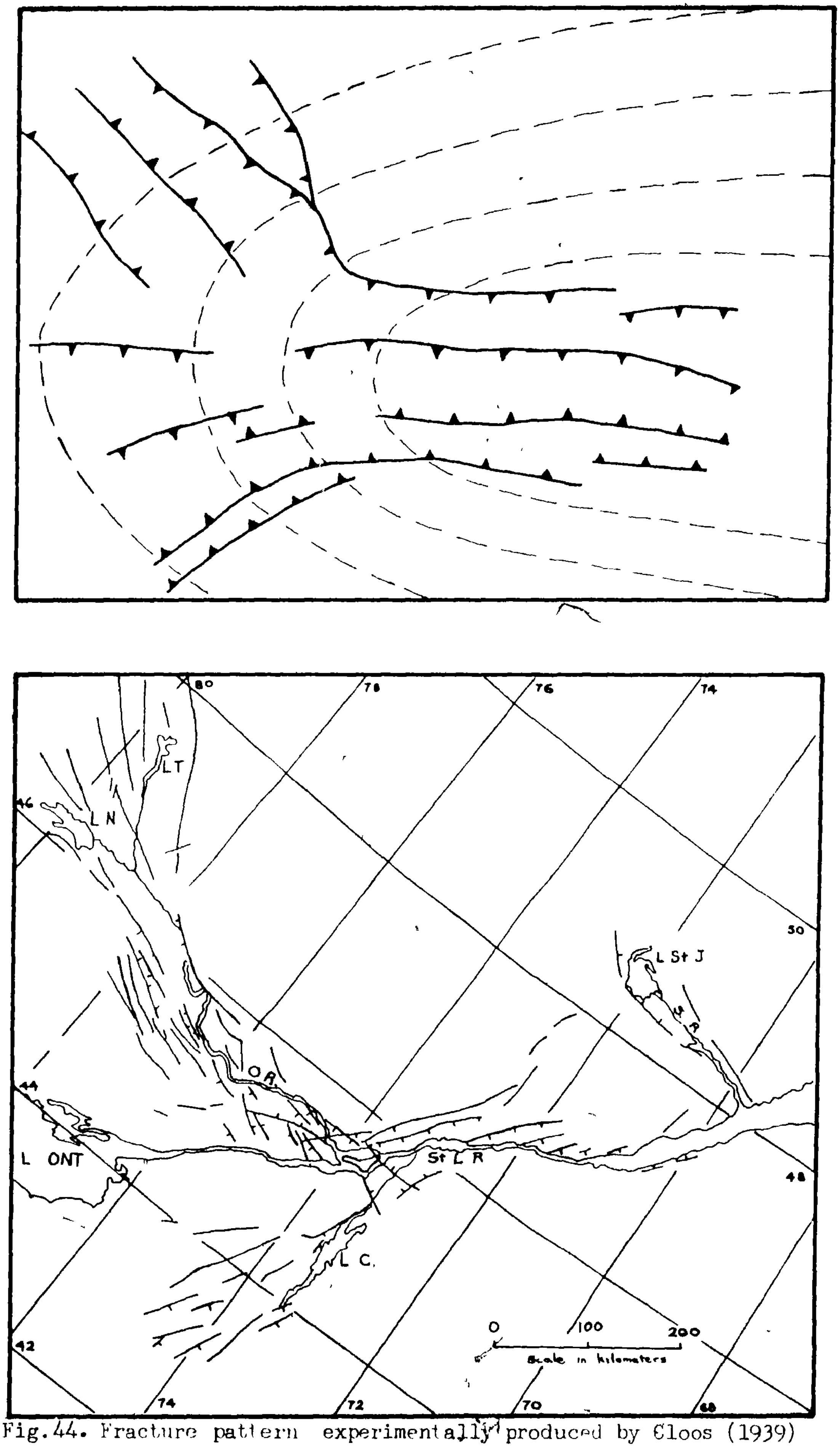
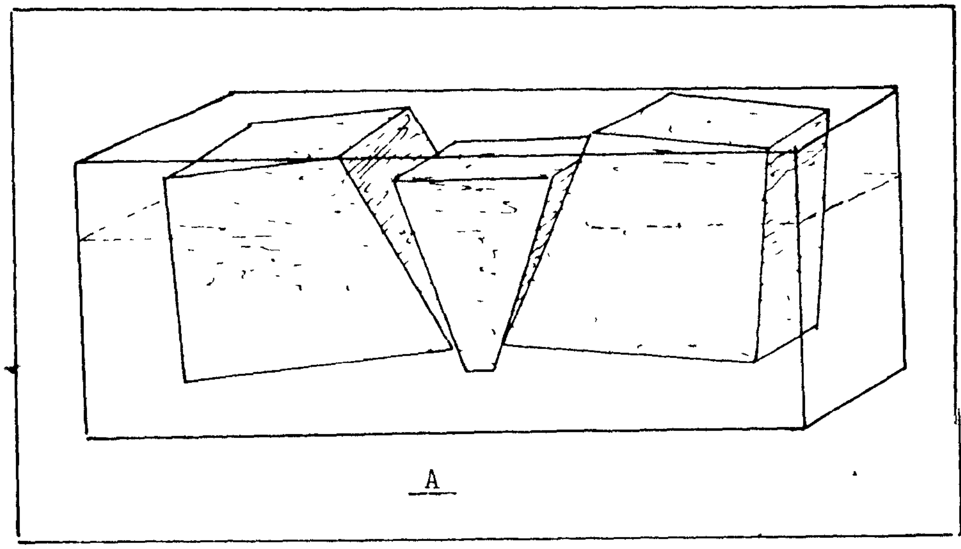


Fig. 44. Fracture pattern experimentally produced by Cloos (1939) by swelling an oval-shaped hot water bottle coated with moist clay (upper diagram) compared with the fault pattern at the south end of the St. Lawrence Valley.



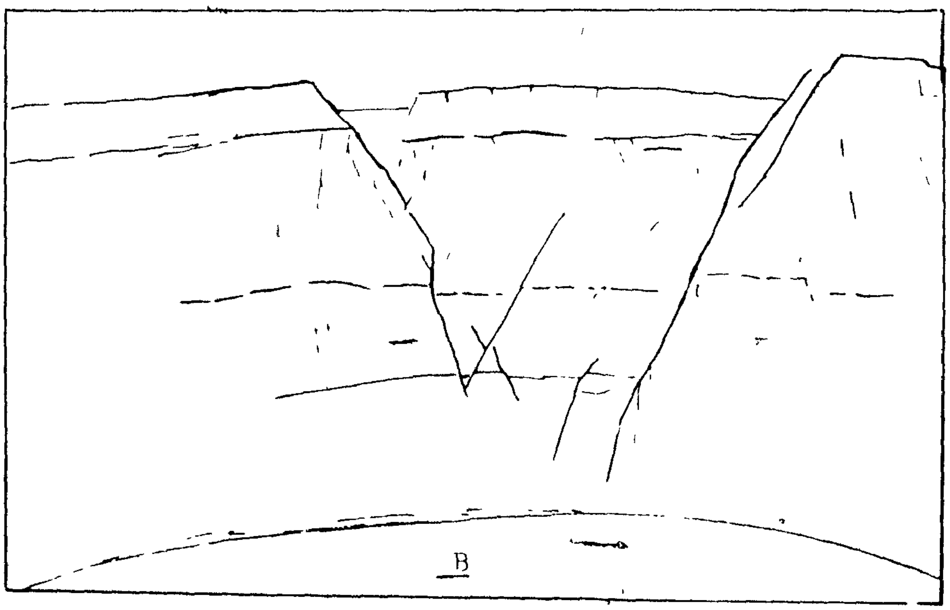


Fig.45A. Wooden blocks floating in water to illustrate the formation of a rift valley and the consequent tilting of the marginal blocks (after Taber 1927).

Fig.45B. Experimental graben produced in a cake of clay arched over a balloon (after Hans Cloos 1939).

hot water bottle coated with moist clay (Fig. 44). Thise close resemblance is compelling evidence that the Marginal Segment Rift Zone and the Shield Segment A graben system are casually related.

DIMENSIONS

Width

The rift zones of the three classical areas are similar in width. Generally, the widths range from about 35 to 60 km. These widths are of the same order of magnitude as the thickness of the continental crust, a relationship also brought out in the model experiments of Cloos (1939). To explain this relationship, Holmes (1964, p. 1061) advanced the simple theory that when the crust is bent slowly, it generally cracks into pieces having average width about the same as its own thickness. Because of the wedge shape of the crustal blocks, however, the width at the surface would be somewhat greater than the thickness of the crust.

Taber's (1927) model of "floating blocks" (Fig. 45A) was one of the earliest adopted for explaining the various features of grabens. Despite its many drawbacks (see Freund 1966, p. 333) it is still widely used, especially by geophysicists (for example see Girdler 1964, p. 720). Adopting this model, Vening Meinesz (1959, in Freund 1966, p. 333) computed that for a crust 35 km thick, the surface width of a graben should be about 65 km.

Table II shows the similarity of widths between some well known continental grabens and the various parts of the St. Lawrence Rift system.

Table II: Widths of Continental Rifts

Rift	Widths - km
Gulf of Aquaba	50
Dead Sea	35
Gulf of Suez	35
East African (common)	30
Rhine	40
Baikal	50
Ottawa graben	
Lower Ottawa Valley	60
Upper Ottawa Valley	55
Nipissing graben	30
Timiskaming graben	30
Saguenay graben	55
Marginal Segment Rift Zone	
St. Lawrence Valley	30 ?
Champlain Valley	35 ?
Esquiman Channel	150 ?

Data on the East African, Rhine and Baikal rifts are from Beloussov (1969). The parts of the St. Lawrence Rift zones listed in the Table II vary in width. In general they are wider at one end and narrow gradually towards the other end. The values given are those of the medial sections. The width of the Marginal Segment Rift Zone is uncertain.

Length

The exact lateral extent of the three classical rift zones is imperfectly known. Near the end of an individual graben, its main faults usually split into several divergent smaller faults - the so-called "splays". The graben structure itself becomes wider and shallower and its topographic expression becomes indistinct. The rift faults may, however, continue beyond this area, for a considerable distance before dying out. Even excluding such possible continuations, the lateral extent of the large continental rift systems is impressive. For example, the East African Rift system, the largest of the continental rifts, extends from the Turkish-Syrian border on the north, to the lower course of the Zambesi River on the south, a distance of over 6500 km. The Baikal rift zone is shorter, yet it extends for a distance of about 2500-3000 km (Florensov 1966, p. 175) from northwestern Mongolia to South Yakutia. The Mittelmeer-Mjoesen Zone, as envisaged by Hans Stille (see Bederke 1966, p. 214) is also about 2000 km long; its better known parts, the Rhine and Oslo grabens are each about 300 km long.

The termination of rift structures of the St. Lawrence Valley system are also poorly known. For example, it is not certain where the Timiskaming graben faults terminate. They may continue several tens of kilometers beyond the northern limit of the depression (see Lovell and Caine 1970; Parkinson 1962, p. 95). Also, the Nipissing graben faults appear to extend into the north channel of Lake Huron (see p. 147) but where they terminate is conjectural. The postulated branch structure along the outer part of the Laurentian Channel might be continuous with the Newfoundland Fracture Zone (see Le Pichon and Fox 1971), which can be traced as far as the mid-Atlantic ridge. Excluding these possible extensions, the St. Lawrence Rift system as measured from the north end of the Timiskaming Depression to the northeast end of the Esquiman Channel Area is about 2200 km long.

GEOMORPHIC FEATURES

General Features

Topographically, the rift zones of the classical areas are commonly reflected as deep broad valleys with steep edges. A chain of lakes may sometimes occupy the valley floor. The bottoms of some of the lakes are well below sea level. For example, the bottom of Lake Tanganyika is about 650 m below sea level. Elsewhere, the valley floors for the most part are flat and are remarkably even. This evenness is commonly due to sedimentation in formerly extensive lakes masking the irregularities on bedrock. The most common interruptions of the plain-like topography of the valley floors are longitudinal asymmetrical ridges reflecting uptilted edges of subsidiary fault blocks, block mountains of various shapes and land forms of igneous origin.

Compared with rift valleys of the three classical areas, those of the St. Lawrence System are shallower and often poorly defined. These differences can be attributed to a combination of long inactivity and deeper erosion of the rift zones. Four large lakes: Timiskaming, Nipissing, Champlain and St. Jean occupy parts of the St. Lawrence Valley system. The bottom of Lake Champlain lies about 100 m below sea-level;the other lakes have their bottoms slightly above sea-level. Mention should be made here, that immediately after the Wisconsin glaciation, much of the now emerged parts of the St. Lawrence Valley system were occupied by the transient Champlain Sea. It lasted for about 1500 years and gave way to a lacustrine phase (mainly in the St. Lawrence Valley) which also was short-lived (Elson 1969).

Large parts of the St. Lawrence and Lower Ottawa Valley floors are nearly flat plains that are partly a reflection of the near horizontality of the bedrock strata and partly due to masking of bedrock irregularities by Champlain Sea sediments. The flat terraces around Lake Champlain and Lac St. Jean show former lake levels. Positive topographic features of tectonic and igneous origin can be found in the following areas of the St. Lawrence Valley system: asymmetrical longitudinal ridges in the Upper Ottawa Valley (the Muskrat, Doré, Eganville, Shamrock and Pakenham scarps; see p. of this thesis); block mountains in the southwestern part of the St. Lawrence Valley and the adjacent areas of the Lower Ottawa Valley (Oka and Rigaud Mountains, St. André Hills); hills related to igneous intrusions in the southern part of the St. Lawrence Valley (Monteregian Hills).

Fault Line Scarps

Fault line scarps, especially those related to the boundary faults of rift valleys, are perhaps the most impressive features of rift valley scenery. Some of the scarps of the East African Rift Valleys are old and deeply dissected by erosion, but the more recent ones are steep and sharply defined. A few are noted for their height. For example, the Livingstone Mountains scarp, on the east side of Lake Nyasa is about 2000 m in height (Dixey 1956, p. 17). More commonly, however, the scarps are not unusually high but yet they appear as imposing features of the landscape because of their characteristic setting against the flat valley floors.

Fault lines scarps of great height are not associated with the St. Lawrence Rift system. This feature also probably reflects the long inactivity and deep erosion of the rift system. Some of the scarps, although deeply dissected and rarely exceeding 500 m in height, are abrupt and appear linear for long distances on small scale maps. Such are the Coulonge and Grenville scarps on the north side of the Lower and Upper Ottawa Valleys respectively (Figs. 46 & 47), and the Laurentide scarp in the area of Parc des Laurentides massif (Fig. 48) Other well-defined scarps are the St. Patrick scarp (Fig. 49) on the south side of the Upper Ottawa Valley and the Adirondack scarp on the west side of the Champlain Valley and the Long Range scarp to the southeast of the Esquiman Channel Area.

Block Mountains

In places elevated fault blocks form block mountains within rift valleys. The highest mountain of this type is Ruwenzori (5125 m), which is by far the highest non-volcanic mountain in Africa. It occurs

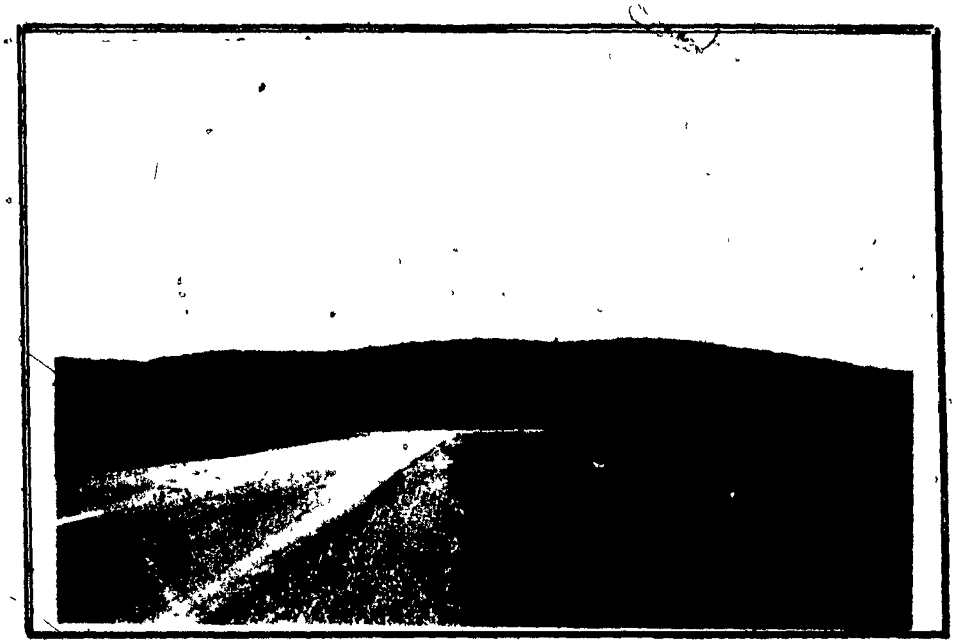


Fig.46. View of the Coulonge scarp, looking north from Chapeau (Allumette Island).

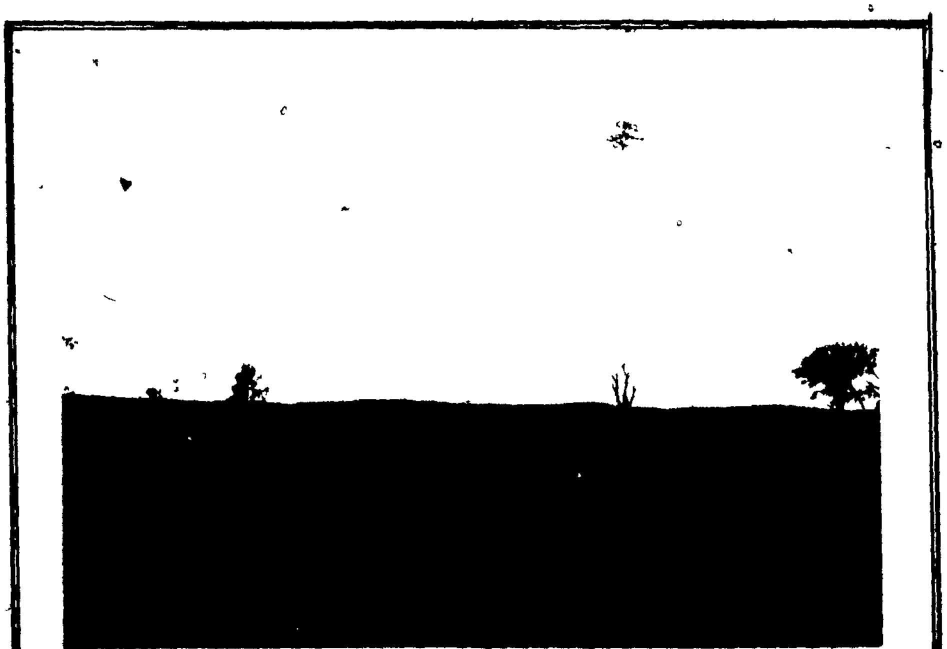


Fig.47. View of the Grenville scarp at Fassett (about 5 km east of Montebello), looking north from Highway 17.

Fig.48. View of the Laurentide scarp, looking northwest from the southeast shore of the St.Lawrence River near Montmagny, about 50 km northeast of Quebec City.

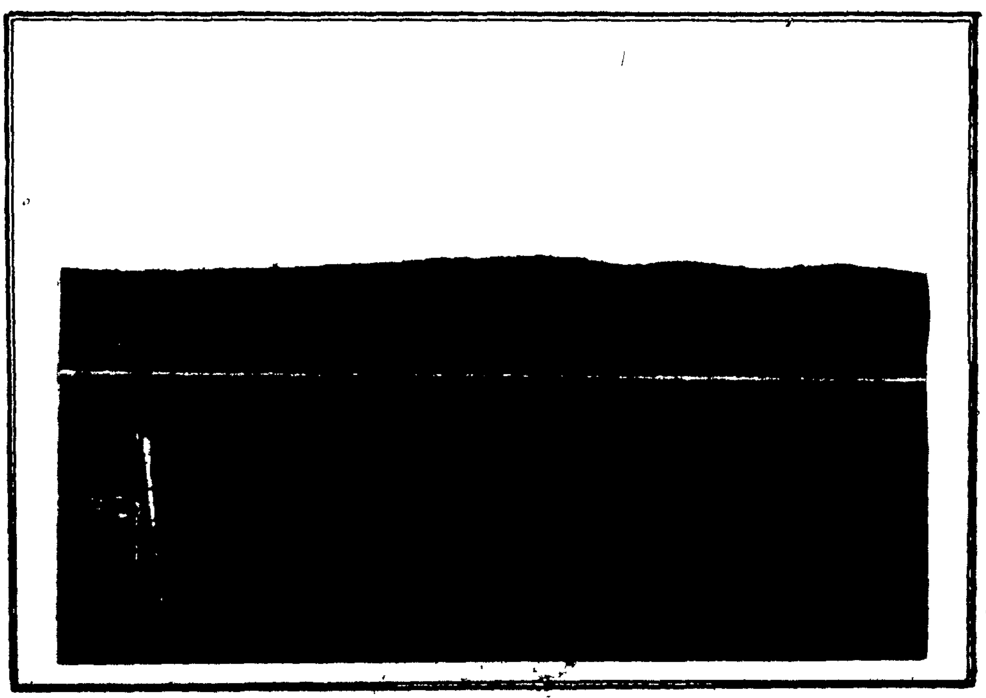


Fig.49. View of the St.Patrick scarp, Looking southwest from Cormac about 5 km west-northwest of Lake Clear.

at the bifurcation of the western Rift Valley and rises much higher (about 4000 m) than any other feature in the general area. Other block mountains of similar setting are Mt. Lebanon, the Danakil Mts. in Ethiopia and Mt. Mebya in Tanzania.

Rigaud and Oka Mountains at the southwest bifurcation of the Marginal Segment Rift Zone appear to be block mountains of the type mentioned above. However, they stand only about 200 m above the general level of the valley floor. Although they are dwarfed by the immensity of the Ruwenzori massif, yet their tectonic setting and structure appear to be quite similar to those of the latter.

The acute angle between the two branches of a bifurcating rift may also be occupied by a block mountain (Brock 1966, p. 108). An example of this is the Sinai Peninsula. In the St. Lawrence Region, the Adirondack Mountains and the Parc des Laurentides massif appear to be block mountains of this type.

Block Mountains also sometimes border rift valleys. Examples are the Vosges and Black Forest areas on either side of the Rhine graben. In the St. Lawrence Region, block mountains of similar setting are the Madawaska Highlands and the northern part of the Long Range Mountains.

TECTONIC SETTING AND TREND RELATIONS

The large continental rifts of the three classical areas are characteristically superimposed on cratons. The East African Rift system is set chiefly in a vast shield area. The gross trend of the rift system is N-S, and the Precambrian basement in the area is also characterized by the same general trend. Despite this overall conformity of trends, notable discordancies between the strikes of the Precambrian rocks and of the rifts are observed when they are considered in detail. Trend relations are essentially similar in the Baikal region (Beloussov 1969, p. 543). The Rhine graben on the other hand is discordantly superimposed on Paleozoic and Precambrian structures. The gross trend of the graben is about 20° east of north, but different parts of the graben trend north-northeast and northwest, whereas the older rocks in the area strike northeast.

Apart from the postulated branch structure along the outer part of the Laurentian Channel, the St. Lawrence Rift system is set on craton. The gross trend relations are similar to those of the East African and Baikal rift systems in that:

1. the gross trend of the St. Lawrence Rift system conforms with that of the Grenville belt on the one hand and that of the Appalachian belt on the other,
2. in detail there are notable discordancies between the trends of the rift zones and those of older rocks; an extreme example is the extension of the Timiskaming graben across the Grenville-Superior boundary.

With regard to tectonic setting and trend relations, the Rhine graben shows similarities to the Shield Segment grabens, rather than to the St. Lawrence Rift system as a whole:

1. the Rhine and the Shield Segment grabens are superimposed on older rocks discordantly,
2. their trends are at right angles to the associated foldbelts; Alpine belt in the case of the Rhine graben and Appalachian belt in the case of the Shield Segment grabens.

STRUCTURE

Structure of Continental Rift Zones

Some of the points which arose from the 1965 UMC/UNESCO seminar on the East African Rift system were as follows:

1. The structure of the rift system is highly complex and variable along its length, so much so, that practically no generalization can be made as to the structure of the entire rift system (Dixey 1965, p. 123). The grabens that make up large segments of the rift system are of the complex type.
2. In addition to segments characterized by graben structures, broad block-faulted and/or downwarped areas are included in the rift system. For instance, in northern Kenya and northern Tanzania, the rift faults diverge and broad downwarped areas are customarily included in the rift system (Baker 1965, p. 82; Pallister 1965, p. 87).
3. On the surface at least, noteworthy discontinuities are apparent in the rift system (e.g. see Tertiary fault map of Tanzania, Pallister 1965, p. 90).

In attempting to compare the gross structure of the St. Lawrence Rift system with that of the classical rifts, a point to be reckoned with is that no less than half of the St. Lawrence Rift system is submerged, and the precise structure of these parts is poorly known. Also some of the parts that are not water-covered have not been studied to the extent that their detailed structure is known (e.g. Saguenay graben). Based on the structure of the better known parts, the following conclusions can be made. The rift system is made up of complex grabens and graben-like zones of downfaulting. Even if a relatively short segment of the rift system is considered, the structure is variable along its length. For example, near Ottawa the structure of the Upper Ottawa Valley is a well-defined complex graben but near Mattawa the structure appears to degenerate into a single tilted-block (compare Figs. 24 and 25). Where rift faults diverge as in the northern part of the Timiskaming Depression, the western part of the Nipissing Depression and the northwestern part of the Saguenay-Lac St. Jean Depression, broad downwarped areas are included in the rift system. Thus, as far as is known, the gross structure of the St. Lawrence Rift system is similar to that of the three classical rifts.

Structural Configuration of Grabens

(a) Transverse configuration

In the East African Rift system, grabens commonly have asymmetrical cross-sections. In fact, symmetrical, two-sided grabens of the type experimentally produced by Cloos (1939) are relatively rare. In a number of cases as in parts of Ethiopia, Kenya, Nyasa and other rifts, a major fault is present only on one side, and the other side is controlled by down-flexing or hinge-faulting; one shoulder may be uplifted considerably more than the other (Dixey 1965). In the Baikal system too, Florensov (1965, 1969) emphasized the structural asymmetry of grabens.

All the better known grabens of the St. Lawrence system appear to have asymmetrical cross-sections.

The only near-symmetrical structure is the Ottawa graben between Ottawa and Pembroke. Extreme structural asymmetry is shown by the Timiskaming graben. In the neighbouring parts of the Ottawa and Nipissing grabens a major fault seems to be present only on the north side, the south side being controlled presumably by hinge-faulting.

(b) Longitudinal configuration

Structural subsidence along trains of grabens that constitute large graben systems vary along their length. Holmes (1964, p. 1065), has pointed out that at least some parts of the East African Rift system consist of cradle-shaped grabens (see Fig. 50) separated by transverse "arches". This is probably why a long rift valley such as the Western Rift Valley of East Africa is occupied by a chain of lakes (Lakes Albert, Edward, Kivu and Tanganiyaka) instead of one long lake.

The St. Lawrence Rift system also contains cradle-shaped rifts. For example, the Marginal Segment Rift Zone is cradle-shaped between the Beauharnois axis and the rise of the basement in the Quebec City area. The Ottawa graben is also cradle-shaped, the deepest part being around Ottawa. The Timiskaming, Nipissing, Champlain and Saguenay grabens may also be cradle-shaped with their deepest parts occupied by lakes, although this cannot be demonstrated with any certainty.

As Holmes (1964, p. 1065) has pointed out, the formation of cradle-shaped grabens separated by transverse "arches" is dictated by the spherical shape of the earth. The top of a narrow strip of earth's crust on which a long chain of grabens is superimposed, has the shape of an arc of a circle. With continued subsidence, such a strip would tend to settle towards the corresponding chord. But as the chord is shorter than the arc, the strip may buckle into a wave-like form, the wave crests forming the transverse "arches". Figure 51 which includes a section through the trunk rift zone of the St. Lawrence Rift system, shows the wave-like configuration of the basement surface.

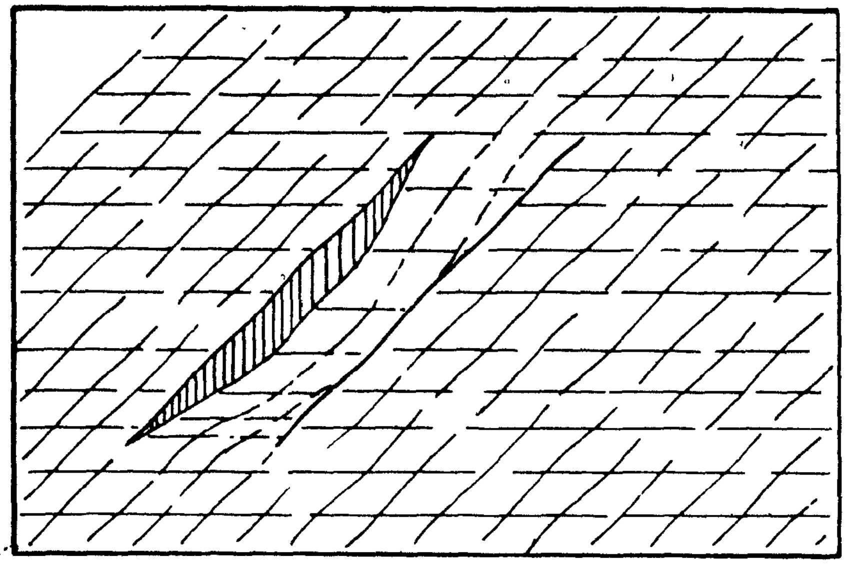


Fig.50. Diagram illustrating the concept of a cradle-shaped graben. After Beloussov (1962).

(c) Internal structure of graben blocks

The complex grabens that make up large segments of the classical rift zones commonly consist of combinations of longitudinal tilted blocks and/or small grabens and horsts. Similar structures characterize the St. Lawrence rifts, at least where the internal structure of rift blocks is known in sufficient detail. An excellent example of a complex graben consisting of a group of longitudinal tilted blocks is the Ottawa graben between Ottawa and Pembroke (Fig. 24). The Marginal Segment Rift Zone in the Champlain Valley area consists of a combination of minor grabens and horsts (Fig. 14). In the southern part of the St. Lawrence Valley and in parts of the upper Ottawa Valley, rift floors are broken into fault blocks of rhomboid shape (Fig. 52E,F).

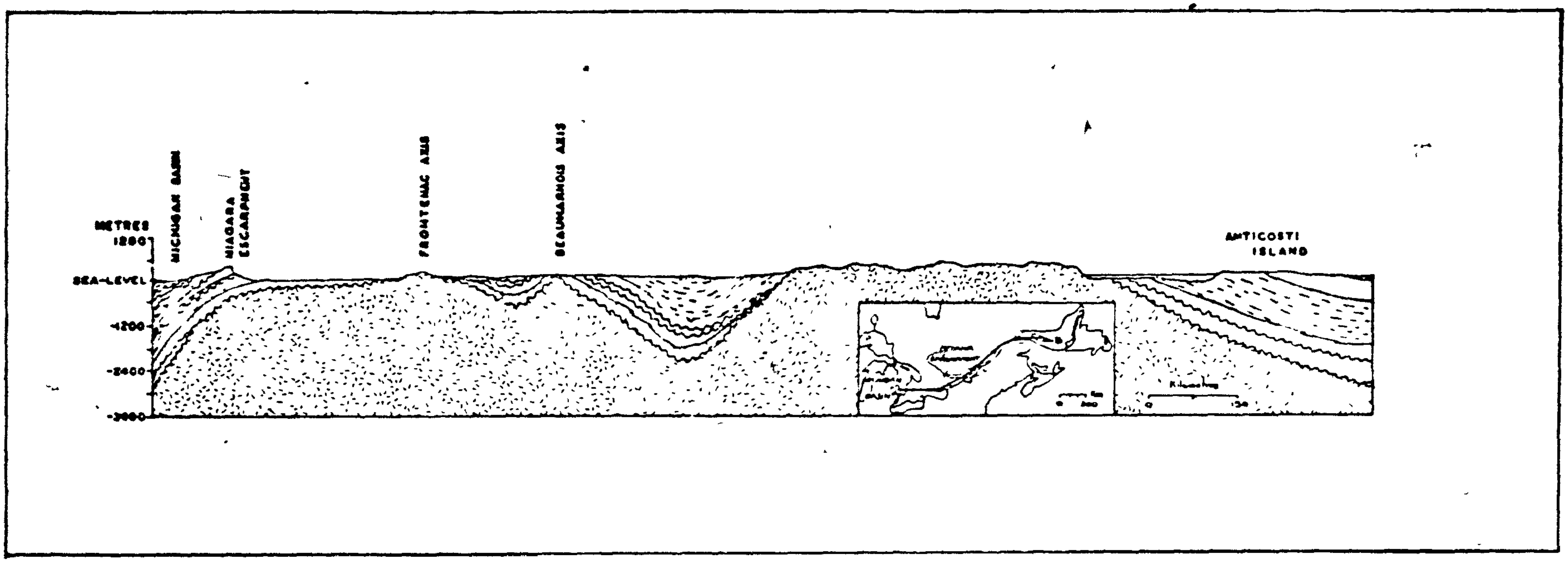


Fig.51. Wave-like surface of the basement along the St.Lawrence Rift system. Diagram adapted from B.V.Sanford in Poole et al.(1970)

Graben Depths

Because the grabens of the large rift zones are commonly cradle-shaped longitudinally, and are asymmetrical transversely, variable structural subsidence is involved. The maximum vertical displacement of the East African and Rhine rifts is estimated to be about 3000 m (Beloussov 1969, p. 539) and that of the Baikal rift system to be more than 7000 m.

The amount of structural subsidence along the St. Lawrence Rift system cannot be determined with any certainty because of deep erosion of the rift system and the fact that there are not many places where structural subsidence could be estimated. The following pieces of evidence are thought to give some idea of the magnitude of post-Ordovician structural subsidence.

1. In the Ottawa graben, the maximum structural subsidence of platformal rocks does not seem to exceed 500 m (see Fig. 22). Also, if it is assumed that (i) much of the Shield Region was once covered by Lower Paleozoic platformal rocks and (ii) no great lowering of the Grenville surface has taken place since its exhumation (Ambrose 1964), then the elevations at which Lower Paleozoic outliers are preserved within the grabens (e.g. Ottawa graben, see Fig. 24), when compared with the elevations of adjacent areas of the Shield Region, indicate that the total vertical movements does not exceed 500 m.

2. In the Marginal Segment, the stratigraphy of platformal rocks indicates that the regional slope of the Grenville surface (during the deposition of the platform cover) was towards the Appalachian geosyncline. This slope, although somewhat interrupted by post-Ordovician faulting has not been changed to any great extent (see Figs. 11 & 17). In fact the transverse structural asymmetry of the Marginal Segment Rift Zone appears to be largely inherited from the configuration, prior to post-Ordovician rifting, of the basement surface and of the platformal rocks. These features indicate that not more than a few hundred metres of structural subsidence can be attributed to post-Ordovician movements. Thus, it appears that the maximum post-Ordovician vertical movements along the rift system are small in comparison with the vertical movements of the classical rift zones. Larger vertical movements may have taken place during the evolution of the ancestral structures but there appears to be no way of determining the magnitude of movements related to these structures.

Faults and Fault Patterns

Faults that make up the classical rift systems (Freund 1966; Beloussov 1969) are either vertical or dip steeply in the direction of the downthrown side (i.e. normal faults). Reverse faults are rare, but strike-slip faults have been recognized from several parts of the East African Rift system (Freund 1966; Freund et al. 1970) notably along the

Dead Sea rift (Quennell 1959). Longitudinal faults are the most prominent. They produce step-like patterns on the sides of grabens. Transverse faults also have rather widespread occurrence. Faults often have curved traces with the maximum throw about midway of their length. An individual fault rarely extends for more than a few tens of kilometers before being replaced by another, usually in en echelon fashion.

The faults of the St. Lawrence Valley system, have been described earlier in Chapters IV to XII under the heading "high-angle faults". The close space association of normal faults, south- and east-dipping high-angle reverse faults and low-angle thrust faults, along the south and east sides of the Marginal Segment, can be explained in term of the marginal habitat of this zone with respect to the Appalachian foldbelt. Apart from this complicated zone, the St. Lawrence Valley system, as far as is known, is outlined by high-angle faults whose nature and arrangement are similar to those of the rift faults in the three classical areas. Some prominent examples of fault patterns are shown in Figure 52. Strike-slip movements of less than 2 km have been postulated on the northwest side of the Esquiman Channel Area (Davies 1968, pp. 270-271, also see p.92 of this thesis) and northwest end of the Saguenay graben (Philpotts 1965, p. 7, also see p.172 of this thesis). It is not certain whether these movements accompanied the rifting itself (see Freund 1966, p. 332), or are due to response of pre-existing faults to stresses related to orogeny in the nearby Appalachian geosyncline.

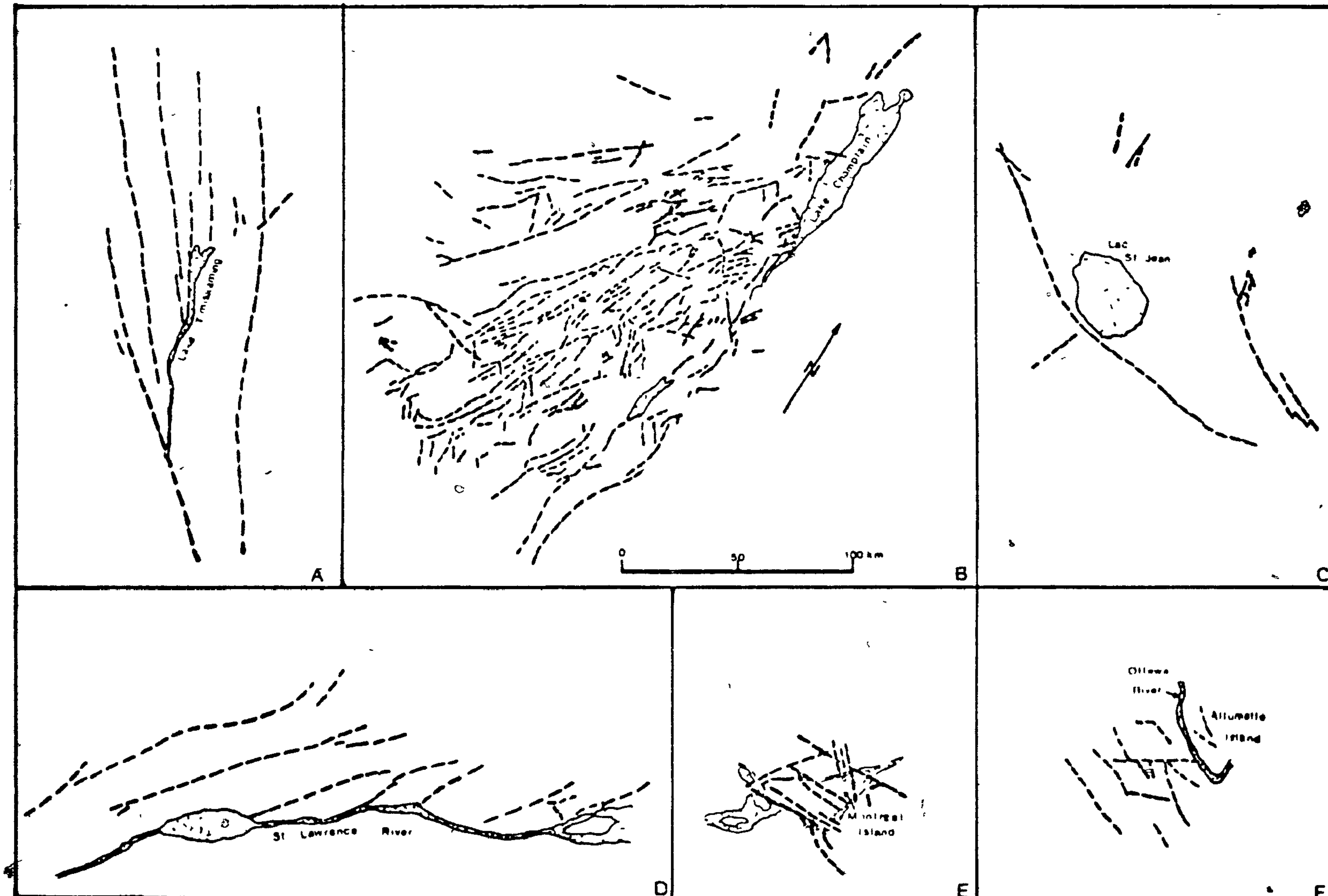


Fig.52. Some selected fault patterns of the St.Lawrence Rift system. A,B,C - divergent fault patterns probably representing splays. D - en echelon pattern. E,F - intersecting patterns. The scale and the north arrow in B apply to all parts of the diagram.

Control of Rift Faults by Older Structural Lines

Dixey (1965;also see Brock 1965) has pointed out that faults of the East African Rift system, which are Miocene and younger, frequently follow structural weaknesses determined by earlier faults (mainly Precambrian) and tectonic trends, particularly where they lay in the general direction of the later stress relief. Florensov (1966, p. 195) notes that in the Baikal Rift system too, the Tertiary and later faults are controlled by late Proterozoic and Caledonian structures.

In several areas of the St. Lawrence Rift system, there are indications that the directions of rift faults were controlled by a pre-existing fracture pattern, oriented prevalently northeast and northwest (see p.42). The rift faults and the rift zones themselves for the most part are oriented or tend to be oriented in one of the other of these two directions. The Marginal Segment Rift Zone, although nearly north-south trending in the northern part of the Champlain Valley, swings nearly to a northeast direction in the southern part of the Champlain Valley. Also, the Saguenay graben which trends nearly east-west between the Marginal Segment Rift Zone and Lac St. Jean, swings to a northwest orientation westwards. The two east-west oriented segments in the Lower Ottawa Valley and in Anticosti Island area appear to play the role of links between northeast and northwest oriented segments of the rift system. The Nipissing graben does not fit into the above scheme of northeast and northwest oriented segments linked by east-west oriented

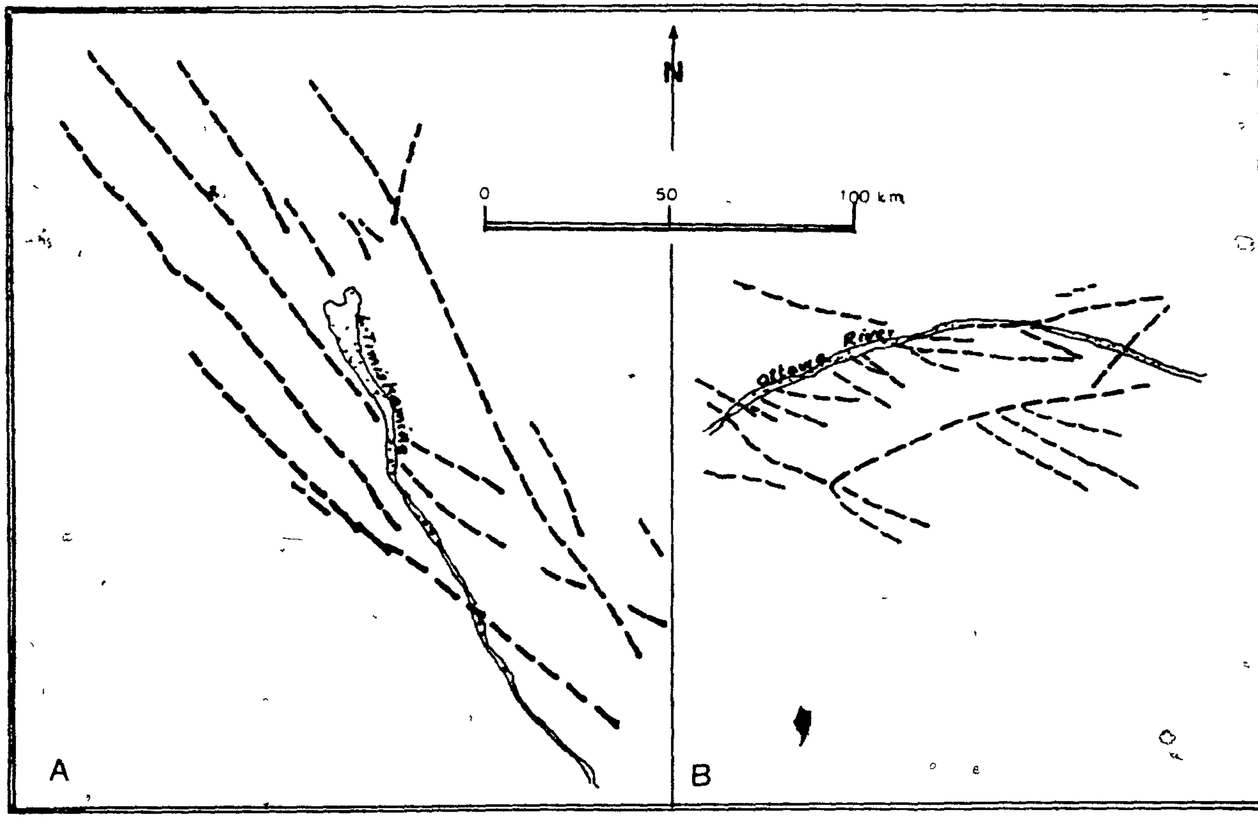


Fig.53. Diagram illustrating the relationship between rift orientation and the degree of development of longitudinal faults. Note that the longitudinal faults of the northwest trending Timiskaming graben (A) are strongly developed whereas the longitudinal faults of the nearly east-west trending Ottawa graben in the Lower Ottawa Valley area (B) are weakly developed.

segments. It is oriented approximately east-west but does not play the role of a link. Its direction might have been dictated by older east-west faults, perhaps by an eastward extension of the Murray Fault system (see Church 1972, p. 355). A feature inherited from northeast and northwest directional control of faults appears to be that in rift segments conforming to these directions, longitudinal faults are usually well developed (e.g. Timiskaming Depression, also Upper Ottawa and St. Lawrence Valleys), whereas in the segments that do not conform to these directions, the longitudinal faults are weakly developed (Fig. 53), and usually consist of relatively short <u>en echelon</u> faults (e.g. Nipissing graben, Champlain Valley, Lower Ottawa Valley).

SEDIMENTARY-VOLCANIC FILL

The fault troughs of the classical rift zones are filled in varying degrees by clastic sediments and volcanics of Cenozoic age. In the deeper parts of the fault troughs, the rift fill can be two to three thousand meters thick (e.g. see Doebl 1970). In one section of the Baikal Rift system, the thickness of sedimentary fill is estimated to be 5000-6000 m (Florensov 1966, p. 173; 1969, p. 452).

The fault troughs of the St. Lawrence Rift system have neither Cenozoic volcanics nor substantial thicknesses of sediments. Their floors are covered by a veneer of Pleistocene and recent sediments but these are usually less than 100 m thick. In the above respects, the St. Lawrence Rifts are similar to the Oslo graben (see Bederke 1966, p. 213) and in both areas the lack of a Cenozoic sedimentary-volcanic fill is consistent with the long inactivity of the rifts.

IGNEOUS ACTIVITY

Igneous Activity Associated with Continental Rifts

Many workers have noted the close association of volcanic activity with continental rifts (e.g. Holmes 1964, p. 1051; Freund 1966, p. 332; Beloussov 1969, p. 543) and have concluded that rifting and the volcanism are interrelated phenomena. Other workers (e.g. Florensov et al. 1968) feel that this association has been rather overemphasized in the literature, and that the Cenozoic volcanism associated with the East African, Rhine and Baikal rifts, is part of widespread phenomenon, and should strictly speaking, be regarded as independent of rift formation. These workers point out that although the East African Rift system is often cited as an example of association between volcanism and rift structures, volcanism also appeared at about the same time in other parts of Africa, such as between Guinea Bay and the Chad Lake depression in Nigeria, in the Comore Islands and in some other areas. However, they admit that when certain quantitative aspects (volume of volcanic products, thickness and number of lava flows, number and size of volcanic forms) of Cenozoic volcanism accompanying the large rift zones are considered, a correlation exists between the scale of volcanic phenomena and the scale of rift structures. For example, the three classical rift systems, arranged in decreasing order of size and accompanying volcanic products are (i) the East African Rift system (ii) the Baikal Rift system (iii) the Rhine graben. This type of correlation does indicate some interrelationship between the size of rift structures and the magnitude of the accompanying volcanism. As has been suggested by several workers (Shackleton 1954;Bailey 1964; Harris 1969; King 1970; Gass 1970), it is quite likely that doming or regional uplift, accompanied by deep faulting and magmatism, are the manifestations of a more fundamental process that originates in the upper mantle. Deep faulting may or may not combine to form large rift systems. Accordingly, the development of large scale volcanism independent of rifts seems reasonable, and yet the association in space and time of volcanism with rift zones does not appear fortuitous.

One rather outstanding characteristic of igneous activity associated with the East African Rifts is the unusually large scale development of alkaline plutonic and volcanic rocks often with foid-bearing differentiates. Petrochemically similar rocks are also associated with the Rhine graben. The most common plutonic bodies of the alkaline series are central ring complexes of ultrabasic rocks and carbonatite. Carbonatite in general (Heinrich 1966) and Kimberlite in particular (Dawson 1970) are rock types restricted to the continental environment. They are intimately associated with deep fault zones related to crustal upwarps. The alkaline-carbonatite areas of the East African and Rhine Rift zones are commonly characterized by diatreme pipes (e.g. Swabian tuffisite pipes; Cloos 1941) and spectacular explosion craters (e.g. ring craters around Ruwenzori; Holmes 1964, p. 1073) indicating the creation of fluidized systems (Reynolds 1954) by high pressure gas movement and sudden release of constricted gas. There is abundant evidence that the gas responsible is CO_2 (Heinrich 1966, p. 296). Compared with igneous rocks of the East African and Rhine rifts, those of the Baikal Rift are less alkaline. They consist dominantly of normal olivine basalts, andesite-basalts and some moderately alkaline types (Florensov et al. 1968). Greatly undersaturated felspathoid-bearing rocks which occur not infrequently in the East African and Rhine lava series, are unknown in the Baikal region and carbonatite and ultrabasic alkaline rocks seem to be altogether absent (Florensov et al. 1968).

Olivine basalts, both normal and mildly alkaline, also form part of the volcanic products of the East African and Rhine Rifts. Furthermore, large volumes of tholeiitic basalts also occur in the East African system (Harris 1969; Gass 1970) but seem to be restricted to the Red Sea area and the Afar depression. Harris (1969) points out that the basalt type changes along the East African Rift system, becoming more alkaline away from the Red Sea area. He also suggests that in areas of intense volcanism, as for example in the Afar triple junction area, the basalt type tends to be less alkaline. Harris (1969) attempts to explain the main differences in the basalt types along the rift system, in terms of differences in the geothermal state of those parts of the upper mantle from which the magmas were derived.

The spatial distribution of igneous activity in relation to the rifts shows, among others, two interesting features. Firstly, even in East Africa where a genetic link between rifts and igneous activity is most manifest, considerable sections of the rift zones are not associated with igneous rocks that can be related to rift formation, and in the Baikal, volcanic rocks occur only in two of the ten large grabens (Florensov et al. 1968). Secondly, although it might seem that the boundary rift faults should control much of the igneous activity, this is actually not the case. It is true that in the East African rift zones, some igneous centres are arranged along longitudinal rift faults, but a majority of them seem to be controlled by lateral fractures, some of which extend for long distances from the rift zones. Thus, many of the igneous centres occur outside the main rifts. This, in fact, is the most common habitat of carbonatite complexes. Within the rift zone, notable concentrations of igneous activity occur at their intersections and bifurcations (Bailey 1964) and also where grabens terminate (e.g. in Bufumbira and eastern Galilee; Freund 1966). The Rhine graben, too, is associated with lateral volcanic lines (Bederke 1966) and also has notable concentrations of igneous activity at graben terminations. The above features show that areas most conducive to magmatism formed outside the main grabens, on the flanks of related upwarps, where the crustal plates have cracked, to provide open fissures. Within the grabens igneous activity has occurred mostly where structures bifurcate, terminate (usually by splaying out) or intersect with other graben structures. Perhaps the enormous weight of the subsiding rift blocks, keeps the boundary faults tightly closed most of the time in most places (Holmes 1964, p. 1951).

Time relations between igneous activity and evolution of continental rifts are complex. In the East African Rifts, for instance, the most intense phases of igneous activity were not synchronous in different parts of the rift system. For example, there is no evidence of Tertiary volcanism along the Western Rift. However, intense volcanic activity began there in the Pleistocene (Beloussov 1969). Also, there are practically no signs of Cenozoic volcanism in the southern parts of the rift system. There, the youngest activity is represented by the rocks of the Chilwa alkaline province of Late Jurassic and Early Cretaceous age.

In attempting to compare the scale and nature of igneous activity in the St. Lawrence Region with those of the three classical rift zones, a point to be reckoned with is that whereas the three latter rift zones are characterized by Cenozoic volcanism, no igneous activity of this age is known in the St. Lawrence Region. Older volcanics, of course, have much less chance of survival.

Igneous Products Associated with the St. Lawrence Rift System

(a) The Grenville dike swarm

This diabase (tholeiitic) dike swarm is about 650 km long and has been discussed earlier (p.111). Its close association with the Ottawa and Nipissing grabens is seen from Figure 54. Despite some deviations, especially in the Lower Ottawa Valley area, the dike swarm and graben structures show a remarkable unity of trends and patterns. In the western part of the Nipissing graben the dike pattern fans possibly simulating the splay pattern of the structure. In the western part of the Ottawa graben there are indications that some of the dikes change direction to conform with the Timiskaming graben trend. The precise ages of these dikes are uncertain. The dikes are definitely post-Grenville and they are not known to cut the Ordovician rocks of the Ottawa Valley area. They are regarded as Hadrynian (See Fahrig 1972, p. 576). The presence of diabase dike swarms in the crust is interpreted by most authors (e.g. Fahrig 1970, p. 134) as indicating a tensional stress environment at the time of their emplacement. In the Ottawa Valley - Nipissing Depression areas, the near-parallelism and close space association of the dike swarm with graben structures indicate that diking and high-angle (mostly normal) faulting occurred under the same stress regime. Whether some of the dikes were feeders for flood-basalt eruptions is an interesting speculation.

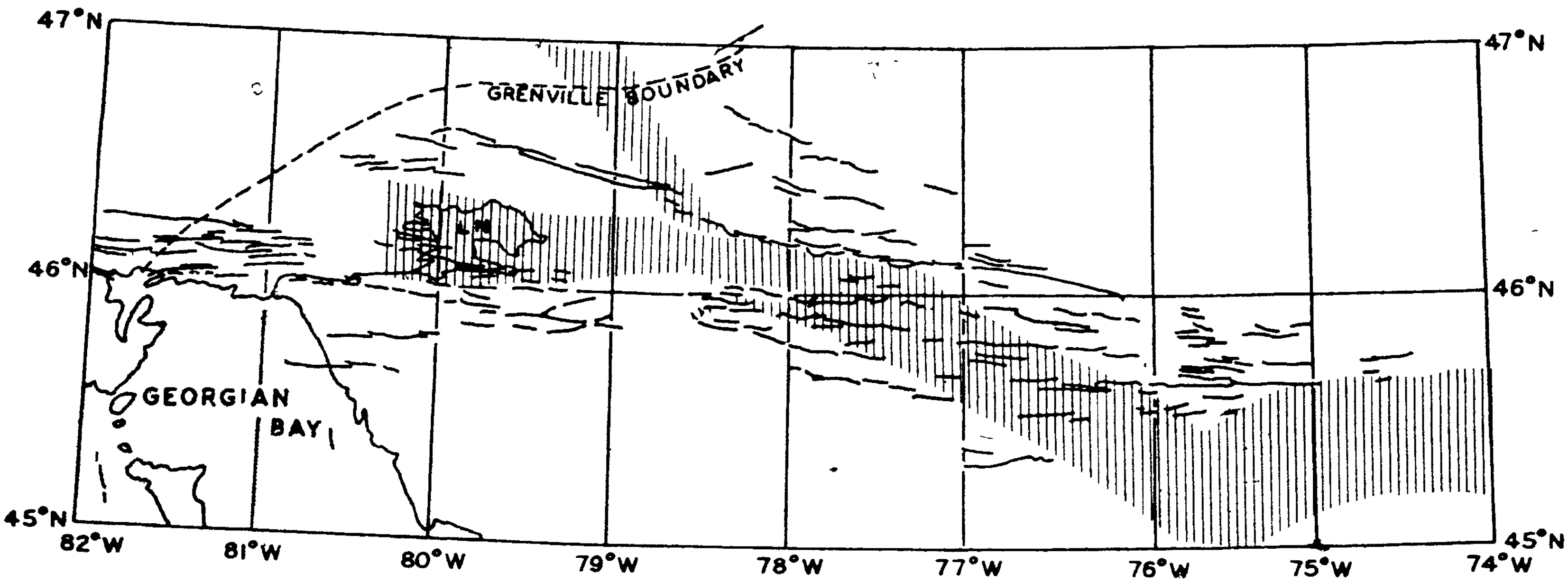


Fig.54. Map showing the Grenville dike swarm (dikes shown as thick lines, after Murthy 1971) and its relation to Ottawa and Nipissing grabens.

(b) Frontenac Axis dikes

In the Frontenac axis there are two sets of post-Grenville diabase dikes that are parallel to faults of the St. Lawrence Rift system. They are:

1. an east-northeast set parallel to the longitudinal faults of the St. Lawrence Valley
2. a northwest set parallel to the faults of the Madawaska Highlands (also of the Upper Ottawa Valley).

Dikes from the Frontenac axis area have yielded K/Ar ages between 400 to 450 m.y. (Park and Irving 1972).

(c) Flood Basalts and Diabase Dike Swarms in the Belle Isle area and in Northern Long Range Mountains

The Hadrynian diabase dike swarm in the northern Long Range Mountains (Pringle et al. 1971) and the presumably consanguineous dikes and flood basalts of the Belle Isle area (Williams and Stevens 1969) have already been discussed (see p.90). The flood basalts are transitional between tholeiitic and alkaline basalts (Strong and Williams 1971) and appear to be similar in petrochemistry to the early Cambrian or late Hadrynian Tibbit Hill volcanics of the Pinnacle formation in northwestern Vermont and adjacent Quebec (see Cady 1969, p. 148). Both volcanic groups occur low in the stratigraphic sequences in the respective areas and their spatial relationships to the St. Lawrence Rift system on the one hand and to the Appalachian foldbelt on the other are similar except that the Tibbit Hill volcanics are a little more in the geosyncline and hence are deformed. It is possibly significant that the Tibbit Hill volcanics occur just to the east of intersection of the Shield Segment A with the Marginal Segment Rift Zone. It will be proposed later, that the ancestral structure of the St. Lawrence Rift system represents a part of a larger continental rift system that developed as a prelude to the Appalachian evolution and that the Tibbit Hill volcanics, Belle Isle volcanics and the associated dike swarms, and the Grenville dike swarm are mutually related in time and cause with the origin and development of the ancestral structure.

(d) Alkaline and Carbonatite Complexes and Related Minor Intrusions

Twenty central complexes and numerous dikes, sills and plugs of alkaline ~~and~~ carbonatite rocks ranging in age from early Cambrian to Early Cretaceous are known to occur in close association with the St. Lawrence Rift system (Fig. 55). These have been described in earlier Chapters. Intrusions in the Nipissing Depression area and in the southern part of the St. Lawrence Valley occur in two well defined petrographic provinces; the Lake Nipissing and Monteregian

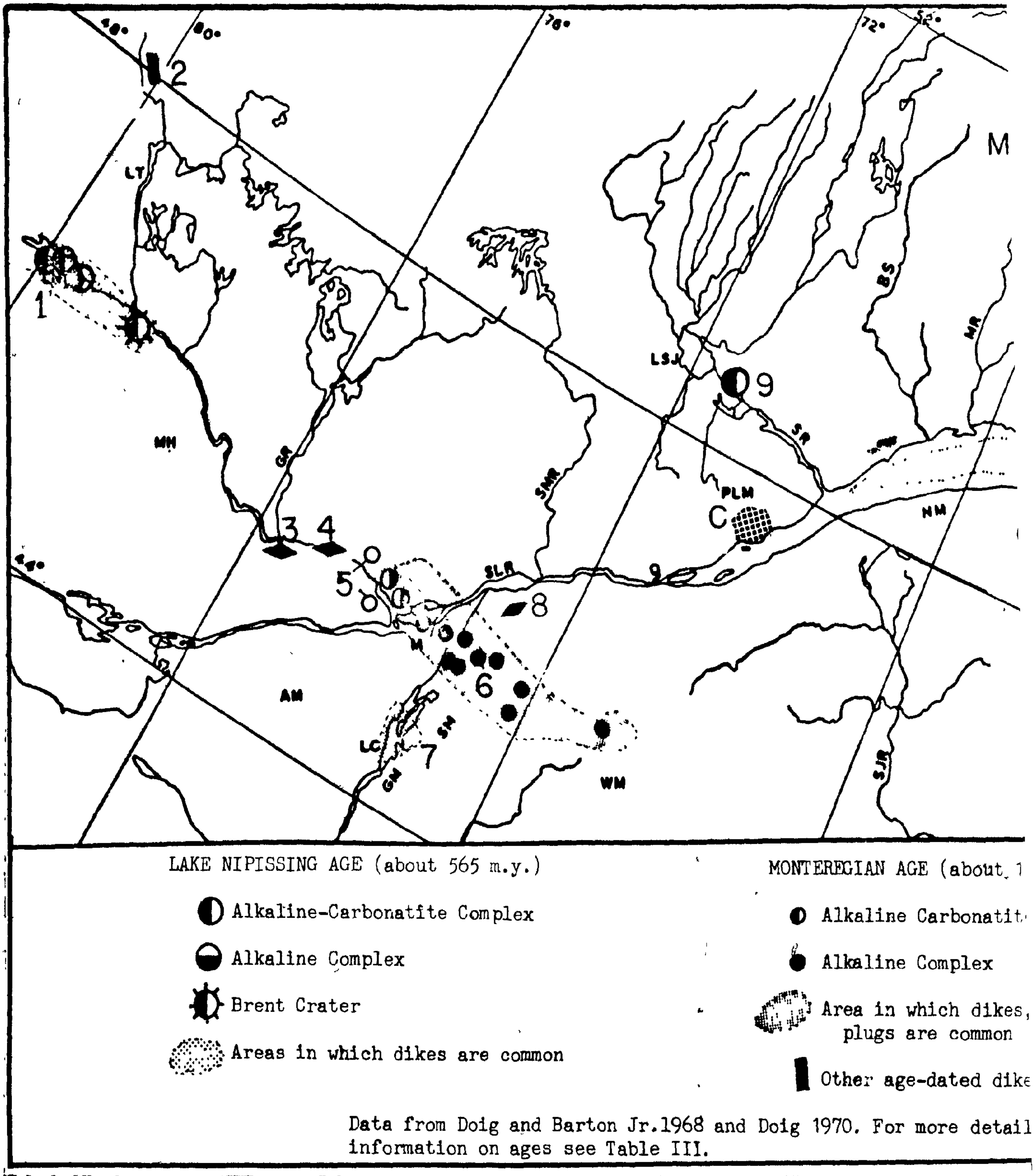
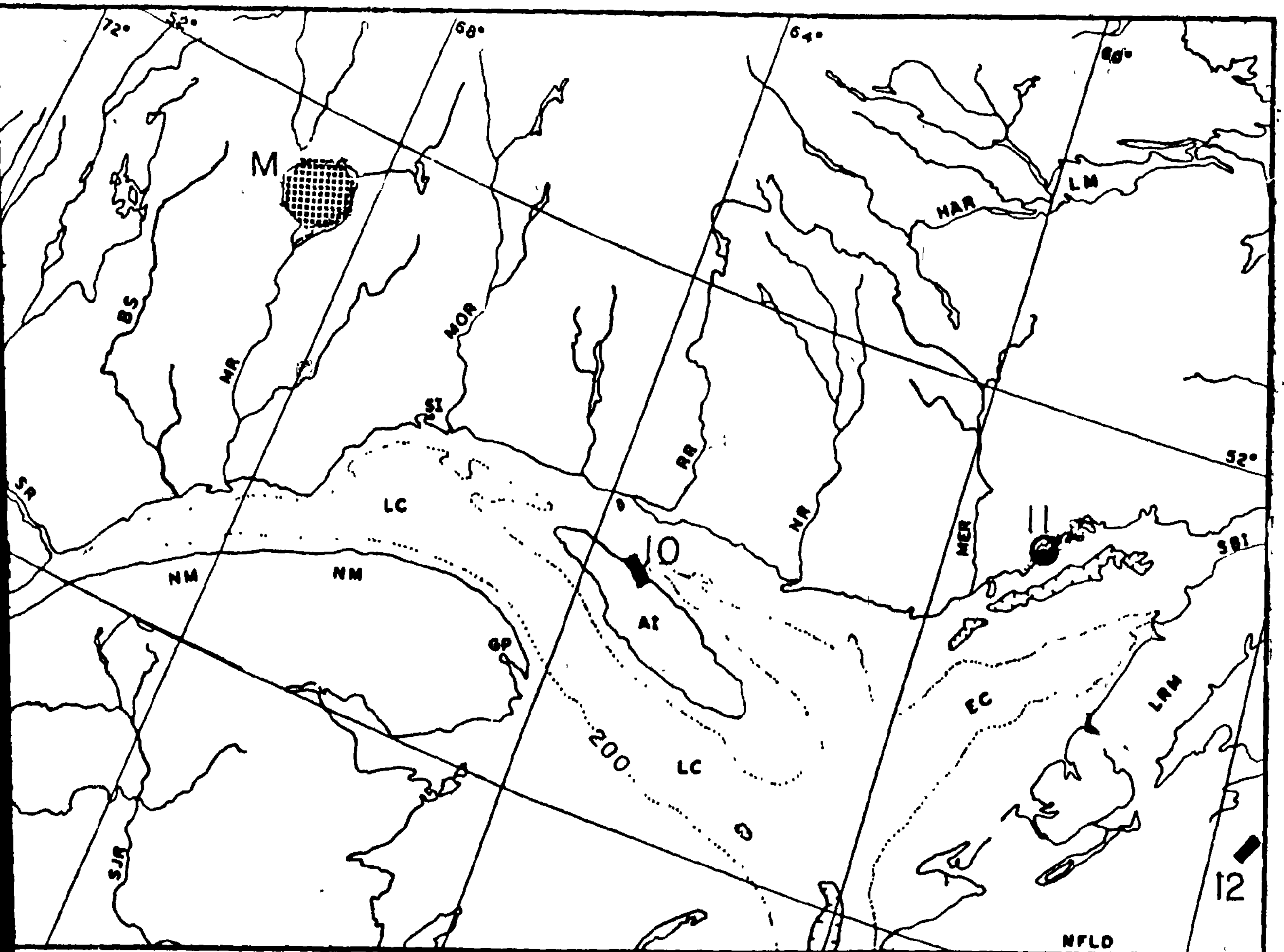


Fig.55. Distribution of post-Grenville alkaline rocks, carbonatites and explosion craters in the Region.

1 of



TEREGIAN AGE (about 110 m.y.)

Alkaline Carbonatite Complex

Alkaline Complex

Area in which dikes, plugs and small plugs are common

Other age-dated dikes

1970. For more detailed

POST-LAKE NIPISSING AND PRE-MONTEREGIAN AGE

Alkaline Complex

Explosion Crater { M-Manicouagan, C-Charlevoix

Dated Dike

KM 100 0 100

200 Bathymetric contours in metres

losion craters in the St.Lawrence

petrographic provinces. In addition to these igneous rocks of Phanerozoic age, there are other alkaline and carbonatite intrusions of Precambrian age in the St. Lawrence Region. These include the Meach Lake carbonatite bodies (Hogarth 1966) just north of Ottawa with K/Ar ages of about 920 m.y.; alkaline-syenite plutons in the Mount Laurier area (Wynne-Edwards et al. 1966, p. 25) about 100 km north of the Lower Ottawa Valley, with K/Ar ages ranging from 822 to 1005 m.y. (Doig and Barton Jr. 1968); alkaline and carbonatite intrusions in the Haliburton-Bancroft area (Chayes 1942) about 100 km southwest of the Upper Ottawa Valley, with K/Ar ages ranging from 900 to 1000 m.y. (MacIntyre et al. 1967). These intrusions might be connected with events that lead to the initiation of the St. Lawrence Rift system (see Sutton 1969, 1970), although this cannot be established with any certainity. But the younger intrusions mentioned earlier are probably related in time and cause with rifting along the St. Lawrence Valley system.

The alkaline rocks of the St. Lawrence Region closely resemble the sub-volcanic assemblages of the East African and Rhine Rift zones. Rocks of all three areas are strongly alkaline. Distinctive rocks such as carbonatite and kimberlite, characteristic of the East African and Rhine alkaline provinces, have their counterparts in the St. Lawrence

Region. For instance, of the twenty central complexes known from the St. Lawrence Region, six contain carbonatite. Kimberlite, is relatively rare: the only known occurrences are dikes at the north end of the Timiskaming graben, and a pipe at Isle Bizard in the western part of the Monteregian province. Alnöite which is closely akin to kimberlite, however, is present abundantly as dikes in the western part of the Monteregian province.

The space relations between alkaline centres and graben structures of the St. Lawrence, East African and Rhine rift zones are also similar in some respects. In the St. Lawrence Region only two of the known alkaline centres (Baie-des-Moutons complex, Chatham-Grenville complex) are located along a boundary fault of the rifts. Of the intrusive centres that occur within the rift zones, the carbonatite complexes in the Nipissing graben and the single known complex in the Saguenay graben occur where the respective graben structures begin splaying out. The Rigaud and Chatham-Grenville stocks and the Monteregian intrusions are largely localized at a major bifurcation of the St. Lawrence Rift system. The Monteregian line of intrusions extends across the Marginal Segment Rift Zone and continues beyond the eastern limit of the rift zone for a considerable distance. Thus, the disposition of the Monteregian line with respect to the rift zone is similar to that of the lateral volcanic lines in the East African and Rhine settings.

However, this line of intrusions may simply be set along the eastward continuation of the Ottawa graben faults which, as discussed later (see p. 256) appear to extend across the Marginal Segment Rift Zone into the Appalachian foldbet.

The alkaline igneous rocks and carbonatites in the St. Lawrence Region show a wide scatter of ages (Doig 1970), a feature which is also characteristic of the East African situation (King 1970). The wide ranges in ages of igneous activity probably indicate similarly wide ranges in ages of tectonic activity. The K/Ar ages of alkaline rocks and carbonatites listed in Table III range from 580 m.y. to 90 m.y. Quantitatively (considering the number of central complexes, dikes sills, plugs, breccia pipes and extent of hydrothermal alteration) however, the bulk of the magmatic products have appeared in two distinct periods: one about the beginning of the Cambrian Period including rocks of the Lake Nipissing alkaline province, St. Honoré carbonatite, Baie des Moutons complex and dikes; the other in late Jurassic - early Cretaceous periods including rocks of the Monteregian province and the kimberlite dikes along the Timiskaming graben. Other igneous products that are included in the latter group are the northwest trending lamprophyre dike on Anticosti Island, with a K/Ar whole rock age of 138 m.y. (Poole et al. 1970, p. 298) and the northeast trending lamprophyre dikes in Notre Dame Bay area, Newfoundland (Fig. 55) with K/Ar ages (on biotite and hornblende) ranging from 115 to 144 m.y. (Wanless et al. 1967, p. 114), although the Notre Dame Bay dikes are

TABLE III

K/Ar ages of post-Grenville alkaline rocks and carbonatites associated with the St. Lawrence Rift system

Area Code in Fig. 55	Igneous Bodies	Detailed Locality reference(s) Fig.	Age, 10^6 yrs	d	n
6	Monteregian rocks	10	90-130	n.a.	n.a.
7	Lake Champlain dikes	13	120-150	n.a.	n.a.
12	Notre Dame Bay Lamprophyre dikes	--	115-144	n.a.	
10	Anticosti Island lamprophyre dikes	--	138	n.a.	
2	Kirkland Lake kimberlite dike	29	151	n.a.	n.a.
4	Buckingham dikes of mica peridotite	21	275		1
3	East view carbonatite dike	21	320	n.a.	n.a.
8	Bon conseil syenite and mica-pyroxenite body	9	428	20	3
5	Chatham-Grenville alkaline syenite stock	21	450	27	6
5	Rigaud alkaline syenite stock	21	450	22	7
-	Ste. Sophie alkaline gabbro dikes	9	520	3	2
9	St. Honoré carbonatite complex	35	564	4	3
1	Alkaline dikes near Brent crater	23	558-576	n.a.	n.a.
1	Manitou Island carbonate complex	27	565	4	2
11	Baie des Moutons alkaline syenite stock	16	568	8	3

Data from Doig 1970 and Doig and Barton Jr. 1968 and other sources. d is the sample standard deviation for the number of age determinations n used to compute the mean age. n.a. - data not available.

located at a considerable distance from the rift system. These two phases of magmatism are widespread in the St. Lawrence Region and may reflect tectonic movements of a general nature along the rift system. The other alkaline intrusions have appeared at various intermediate ages in different localities along the rift system and probably are related to movements of a more localized nature.

(e) Explosion Craters

Explosion craters and diatreme breccia pipes characteristic of the alkaline-carbonatite areas of the East African and Rhine grabens have already been discussed. In the alkaline-carbonatite areas of the St. Lawrence Region, also, high pressure gas movement creating fluidized systems are indicated by the presence of numerous diatreme breccia pipes, especially in the western part of the Monteregian province. Also closely associated in space with the St. Lawrence Rifts are three explosion craters: Brent, Charlevoix and Manicouagan (Fig. 55). The Brent crater is on the south boundary fault zone of the Ottawa graben and is in the area of the Lake Nipissing alkaline province. The Charlevoix crater is on a north boundary fault of the Marginal Segment Rift Zone. The Manicouagan crater is located on what seems to be the splay-out of a cross-fracture in which the Manicouagan River is entrenched. All three have igneous rocks associated with them. Brent has one dike-like mass of olivine-normative alkaline trachyte, and several dikes of lamprophyre (ocellar monchiquite) which are similar petrochemically and in age (Currie 1971a) to the Lake Nipissing alkaline rocks. The Charlevoix crater has hydrothermally-altered breccia some of which is similar to the suevite of Ries, Germany, and pseudotachylyte dikes and in the glacial deposits in the area of the crater are blocks of fused rocks ("impacticite" of Rondot 1971). Small calcite-fluorite veins with galena and sphalerite presumably related to the igneous rocks are also present nearby (Rondot 1968). K/Ar age determinations of the igneous material have yielded ages in the range of 320 to 370 m.y. (Rondot 1971, p. 5421). Currie (1972, p. 140) states that at Manicouagan the pile of igneous material includes, from bottom to top, tuff breccia (suevite), alkaline basalt with ultrabasic inclusions, recrystallised aphanitic doreite, fine-grained doreite and coarse-grained doreite about 200 m thick. A K/Ar age of 210 ± 4 m.y. has been obtained from the igneous material (Wolfe 1971). Shock metamorphic features have been found in all three areas; the main ones being the presence of maskelynite and shock lamellae in quartz indicating high strain-rates and melting of quartz to lechatelierite indicating high temperature effects (Short and Bunch 1968).

The Brent and Charlevoix craters are located on major faults of the St. Lawrence Rift system. The Manicouagan crater also appears to be associated with a lateral fracture related to the rift system. The ages of igneous rocks of the craters are within the span of time (early Cambrian to Early Cretaceous) of alkaline magmatism associated with the rift system. Also, the igneous rocks from the Brent crater resemble the alkaline rocks and carbonatites of the Lake Nipissing alkaline province with regard to age and petrochemistry. The above relationships suggest that the craters are manifestations of volcanism related to the St. Lawrence Rift system. However, the shock metamorphic features associated with the craters have been interpreted by some in favour of a meteorite impact origin for them (Brent: Dence 1965, 1968; Dence et al. 1968. Charlevoix: Robertson 1968, Rondot 1969, 1970, 1971. Mancouagan: Beals et al. 1963, Dence 1965). The premise on which this interpretation is based is that such features as shock lamellae in quartz and the formation of maskelynite, require extremely high shock pressures of several tens or even hundreds of kilobars (see Hörz 1968) and that such pressures are significantly outside the domain of normal geological processes (see French 1968). This premise, however, is not accepted by all geologists (for example see Currie 1968, 1971b). In this connection it may be pointed out that the energy expended during the 1883 volcanic explosion of Krakatao, Indonesia has been estimated at not less than that could be liberated by 5,000 megaton hydrogen bombs (Holmes 1964, p. 339). The above energy figure is much larger than the yield of underground nuclear explosions that have produced shock metamorphic effects comparable with natural examples. Also, shock metamorphic features are present in some breccias and adjacent rocks in the vicinity of some porphyry copper deposits (Godwin 1973) and these must surely have formed by some endogenetic process. Thus, in the present state of knowledge, it appears that both impact and endogenetic processes produce shock metamorphic features in rocks. Accordingly, the endogenetic hypothesis of Bucher (1963) and others (McCall 1964; Snyder and Gerdemann 1965), based largely on the fact that a considerable number of known explosion craters are systematically related to earth structures and areas of well-known igneous activity, should be given first preference in explaining craters that show such relations. Thus, an impact hypothesis for the three discussed explosion craters seems improbable because they are so patently related to the St. Lawrence Rifts.

GEOPHYSICAL CHARACTERISTICS

General Statement

To a large extent, geophysical characteristics of continental rifts are similar. They are usually characterized by negative Bouguer gravity anomalies, above average heat flow, and shallow seismicity. In the following sections the geophysical characteristics of the St. Lawrence

Rift system are compared with those of the three classical rift zones.

Gravity Anomalies

Bullard's (1936) gravity measurements in East Africa demonstrated, for the first time, the presence of negative Bouguer anomalies over the rift valleys of that region. Later surveys have revealed similar anomalies over other rift valleys and it is now believed that these anomalies are a characteristic feature of large continental rifts (Girdler 1964). The maximum amplitude of the anomalies is about 50 milligals.

There is little agreement among students of graben problems as to the cause of these negative anomalies. Most workers attribute the anomalies to the effect of thick accumulations of relatively low density unconsolidated sediments that commonly occur in the fault-troughs (e.g. see Zorin 1966a). Others think that the gravity lows cannot be fully explained this way (e.g. Mueller 1970, p. 31) and seek additional causes of deep-seated origin, but there is no agreement amongst them as to the nature of these causes. Thus, various models with differing crust-upper mantle parameters (below rift zones) have been proposed: e.g. thicker-than-average crust (Bulmasov 1960); thinner-than-average crust with an "antiroot" (Freund 1966); an upper mantle with a crust-mantle mix (Florensov 1969).

The Bouguer anomaly map of the St. Lawrence Region (Bouguer gravity anomaly map of Canada 1969) shows that, unlike the rift zones of the three classical areas, the St. Lawrence Rifts are not associated with conspicuous Bouguer gravity lows. The only part of the St. Lawrence rifts with discernible gravity lows ($\sim$ 20 milligals) is the Marginal Segment Rift Zone along the Champlain and St. Lawrence Valleys. These gravity lows probably reflect downbowed platformal strata (Cambro-Ordovician) whose basal formation (Potsdam) is a low density (2.5) sandstone. The Oslo graben is another rift, which according to the Norwegian gravity map shows no gravity low (Thompson in Bederke 1966). It is interesting to note that both the Oslo graben and St. Lawrence Valley system have no noteworthy amounts of low density sedimentary fillings. Thus, it seems reasonable to conclude that the absence of conspicuous gravity lows over the St. Lawrence rifts is largely due to the lack of thick sedimentary fills in the fault troughs. However, because of the long inactivity of the St. Lawrence Rift system, if any deep seated causes contributory to a negative gravity field did exist originally, such causes could have largely disappeared with time.

Heat Flow

Of the three classical rifts, heat flow measurements are most advanced in the Baikal. Measurements in the other two areas are scanty but the indications are that the heat flow characteristics of all three rift zones are similar. In the platform region on the north-east side of Lake Baikal, the heat flow values, on the average, are about 1.0 HFU and increase eastwards to about 1.6 HFU on the west coast of the lake and to 2.6 HFU on the axis of the rift along the lake (Lubimova 1969).

From the analysis of heat flow data by Horai and Simmons (1969), it appears that the average heat flow for the Appalachian Region (1.32 HFU from 32 values with a 0.38 standard deviation of mean) is somewhat higher than for the Canadian Shield (1.09 from 21 values with a 0.46 standard deviation of mean). Within the St. Lawrence rifts, a few heat flow values have been determined in the St. Lawrence Valley (Saull et al. 1962; and Crain 1967) and in the Ottawa Valley (Jessop and Judge 1971). The indications are that the heat flow in the St. Lawrence rift zones is not different from that of the Shield Region (also see Diment et al. 1972). Thus, it appears that unlike in the Baikal and also in the other rift zones (see Hanel 1970; Von Herzen and Vacquier 1967) there is no abnormal heat flux through the rift zones of the St. Lawrence. This difference can also be explained as a result of the long inactivity of the St. Lawrence Rift system.

Seismicity of Continental Grabens

The main continental seismic zones of the world include the Baikal and the East African Rift Zones (Miyamura 1969). The records of the Baikal region are more complete as regards geographical coverage and go further back in time than those of East Africa. The Baikal records show that the level of seismicity there (see Solonenko 1968b) whether reckoned from the annually recorded number of earthquakes or by their intensity, is higher than that of East Africa (see Wohlenberg 1970). The Rhine rift zone shows only mild seismicity (Hägle and Wohlenberg, 1970).

The earthquakes of the continental rift zones are shallow focus; the depths of foci are commonly less than 40 km in East Africa (Wohlenberg 1970) and Baikal (Treskov 1968), and less than 25 km in the Rhine (Ahorner 1970).

The characteristics of the stress state in the Baikal and East African seismic zones have been interpreted from studies of focal mechanisms of earthquakes by Bakalina and others (1969). These interpretations, however, are based on certain assumptions (see Garland 1971, pp. 87-91) and are not unequivocal. Balakina and coworkers (1967) state that the Baikal region is characterized by non-uniform horizontal "expansion", mainly normal to the rift zones, and the East African rift zones are characterized by horizontal compression parallel and horizontal tension normal to the graben trends. These inferred stresses are compatible with the generally accepted view of graben formation by tensional stresses. In the Rhine region, however, the contemporary stress state appears to be different from the stress regime under which crustal updoming and graben formation took place. The focal mechanism studies indicate horizontal compressional stresses of regional extent in the area, the greatest principal stress direction being northwest-southeast (Ahorner 1970, pp. 164-166).

Examination of epicentre maps (East Africa: Fig. 1 of Wohlenberg 1970. Rhine: Fig. 1 of Ahorner 1970. Baikal: Treskov 1968) of the classical rift zones show certain space relationships between the geographic distribution of epicentres and rift structures.

1. The gross patterns of rift structures and epicentres correspond fairly well especially when plotted on small scale maps.
2. The epicentre patterns spread over wider areas than the rift zones themselves.
3. Some parts of the seismic zones have unusually great accumulations of epicentres.
4. Some parts of the rift zones are aseismic.
5. Some of the associated seismic zones cannot be correlated with any known geological structures.

Seismicity of the St. Lawrence Region

The St. Lawrence Region is seismically active. Five shocks of magnitudes ranging from 5.9 to 7.2 have occurred in this region since the turn of the century, and on the average about 20 minor shocks (M = 2-5) per year were recorded during the period 1954 to 1959 (Smith 1964). In Table IV are listed shocks of M $\geq$ 5 that are known to have occurred in the region.

The earthquakes in the region are described in a series of catalogues (Smith 1962; Milne and Smith 1963, 1964; Smith 1966a; Stevens et al. 1972, 1973) and are discussed in several papers (Smith 1966b); Hamilton 1966; Smith 1967; Milne 1967; Milne et al. 1970). Epicentre maps for the periods 1534 to 1927 and 1928 to 1959 have been

TABLE IV

Larger earthquakes in the St. Lawrence Region. Data from Hodgson 1965 and other sources

Date	Approximate Location of Epicentre	Estimated Magnitude
1638	St. Lawrence River near the mouth of the Saguenay	7
1663	Near the mouth of the Saguenay	7.5 - 8
1665	Near the mouth of the Saguenay	6.4
1732	At Montreal	7
1791	St. Lawrence River near the mouth of the Saguenay	6.4
1816	Near Montreal	5.5 - 5
1831	Two earthquakes of about equal intensity at the mouth of the Saguenay	5.5 - 6
1860	Mouth of the Saguenay	6.5 - 7
1861	At Ottawa	5.5 - 6
1870	St. Lawrence River near the mouth of the Saguenay	7
1897	Near Montreal (two earthquakes)	5 - 6
1924	In the Upper Ottawa Valley	6.1
1925	St. Lawrence Valley near the mouth of the Saguenay	7
1929	Grand Banks, Newfoundland	7.2
1935	Timiskaming Depression (at Timiskaming)	6.2
1944	At Cornwall, Ontario	5.9
1967	Lower St. Lawrence River southeast of Sept Iles	5.3

published (Smith 1962, p. 1966). The former map contains mostly non-instrumental data and is bound to be biased by population distribution; the latter which contains only instrumental data is reproduced in Fig. 56. This map is probably satisfactory for inferring broad seismo-tectonic trends but is unsatisfacotry for detailed seismic-structural correlations because of uncertainties (± 20' of arc) in epicentral locations. Such uncertainties act to diffuse trends. Also included in Figure 56 are the rift zones of the St. Lawrence system. An examination of the Figure shows the following seismo-tectonic patterns.

1. When the number of epicentres and/or the size of shocks are considered, the greater part by far of the seismicity of the general region is closely associated with the St. Lawrence Rift zones.
2. There are two areas of marked epicentre concentrations: one extending from Quebec City downstream to about Anticosti Island and more or less straddling the Marginal Segment Rift Zone and the other extending across the south end of the St. Lawrence Valley. These two zones will be referred to as Zones A and B respectively.
3. There is a minor, rather diffuse zone of epicentres along the southeast part of the Appalachians. It will be referred to as Zone C.

4. A closely grouped cluster of half a dozen or so epicentres occurs at the mouth of the Laurentian Channel.
5. Two segments of the rift system are nearly aseismic. One is between Zones A and B and the other is east of Zone A (except for the cluster of epicentres at the mouth of the Laurentian Channel).

Zone A follows the Marginal Segment Rift Zone for a distance of about 500 km. The part between Quebec City and the Saguenay outlet is characterized by a dense cluster of epicentres. This cluster of epicentres straddles the north boundary fault zone of the Marginal Segment Rift Zone and is centered around the highly faulted area associated with the Charlevoix crater. It is also in an area of neotectonic movements (Frost and Lilly 1966; Vaníček and Hamilton 1972; Nyland 1973) and is characterized by intense microearthquake activity (Leblanc et al. 1973).

Zone B extends from the northern part of the Champlain Valley, northwestwards across the southern part of the St. Lawrence Valley and thence for another 300 km or so. In this latter part, the seismic zone fans out to include the Shield Segment A (excluding the Nipissing graben) and a large area of the Shield Region to the northeast. As seen from Figure 56, there is a suggestion that the Zone B extends south-eastwards to include the cluster of epicentres (grouped with Zone C) just north of Boston. Recently, several workers have commented on the possible

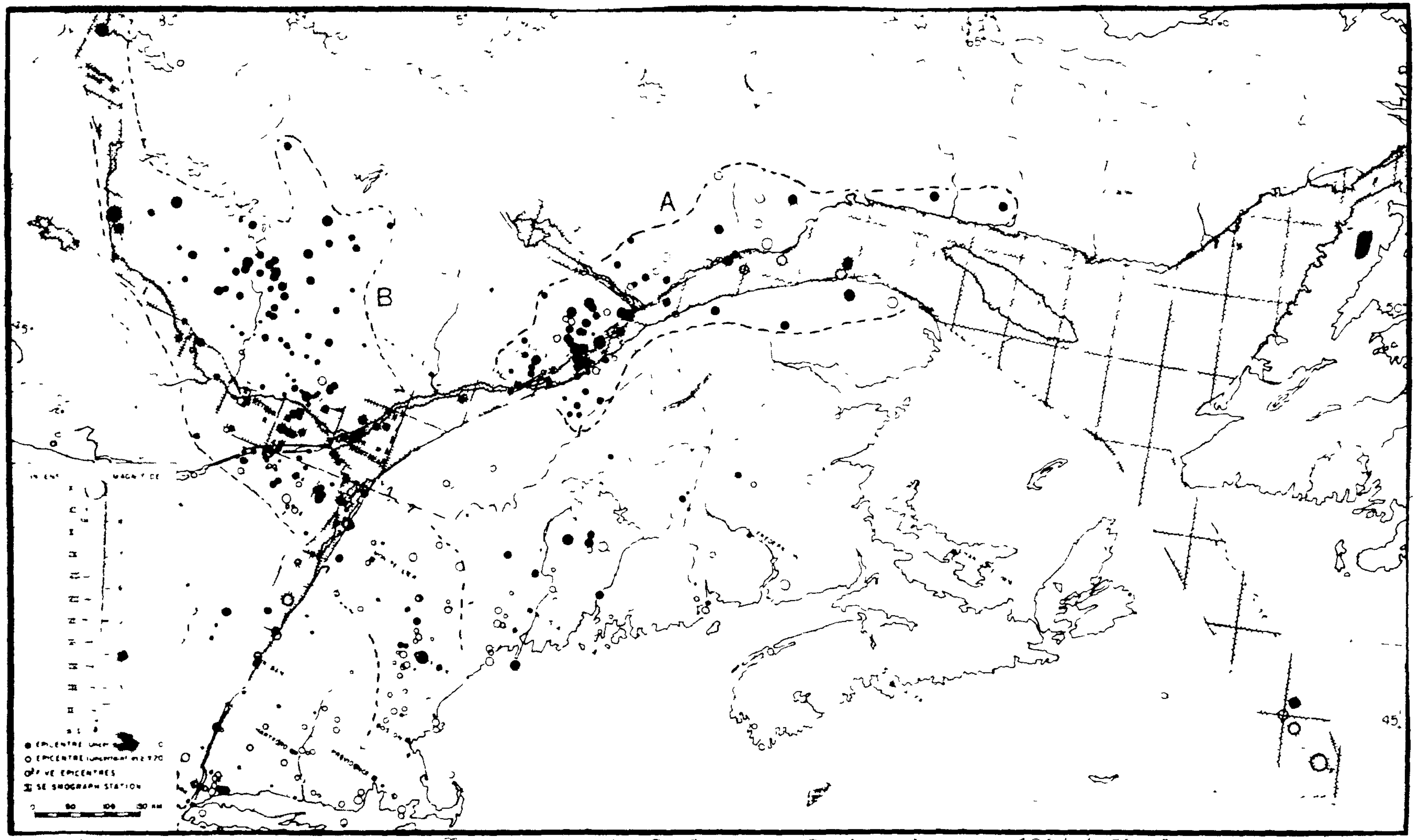


Fig.56. Map showing earthquake epicentres of the St.Lawrence Region, between 1928-1959. Epicentre uncertainities: solid circles< ± 20', open circles > ± 20'. Shaded areas are the rift zones of the St.Lawrence system. Dashed lines inserted to show possible boundaries of seismic zones A and B. Earthquake data after Smith 1966.

significance of this so-called Ottawa-Boston seismic zone (e.g. Leblanc et al. 1973; Sbar and Sykes 1973). Five possibly interrelated lineaments lie along the Ottawa-Boston seismic zone.

1. The Ottawa graben (also part of the Timiskaming graben) and the east-west trending faults in the southern part of the St. Lawrence Valley. These latter faults combine to form a graben-like crustal sag (see Clark 1972, Fig. 18) which may actually be the eastward continuation of the Ottawa graben across the Marginal Segment Rift Zone. The area over which the epicentre pattern fans out (Fig. 56) is a part of the Shield Region where the prevalently northeast and northwest oriented "regional fracture pattern" is well developed (e.g. see Fig. 6). This fracture pattern could conceivably control the epicentre pattern, but nothing definitive can be said because of the uncertainities of epicentre locations.
2. The Beauharnois axis (see Fig. 9).
3. The Monteregian line of intrusions excluding Mt. Megantic.
4. The line of intrusions of the White Mountain Magma Series (see Chapman 1968, p. 389).
5. The axis of the New England Salient (of the Appalachian foldbelt) which is coincident with a transverse geosynclinal trough that first appeared late in the Precambrian and persisted

as a transverse trough throughout the evolution of the Appalachian foldbelt (Cady 1969, p. 35).

The Ottawa graben, the graben-like crustal sag across the south end of the St. Lawrence Valley and the transverse geosynclinal trough along the axis of the New England Salient, succeed one another along a linear zone (trending on the average S60°E) suggesting that they may be aspects of a single continuous lineament. It is proposed that during the early development of the St. Lawrence Rift system, the ancestral Shield Segment A extended across the southern part of the St. Lawrence Valley and further southeast along the site now occupied by the transverse geosynclinal trough coincident with the axis of the New England Salient. Reactivation of deep faults of the postulated extension of Shield Segment A could have been mutually related to magmatism of the Monteregian Province "proper" and of the White Mountain Magma Series; contemporary stress release on some of the unhealed faults may account for the extension of the seismic zone B towards Boston. Recently, Diment et al. (1972) have suggested a connection between the Ottawa-Boston seismic zone and the Kelvin seamount chain which Le Pichon and Fox (1971) have interpreted as a fossil transform fault - the Kelvin Fracture Zone (see Fig. 58). This transform fault is believed to have formed during the early opening of the Atlantic Ocean in the Jurassic and Cretaceous (Uchupi et al. 1970; Le Pichon and Fox 1971), a time period which is roughly coincident with the timing of the Monteregian and White Mountain Magma Series magmatism. Still more recently Sbar and Sykes (1973, p. 1875) have pointed out that the Kelvin Fracture Zone and the Ottawa-Boston seismic zone lie on the same small circle about the centre of rotation for plate movements during the same period. They suggest that the seismic zone is located along the continental extension of the fracture zone and that the creation of stresses in the area of the seismic zone may be due to a change in the driving force of lithospheric plates at the present time. While the scheme suggested by Sbar and Sykes (1973) could explain the build up of contemporary stresses in the area of the seismic zone, the tectonic feature to which the seismicity appears to be related is the St. Lawrence Rift system. The Kelvin Fracture Zone itself may be a feature inherited from a line of structural weakness created by a southeast extension of the Shield Segment A - New England Salient axis tectonic line. Similarly, the cluster of epicentres at the mouth of the Laurentian Channel may be related to the St. Lawrence Rift system on the one hand and to the Newfoundland Fracture Zone on the other. Thus it appears that excepting the rather weak seismic Zone C (which is located along the Appalachian foldbelt and a part of which may belong to the Ottawa-Boston Seismic Zone), the largest single tectonic feature to which the seismicity of northeastern North America is spatially related is the St. Lawrence Rift system.

Hodgson (1964) and later Woollard (1969) recognized a seismic trend to include Zones A and B of the St. Lawrence Region and extending southwestwards through Lakes Erie and Ontario to the head of Mississippi Embayment. This trend, however, has little relation to known tectonic features except in the area of the St. Lawrence Rift system and at the head of the Mississippi Embayment where the earthquake pattern appears to be spatially related to a zone of normal faulting. Sbar and Sykes (1973) argue that the seismic trend is not, in fact, as continuous as it appears on Woollard's epicentre maps (1969, Fig. 1) which contain both instrumental and non-instrumental data mixed together.

In relation to other seismically active zones of the earth, the St. Lawrence Rift zones may be described as mildly active, an expression which also appropriately describes the seismicity of the Rhine and Oslo grabens. Nevertheless the occasional large shocks ($M \geq 7$) that are characteristic of the St. Lawrence Region, have no known counterparts in the Rhine or Oslo Rift zones (Max < 6). In comparison with the seismic activity of the Baikal and East African Rift zones, the activity of the St. Lawrence Region is distinctly lower. Also, the epicentral pattern of the St. Lawrence Region does not outline the gross pattern of the rift zones to the same degree as in the case of the other three rift zones. Accurate depth determinations of the St. Lawrence Region earthquakes are lacking, but they appear to be shallow focus ($\sim$ 25 km, see Smith 1967), comparable with the depth of focii of the other three rift zones.

Seismicity and Contemporary Stresses of the St. Lawrence Region

In the preliminary paper on the St. Lawrence Rift system, Kumarapeli and Saull (1966a) suggested that the seismic activity of the St. Lawrence Region may be a manifestation of present tectonic activity of the St. Lawrence Rift system. Since the writing of the paper results from <u>in situ</u> stress measurements combined to a limited extent with fault plane solutions, have shown that the stresses throughout a large part of eastern North America including the St. Lawrence Region are compressional, and that the greatest principal stress is large, nearly horizontal and dominantly east to northeast trending (Sbar and Sykes 1973). Voight (1969) was the first to draw attention to this compressive stress field and he proposed that the St. Lawrence Rift system is not an active extensional feature at the present time, a situation not unlike that of the Rhine Region (see Ahorner 1970). He further suggested that the large horizontal compressive stresses (which also appear to be present in Europe) might be related to the mantle flow pattern that drives sea-floor spreading as well as to glacio-isostatic rebound. Regardless of what the causes of the large stresses are their existence seems sufficiently well established to warrant the conclusion that the seismicity is caused by the release of these stresses (Voight 1969) on unhealed faults.

The close spatial association of the seismicity with the St. Lawrence Rift system suggests that unhealed deep faults in northeastern North America are mainly those of the St. Lawrence Rift system.

SUMMARY AND CONCLUSIONS

Table V gives in summary form the main features discussed earlier in detail, of the St. Lawrence Rift system and of the three classical rift zones. The following are the main points arising from the discussion.

1. The St. Lawrence Rift system is over 2200 km long. Its length, although only about one third of the East African Rift system, is comparable with the length of the Baikal Rift system or that of the Mittellmeer-Mjoessen Zone.
2. Widths of the St. Lawrence Rift zones are comparable with the common widths of grabens (35 to 60 km) of the classical rift zones.
3. For convenience, rifting along the St. Lawrence system can be assigned to two time phases (i) post-Ordovician (ii) ancestral (Ordovician and earlier). Most of the post-Ordovician rifting may have taken place in the mid-Mesozoic and the ancestral rifting appears to have begun sometime in the Hadrynian.
4. The classical rift zones have been active systems during the Cenozoic Era. The East African and Baikal rifts

TABLE V. Comparison of some selected features of the rift systems of East Africa, the Baikal, the Rhine and of the St. Lawrence.

FEATURE COMPARED	EAST AFRICAN RIFT SYSTEM	BAIKAL RIFT SYSTEM	RHINE GRABEN	ST. LAWRENCE RIFT SYSTEM
Length in km	6500	2500	300	> 2200
Common widths of rifts	50	50	30	30-60
Tectonic Setting	On crustal swell in Craton	On crustal swell partly in craton and partly in foldbelt (Transbaikalian) but the rift mainly on craton.	On crustal swell in craton	On craton. A crustal swell cannot be discerned in the area.
Gross Trend	N-S conforms with the general trend of the Precambrian basement.	NE-SW conforms with the general trend of the Precambrian structures	N-S Discordantly superimposed on the Paleozoic and Precambrian structures	NE-SW Conforms with the general trends of the Precambrian and Paleozoic structures
Structure	Complex grabens bounded for the most part by step faults and *en echelon* fault zones. Commonly asymmetrical in cross-section and cradle-shaped longitudinally. Vertical displacements up to several thousand metres, largest in the Baikal (> 700 m)			Shield Segments underlain by complex grabens, structurally similar to grabens that make up the classical graben systems. Magnitude of vertical movements not known. The Marginal Segment appears to be underlain by a rift zone which in places has graben-like cross section
Sedimentary-volcanic fill	Fault troughs partly filled with Cenozoic sediments and volcanics, which in the deeper parts of the troughs are several thousands of metres thick, thickest in the Baikal, 2000 to 6000 m.			No Cenozoic volcanics. Floors of the fault troughs covered only by a veneer of unconsolidated sediments usually less than 100 m thick.
Associated Igneous rocks	Olivine basalts normal and mildly alkaline; tholeiitic basalts. Also plutonic and volcanic equivalents of ultramafic alkaline rocks and carbonatites	Similar to the igneous rocks associated with the East African Rift System but greatly undersaturated alkaline rocks and carbonatites are unknown.	Igneous rock assemblages similar to those of the East African Rift System.	Tholeiitic dike swarms, minor flood basalts, alkaline complexes and carbonatites.
Geophysical Characteristics	Rift zones commonly characterized by negative Bouguer anomalies (maximum about 50 milligals), above average heat flow and shallow focus seismicity.			Appears to have no geophysical expression except for mild shallow focus seismicity.
Contemporary stresses	Horizontal tension normal and horizontal compression parralel to the rift zones.		Horizontal compressional stresses of regional extent with the greatest principal stress direction northwest-southeast	Horizontal compressional stresses of regional extent with the greatest principal stress direction east to northeast.

are zones of active rifting at the present time but the contemporary tectonic stresses in the Rhine region appear to be horizontal compressional and evidently different from tectonic stresses under which rifting took place. Unlike the classical rift zones, the St. Lawrence rift system appears to have been an inactive rift during the Cenozoic. Mild seismicity associated with the rift system is probably due to release (on unhealed rift faults) of horizontal compressive stresses that are present over large parts of eastern North America.

5. The pattern of the St. Lawrence Rift system resembles the branching patterns of the classical rift zones.
6. A crustal swell similar to those on which the classical rift zones are set cannot be discerned in the St. Lawrence Region. However, certain features of the Shield Region (e.g. deep erosion, deeply entrenched drainage lines) are compatible with the hypothesis of a crustal swell that has been deeply eroded. Such a crustal swell, if real was not related to the post-Ordovician rifting but to the early development of the ancestral rifts. The block mountains in the areas of Parc des Laurentides massif, Adirondack Mountains and Madawaska Highland may represent skeletal

resurrections of parts of an old crustal swell, during post-Ordovician rifting.

7. With the exception of volcanic features, landforms that are associated with classical rift zones have their counterparts in the St. Lawrence Region. However, the rift valley scenery of the St. Lawrence Region is distinctly less pronounced whether one considers positive or negative landforms. This, as well as the absence of landforms of volcanic origin, can be explained as a result of long inactivity and deep erosion of the St. Lawrence Rift system.
8. Excepting the possible branch structure along the outer part of the Laurentian Channel, the rest of the St. Lawrence Rift system is set on craton. All three classical rift systems are also set on cratons.
9. The gross trend of the St. Lawrence system conforms with the general tectonic trend of the craton on which the rifts are superimposed, but when the trends are considered in detail notable discordances are evident. The above trend relations are not unlike those of the East African Rift system or of the Baikal Rift system. The Rhine graben is discordantly superimposed on the older structures of the region, a situation similar to that of the Shield Segments of the St. Lawrence Rift system.

10. Almost all the structural variants shown by rifts in the three classical areas have their counterparts in the St. Lawrence Rift system. The post-Ordovician rifting does not seem to have brought about more than 500 m of structural subsidence, whereas up to several kilometers of Cenozoic structural subsidence have taken place in the classical areas. Greater structural subsidence may have taken place along the St. Lawrence Rift system, during the evolution of the ancestral structures, but the magnitude of any such movements cannot be determined.

11. The fault troughs of the St. Lawrence Rift system have no noteworthy thicknesses of Cenozoic sediments whereas some parts of the classical rift zones have a few kilometers of sediment in them. The lack of sediment fill in the St. Lawrence fault troughs can be understood in terms of their long inactivity and deep erosion.

12. Tholeiites and transitional types between tholeiites and alkaline basalts appear to have accompanied early rifting along the St. Lawrence Rift system. Later igneous products have a steep bias towards being alkaline (e.g. alkaline-carbonatite complexes of the Lake Nipissing and Monteregian provinces). Apart from the lack of large volumes of volcanics (this can be explained as resulting from long inactivity and deep erosion of the rift system), the igneous products

associated with the St. Lawrence Rift system are typical of continental rift zones.

13. Whereas the rift zones in the three classical areas are associated with conspicuous geophysical anomalies, no notable geophysical anomalies appear to be associated with the St. Lawrence Rift system. The mild seismicity in the St. Lawrence Region appears to be unrelated to rifting processes. The lack of geophysical anomalies is compatible with the long inactivity of the St. Lawrence Rift system.

From the above, it can be seen that the similarities of the St. Lawrence Rift system to the rift systems in the three classical areas are not restricted just to gross patterns but apply to a whole range of criteria including geomorphic, structural and magmatic aspects. There are differences, however, the main ones being the lack of conspicuous geophysical anomalies and the absence of large volumes of volcanic products along the rift system. As already mentioned, these differences can be understood in terms of the long inactivity and deep erosion of the rift system, but they also may partly be related to the rift system's mode of origin which will be discussed in Chapter XV.

Chapter XV: THE ST. LAWRENCE RIFT SYSTEM: GENERALIZATIONS AND SPECULATIONS ON ITS AGE AND ORIGIN

Ages of the St. Lawrence Rift System

As pointed out earlier (see p.194) it is convenient to divide the St. Lawrence Rift structures into two age groups (i) post-Ordovician structures and (ii) ancestral structures. While the evidence for post-Ordovician rifting is conclusive, at least in the better known parts of the rift system, conclusive evidence for ancestral rifting is difficult to obtain. The following are the main lines of evidence for the earlier rifting.

1. The occurrence of regional diabase dike swarms of Hadrynian age, along the entire length of the Shield Segment A and in parts of the Esquiman Channel Area. In the latter area dikes are associated with flood basalts.
2. The occurrence in widely separated parts of the St. Lawrence Valley system of alkaline-carbonatite complexes (see Table III) and presumably related dikes, giving K/Ar ages near the Hadrynian-Cambrian boundary.
3. The lithological and sedimentalogical characteristics of the basal arenites of the platformal sequence (St. Lawrence Valley: Lewis 1971 and p.59 of this paper. Champlain Valley: Wisnet 1961 and p. 76 of this paper. Esquiman Channel Area: Clifford 1969 and p. 93 of this paper. Lower Ottawa Valley: Wilson 1946 and p.112 of this paper) indicate that

the arenites accumulated in fault-margined, graben-like basins. The upper parts of these arenites are Cambrian but the lower parts may be Hadrynian.

In addition to the alkaline-carbonatite complexes with K/Ar ages near the Hadrynian-Cambrian boundary, there are in the St. Lawrence Region, older alkaline-carbonatite complexes with K/Ar ages as much as about 1000 m.y. (see p. 238). Although the emplacement of these older intrusions might be connected with events that led to the formation of the St. Lawrence Rift system, the earliest and most pervasive igneous activity along the St. Lawrence Valley system has been the emplacement of diabase dike swarms along the entire length of the Shield Segment A (Grenville dike swarm) and along the north-eastern parts of the Esquiman Channel Area. In the latter area the dike swarms are associated with flood basalts (Light House Cove formation). These flood basalts appear to be remnants of once extensive plateau basalt fields. The writer proposes that the emplacement of these dike swarms, possibly accompanied by large scale flood basalt eruptions, is related in time and cause with early rifting along the St. Lawrence Rift system. The Tibbit Hill volcanics may also be a remnants of the early rift volcanics (see p.235). The precise ages of the above igneous rocks are poorly known. The dikes in the north-eastern Long Range Mountains, which are presumed to be comagmatic with dikes and flood basalts in the Esquiman Channel Area, have yielded

K/Ar ages of about 805 m.y. (Pringle et al. 1971). The isotopic age of the Grenville dike swarm (along Shield Segment A) is poorly known, but has been quoted as 675 m.y. (Fahrig 1972, p. 576). The Tibbit Hill volcanics may be early Cambrian or late Hadrynian (Cady 1969, p. 148). Thus, the existing evidence indicates that the development of the above diabase dike swarms and volcanic rocks may have continued from approximately 800 m.y. ago to the beginning of the Phanerozoic. Such an age spread is perhaps ~~not unusual~~, for as Sutton (1972, p. 366) has pointed out that diabase dike swarms along rift systems develop over periods of 200 - 300 m.y., although this is not established with certainty. The alkaline-carbonatite complexes with K/Ar ages of about 565 m.y. (see Table III and Fig. 55) and the comagmatic dike swarms appear to have been emplaced towards the end of the period of diabase diking and basaltic eruptions along the rift system. Thus the indications are that the St. Lawrence Rift system began about 800 m.y. ago. The Light House Cove lavas and Tibbit Hill volcanics occur at or near the bases of the Appalachian stratigraphic sequences. Therefore, in the areas where these volcanics occur, rifting appears to have taken place before or about the time the geosynclinal fills began to accumulate.

It was suggested earlier that certain features of the Shield Region are compatible with the concept of a deeply eroded crustal swell which may have formed as a prelude to the initiation of the St. Lawrence Rift system (see pp. 197 to201). If such a crustal swell did exist and did rupture, rapid erosion of its steep edges (facing the rifts), would lead to accumulation of coarse immature clastics in the fault troughs. In the Shield Segments no such clastics are known at the base of the early Paleozoic platform cover. One explanation is that clastic accumulations were removed by deep erosion of the rifts, before the early Paleozoic marine transgression. In parts of the Marginal Segment Rift Zone, however, the platformal sequence begins with coarse immature clastics. Examples of such deposits are the Covey Hill formation in the St. Lawrence and Champlain Valleys (see p. 47) and the Bateau formation in the Belle Isle area (see p. 85). The Bateau formation is probably Hadrynian because it underlies the Light House Cove formation of probable Hadrynian age, and as already mentioned earlier, the lower part of the Covey Hill formation may also be Hadrynian (Hofmann 1972, p. 4).

Recently, Rodgers (1972) has pointed out that along the western Appalachian basement anticlinoria, the Appalachian miogeosynclinal succession usually begins with large volumes of coarse clastics (e.g. Labrador group in northwestern Newfoundland; Mendon group to Cheshire quartzite in central Vermont) that appear to have accumulated in rather uneven basins during a time period extending from the Early Cambrian well back into the Hadrynian but always post-Grenville. He states (p. 511) that

> "the provenance of the clastic material is uncertain; directional indicators in later better sorted sediments suggest mainly a northwestern source. The coarseness of the debris and the abundance of feldspar in the less well sorted rocks suggests a relatively mountainous terrain exposing much granite or gneiss. This terrain, might have been a mountain chain produced in the Grenville belt by the Grenville orogenic cycle, or at least by the later movements of that cycle".

The writer proposes that the clastic debris were derived from the Shield Region (principally the Grenville Province) following the domal uplift of the area during the early development of the ancestral structures of the St. Lawrence Rift system. Inclusion of these clastics as rift sediments implies that the deeper fault troughs related to the ancestral structures occurred to the east and south of the Marginal Segment Rift Zone. Such a configuration of the ancestral structures is consistent in the context of the rift system's origin as hypothesized in the section to follow.

Origin of the St. Lawrence Rift System

There are two main aspects to the problem of origin of the St. Lawrence Rift system. One is the question of the origin of large rift systems in general, and the other is the close space and time association of the St. Lawrence Rift system with the Appalachian foldbelt.

The first aspect of the problem has been extensively studied in relation to other large rifts (continental as well as oceanic) and the results of these studies are briefly discussed in Appendix VI. The second aspect which is specifically related to the St. Lawrence Rift system will be the main subject of the discussion in this section.

Any genetic model of the origin and evolution of the St. Lawrence Rift system must provide a coherent rationale for the following space and time relations between the rift system and the Appalachian foldbelt.

1. The Marginal Segment Rift Zone lies along the northwestern edge of the foldbelt.
2. The Shield Segment A graben system extends into the continental interior from the axis of a salient - the New England Salient - of the foldbelt. (The Shield Segment B graben which appears to be one of several lateral fractures related to the St. Lawrence Rift system (see Fig. 43) is apparently not associated with any special feature of the foldbelt).
3. The postulated branch of the rift system, along the outer part of the Laurentian Channel extends from a reentrant of the foldbelt.

4. The first appearance of the ancestral structures of the rift system and the beginning of the Appalachian stratigraphic sequence may be near contemporanous. The ancestral structures appear to have undergone reactivations during the evolution of the Appalachian foldbelt.
5. The greater part of the post-Ordovician structures of the rift system may have formed in the mid-Mesozoic after the Appalachian geosyncline ceased to be a mobile belt.

From the above, it can be seen that the rift system and the foldbelt are intimately related in space. Furthermore, the evolution of the ancestral structures of the rift system, appears to be related in time with the geosynclinal evolution. These space and time relationships suggest a causal relationship (see Saull 1967) indicating that not only the foldbelt but also the ancestral structures of the rift system are products of geosynclinal evolution. If this were so, it should be possible to explain the origin of the ancestral structures of the rift system within the framework of models of Appalachian evolution.

Recently, proposed models of the Appalachian evolution have already been discussed earlier in Chapter II. Of these, Bird-Dewey model (1970) and Chidester-Cady model (1972) were considered as representative of the two main schools of thought. Although the two models are fundamentally different, both suggest that the geosynclinal activity began, late in the Precambrian, when a rift system developed somewhere along the present site of the Appalachian eugeosynclinal belt, possibly just east and south of the edge of the Cambro-Ordovician North American continent as interpreted by Rodgers (1968, also see Fig. 4). The rift system which will hereafter be referred to as the eo-Appalachian rift system was set in a continent consisting of North America and Africa then joined together. The rift later widened to form the Proto-Atlantic Ocean (Bird-Dewey model) or a downwarped zone occupied by a broad inland sea similar to the present Mediterranean with a dominantly sialic floor (Chidester-Cady model). Whichever the case might be, the writer proposes that the St. Lawrence Rift system can be understood as a feature inherited from the eo-Appalachian rift system, which with continued activity led to the development of the Appalachian geosyncline.

The crustal upwarping and distension related to the formation of the eo-Appalachian rift system and further crustal distension related to subsequent widening of the rift (Bird and Dewey 1970; Chidester and Cady 1972) led to extensive block faulting of the crustal plates on either side of the main rift. To the north and west of the rift, the block faulting extended as far as the present limit of the Marginal Segment. Much of this block-faulted terrain was subsequently incorporated into the foldbelt except in the Marginal Segment Rift Zone, where high-angle faults remained largely unhealed and susceptible to later reactivation. The eo-Appalachian rift system may have contained two triple junctions which formed over local

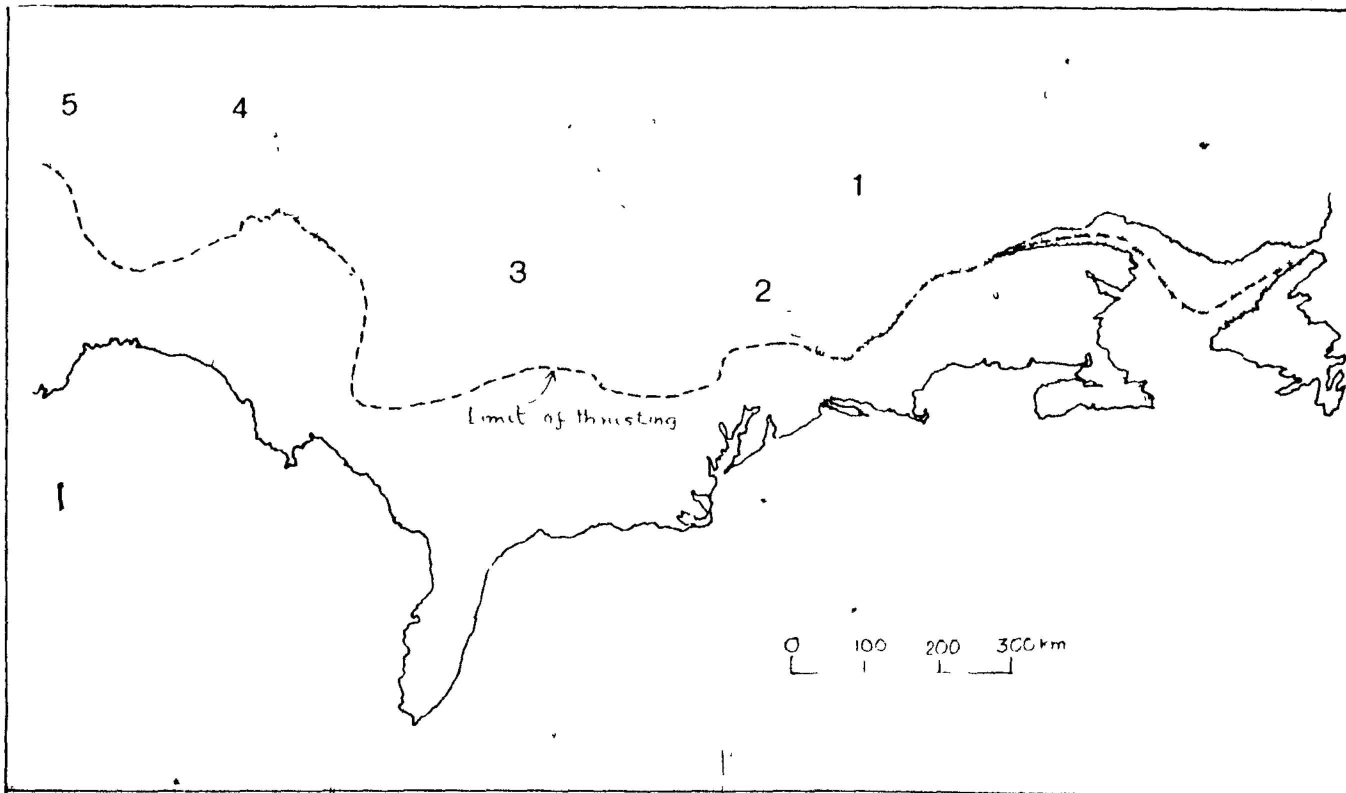


Fig.57. Map showing possible aulacogens associated with the Appalachian and Ouachita foldbelts. 1. Shield Segment A graben system 2, South New York graben 3. Rough Creek graben 4. Wichita graben 5. Marathon graben. **Base map from Wilson 1959.**

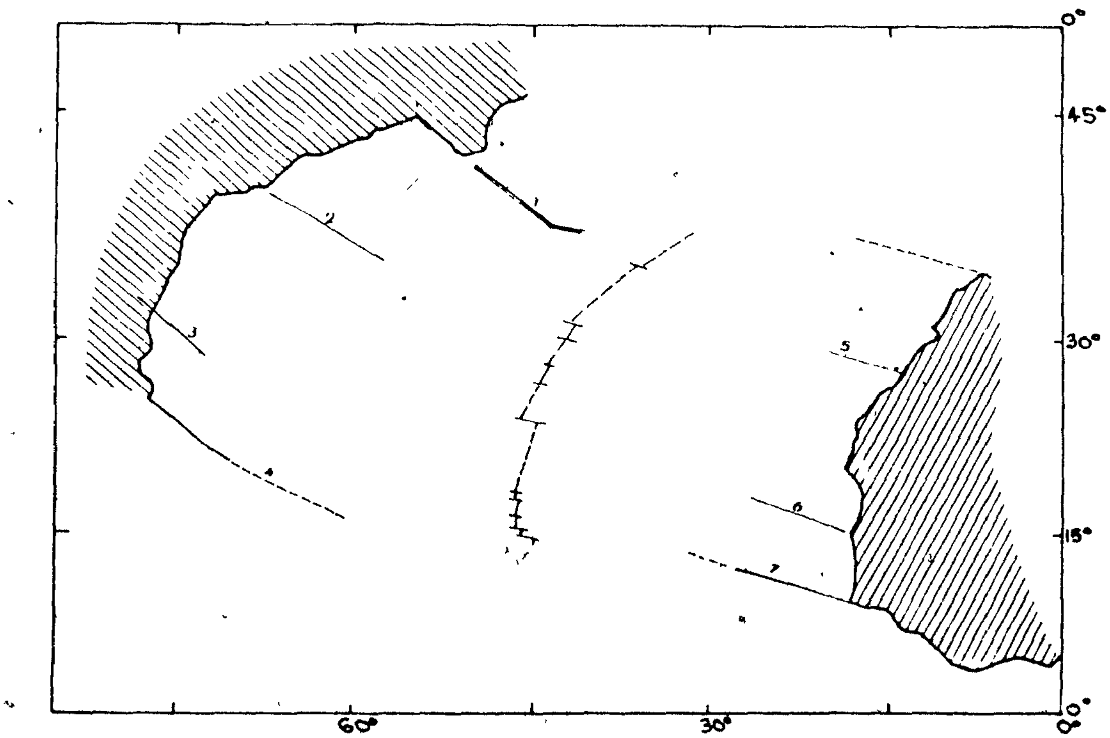


Fig.58. Possible fossil transform faults of the North Atlantic sea floor: 1. Newfoundland Fracture Zone 2. Kelvin F.Z. 3. Cape Fear F.Z. 4. Bahama F.Z. 5. Canary F.Z. 6. Cape Verde F.Z. 7. Guinea F.Z. After LePichon and Fox (1971)

mantle upwellings or plumes of hot material (see Morgan 1972). An arm of one of the triple junctions extended along the axis of the New England Salient and thence across the southern part of the St. Lawrence Valley and along the Shield Segment A (see Burke and Dewey 1973, pp. 420-421). This arm later became an aulacogen* and the other two arms became parts of the geosyncline. This explains the transverse trend of the Shield Segment A to the foldbelt and the location of the former at a salient of the latter (see Fig. 57). The second triple junction lay under the present site of the Gulf of St. Lawrence. One arm of this three-pronged rift extended along the site of the outer part of the Laurentian Channel, the other two arms became parts of the geosyncline. Although the former lay across the site of the geosyncline, it survived the mountain building processes as a relic fracture zone, just as the eastward extension of the Shield Segment A graben system seems to have survived as an unhealed fracture zone (see p. 256). It is proposed that as a result of tensional stresses related to the opening of the Atlantic Ocean in the Mesozoic, the unhealed faults along the Marginal Segment, the

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*Unlike the "type" examples of aulacogens (Shatski 1946, 1947, 1955; Nalivkin 1963) in the USSR (the subsurface Pachelma and Dnieper-Donec, aulacogens in the southeast part of the Russian platform), the Shield Segment A graben system does not contain great thickness of sediments. Neither is there any evidence that it developed into a wide intracratonic downwarp and went through a mild compressional phase during later stages of its evolution. However, it has the basic characteristics of an aulacogen in that it is an intracratonic graben that extends from a foldbelt far into the interior of the foreland craton (Hoffman <u>et al</u>. 1973). Moreover, if rifts that did not develop into geosynclines are regarded as aulacogens (Nalivkin 1963; Hoffman <u>et al</u>. 1973) then according to the genetic model presented in this paper, the Shield Segment A is an aulacogen.

Shield Segment A and the relic of the branch structure along the outer part of the Laurentian Channel became reactivated and combined to form the St. Lawrence Rift system as proposed in this thesis.

When the initial break of the continents occurred in the Mesozoic, the fracture zone along the outer part of the Laurentian Channel led to the creation of a continental margin offset (see Fig. 58). This continental margin offset in turn led to an offset on the plate accretion axis (see Wilson 1965), i.e. a transform fault. As the Atlantic Ocean continued to open, the residual inactive trace of the transform fault became the Newfoundland Fracture Zone (Le Pichon and Fox 1971). The above genetic model takes into account all five requirements mentioned earlier. According to the model, the ancestral structure of the St. Lawrence Rift system originated as parts of the eo-Appalachian rift system. Thus, the genetic relations are the same whether one relates the Belle Isle and Tibbit Hill volcanics (p.235) to the ancestral structures of the St. Lawrence Rift system or to the eo-Appalachian rift system. The same can be said of the dike swarms(p.233-234) alkaline carbonatite centres (p.236) and sediments (p.269) which were interpreted as related to the development of the ancestral structure. The model implies that the pattern of the St. Lawrence Rift System is also largely inherited from the pattern of the eo-Appalachian rift system. Also the sigmoid shape of foldbelt is interpreted as a feature inherited from the branching pattern of the eo-Appalachian rift system. It follows that the apparent skewness of the Newfoundland arc with respect to the mainland arc (see p.25) also can be viewed as a feature inherited mainly from the shape of the eo-Appalachian rift system. Mention should be made here that Black's (1964) finding of a 30-degree discrepancy between paleomagnetic pole positions of Lower Cambrian rocks of western Newfoundland and rocks of similar age in New Brunswick and Nova Scotia, has commonly been cited (e.g. Clifford 1969) as evidence of a 30-degree counterclockwise rotation of Newfoundland. Pole positions of Carboniferous rocks from Newfoundland and the mainland apparently coincide and hence the rotation is believed to have taken place in the Devonian or earlier. Later work by Robertson et al. (1968) has shown that because of uncertainties involved in detecting the components of remanent magnetization, the available paleomagnetic results cannot be used to establish conclusively the hypothesis of a rotation of Newfoundland. They state that paleomagnetic data do not provide independent evidence that rotation of Newfoundland has occurred.

Coarse immature clastics similar in age and lithology to those that were interpreted as rift sediments of the eo-Appalachian system (p. 269) also occur along the western basement anticlinoria of the central and southern Appalachians (Rodgers 1972, p. 510). Also volcanics similar in petrochemistry and age to the Belle Isle and Tibbit Hill volcanics (which were interpreted as eo-Appalachian rift volcanics) occur further to the south (e.g. Catoctin lavas: see Reed, J.C., Jr. 1955, Reed and Morgan 1971, pp. 526-548, Strong and Williams 1972, pp. 43-54) along the west side of the Appalachian foldbelt. Thus, if rifting of a larger continent led to the initiation of the geosynclinal cycle in the northern Appalachians, it is probable that the rifting occcurred along the entire length of the site now occupied by the foldbelt (see Rodgers 1972, p. 517). Also, the tensional stresses that led to the opening of the Atlantic Ocean must have prevailed along the entire length of the present continental (North American) margin. Therefore, it follows that rifts similar to the St. Lawrence Rift system are to be expected in association with the central and southern Appalachians. Also other fossil transform faults similar to the Newfoundland Fracture Zone can be expected in the North Atlantic, off the continental margin of North America. No feature comparable to the Marginal Segment Rift Zone is known from the central and southern Appalachians, but there is at least one prominent rift zone or aulacogen - the Rough Creek graben (Fig. 57) - which like the Shield Segment A graben system, extends from a salient of the foldbelt into the continental interior (Wilson 1959). A possible basement graben referred to by Wilson (1959) as the south New York graben (Fig. 57), also appears to have space relations (with respect to the Appalachian foldbelt) similar to that of Shield Segment A. The origin of Kelvin Fracture Zone, Cape Fear Fracture Zone and Bahama Fracture Zone (Fig. 58) may be similar to that of the Newfoundland Fracture Zone.

MAJOR CONTRIBUTIONS

Some of these contributions were anticipated in a joint paper by Kumarapeli and Saull (1966a). The others have been developed independently, since then, except as noted in the body of the text, where collaboration and prior work is acknowledged.

1. Evidence has been collected, from previous published material and from field and airphoto studies of the St. Lawrence Region and synthesized to show that the St. Lawrence Valley system is carved out mainly along rift zones. This work has led to new interpretations as grabens or graben-like forms for the structures along the St. Lawrence and Champlain Valleys, inner and middle parts of the Laurentian Channel and the Esquiman Channel Area. The outer part of the Laurentian Channel has been interpreted as carved out along a fracture zone of uncertain type.
2. A mega-structure - the St. Lawrence Rift system previously unknown in the form and extent described in this paper, is proposed. A hypothesis of its origin is advanced.
3. The St. Lawrence Rift system is compared and contrasted with the continental rifts of the three classical areas, namely East Africa, Rhine and the Baikal. The main differences of the St. Lawrence Rift system are explained in terms of its long inactivity and its mode of origin.
4. Some of the geomorphic peculiarities, such as the deep valleys of the Shield Region and the unusually deep erosion of the

Grenville Province are interpreted as features inherited from deep erosion of a crustal swell in the St. Lawrence Region.

5. Ten gravity profiles across Logan's Line which extends along a part of the east margin of the St. Lawrence Rift system have been compiled and their results are interpreted.

6. One gravity profile across the Shield margin on the west side of the St. Lawrence Valley has been compiled and the results are interpreted (Kumarapeli and Sharma 1969).

7. The recurrent alkaline magmatism in the St. Lawrence Region has been interpreted as related to tectonic activity along the St. Lawrence Rift system.

8. A seismotectonic scheme, that the mild seismicity of the St. Lawrence Region is related to release of contemporary regional stresses on the fractures related to the St. Lawrence Rift system, is presented.

9. A geological basis for the presence of explosion craters in the St. Lawrence Region is proposed.

10. A hypothesis explaining the apparent skewness of Newfoundland with respect to the mainland is advanced.

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APPENDIX I: PHYSIOGRAPHIC CHARACTERISTICS OF THE APPALACHIAN REGION

Physiographic Expression

The physiography of the Appalachian Region (Goldthwait 1924, Bostock 1970, King 1972a) is dominated by uplands - the Atlantic Uplands - whose height above sea level varies from about 200 m to about 800 m. The uplands consist largely of rolling terrain, developed over relatively resistant rocks. Their continuity is interrupted by belts and broad areas of lowland that are collectively referred to as the Carboniferous-Triassic Lowlands. These lowlands have developed on less resistant rocks.

The surface of the Atlantic Uplands slopes gently to the southeast. The higher northwestern part is characterized by prominent mountain ranges: Long Range Mountains (altitude of the highest peak 814 m), Notre Dame Mountains (1268 m), Sutton Mountains (968 m) and Green Mountains (1130 m) (Fig. 1). Broadly viewed, these ranges form a high rim that overlooks the adjacent St. Lawrence Valley system depressions. The southeast sloping surface of the region meets the waters of the Atlantic Ocean in a drowned coastline which has been subsiding at the rate of 15 cm/century during the past 4000 years (Grant 1970).

Physiographic Evolution

Earlier workers (e.g. Goldthwait 1924, Twenhofel and MacClintock 1940) recognized dissected and tilted peneplains (see Fig.59)

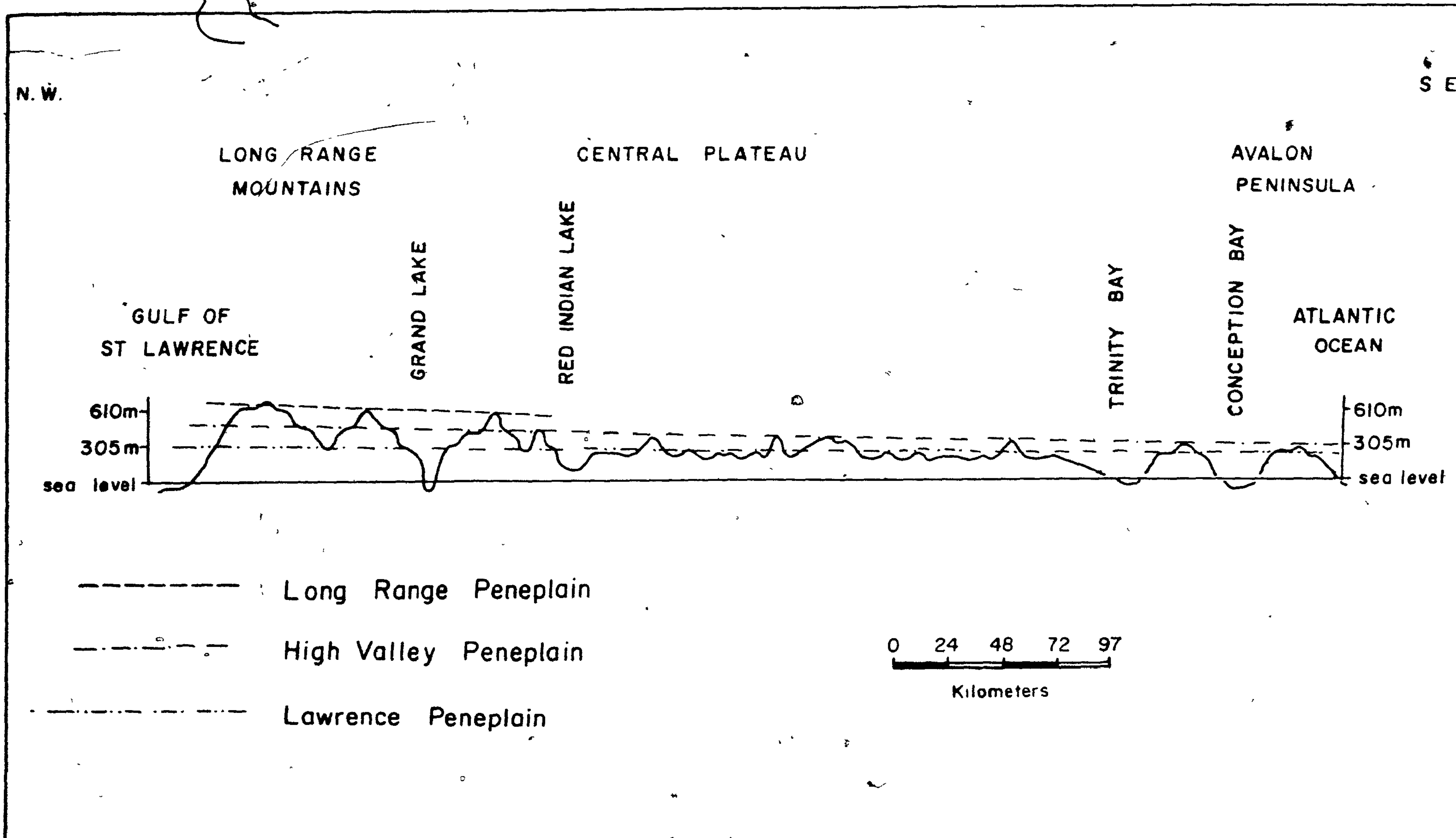

Fig. 59. Three east sloping erosion surfaces of Newfoundland as postulated by Twenhofel and MacClintock (1940).

in the upland surfaces of the Appalachian Region and discussed the physiographic evolution of the region in the light of Davis' hypothesis of cyclical erosion (Davis 1889). Modern geomorphic theory (Hack 1960), however, considers that naturely dissected landscapes such as those of the Atlantic Uplands are essentially in a state of balance between the processes of erosion and resistance of the rocks. King (1972a) has recently suggested that the uplands and lowlands of the Appalachian Region developed progressively together in response to a long and possibly continuous erosion cycle that had its beginning in the Jurassic Period. King (1972b) further speculates that the uplift which initiated this erosion cycle might have been related to ramping (see Vogt 1970) of the continental margin, during the initial stages of continental rifting that led to the opening of the present Atlantic Ocean. Uplift was followed by subsidence and sedimentation along the continental margin and the adjoining ocean basin; the latter processes seem to have persisted to at least the end of the Tertiary Period (King 1972a).

APPENDIX II:

PHYSIOGRAPHIC CHARACTERISTICS OF THE SHIELD REGION

Physiographic Setting

The Canadian Shield is approximately elliptical in plan. Its morphology, considered in the broadest sense, has been described as crudely saucer-shaped (Cooke 1947, p. 11), because in its central part is a great depression - the Hudson Bay depression - outward from which land generally rises in all directions to form a wide rim of highlands. The Shield Region is in the southeastern segment of the highland rim. A major portion of the region is referred to as the Laurentian Highlands (Bostock 1967). The part near the Atlantic seaboard is known as the Mecatina Plateau (Bostock 1967).

A large part of the high rim of the Canadian Shield is fairly uniform in elevation. Altitudes commonly vary between 200 m and 500 m. The rim is considerably higher in parts of the Shield Region, where elevations of 500 m to 1,000 m are common. The Shield margin attains mountainous heights in northern Labrador (2,000 m) and in Baffinland (2,500 m to 3,000 m).

Topography

The Shield Region can be regarded as a composite plateau; much of its surface stands about 400 m above sea level. Within the region, several areas with elevations systematically higher than those of the surrounding areas, can be recognized. A large area of higher elevations lie between the Betsiamites and Romain Rivers. Much of the surface in this area lies between 500 m and 1,000 above sea level. To the northeast of this area, the Mecatina Plateau has elevations of 500 m or less, and to the southwest, the elevations generally decline towards Georgian Bay but not without interruptions. These interruptions occur in the areas of the Parc des Laurentides massif (maximum elevation 1,166 m), Adirondack massif (1,639 m) and Madawaska Highlands (587 m). These three upstanding topographic masses are fault-block mountains and stand several hundreds of meters above the general level of the surrounding areas.

At the margins of the St. Lawrence Valley system depressions, the surface of the Shield Region descends abruptly towards the depressions. These steep edges of the region are mostly interpreted as fault-line scarps, 100 to 300 m high. The related faults are discussed in detail in Chapters IV through to XII. The scarps are deeply dissected, so much so that when the Shield Region is viewed from within the St. Lawrence Valley system depressions, it has a mountainous appearance, and in some places ruggedly so.

Generally speaking, the surface of the Shield Region is characterized by low to moderately high rocky hills and ridges, interspersed with valleys and sundry-shaped depressions, the floors of which are invariably mantled by glacial debris and often occupied by swamps and lakes. Some of the valleys have remarkably linear or zig-zag courses that range in length from a few kilometers to several tens of kilometers.

The areas of higher elevations mentioned earlier are deeply dissected and are characterized by rugged topography. Relief of about 300 m is common in these areas. The rest of the Shield Region is less deeply dissected, relief on the average being about 100 m.

Drainage Characteristics

The present drainage system of the Shield Region has been modified to some extent by Pleistocene glaciation. A typical main drainage line consists of a succession of lakes connected by stretches of fast water. The lakes are merely expanses of quiet water impounded in valleys by dams of glacial debris. Here and there, however, deep and well-defined valleys persist for long distances. Some of these valleys attain depths of 300 m or more and yet even the deepest of them is filled to a greater or lesser extent with glacial debris. Examples of deep valleys are those of rivers Jacques-Cartier, Gouffré, Saguenay, Ste. Marguerite, Betsiamites, Manicouagan and Moisie. These valleys are more or less transverse to the Shield Region and at their lower ends join either the St. Lawrence Valley or the Laurentian Channel.

Typically, the deep valleys are drowned for a few kilometers upstream from their lower ends, and beyond that for several kilometers they commonly have a fairly uniform grade on reworked glacial sediments. Above the graded section there is usually a stretch of narrow gorges or canyons and at some point in the latter section the present river descends abruptly from the plateau above, into the deep canyon below (thus, these rivers have great hydro-electric potential). These valleys did not form in post-glacial time for post-glacial streams are still in the process of eroding and reworking the glacial fill in them. According to Ambrose (1964) not only the deep valleys but also the entire drainage system of the Shield Region was established and structure-adjusted in pre-Ordovician times and became reactivated when it was gradually exhumed in relatively recent times.

A noteworthy feature of rivers in the deep valleys is that their sources lie to the north of the present height of land but the deep canyons enable them to flow southwards against the regional slope. For example, the Saguenay after heading in Lac St. Jean, which is no more than 100 m above sea level, crosses a plateau whose altitude is about 350 m. The zig-zag course of this river across the high-plateau, however, lies along what appears to be deeply eroded fault lines. The other rivers in deep valleys also have conspicuously linear or zig-zag courses and appear to be controlled by faults and/or joints.

The drainage pattern of the Shield Region, with many of its main rivers deeply entrenched and flowing against the regional slope in their upper courses, fascinated early workers (e.g. Wilson 1903; Cooke 1929). A view commonly expressed by them, is that the rivers in deep valleys are examples of antecedent drainage lines: originally, the rivers had established their southward flow as consequent streams and were able to maintain their course against a reversal of regional slope. This hypothesis calls for a state of rather delicate balance to exist for a long period of time, between the presumed rates of reversal of the regional slope and of down-cutting of rivers, a situation perhaps unlikely or at most rare in nature. Hence, this view is now largely abandoned, although the writer knows of no alternate view advanced to explain the origin of these unusual drainage lines. It is argued in Chapter XIV that the peculiarities of these drainage lines were inherited by deep erosion of structures impressed on the Shield Region during the initiation and development of the ancestral structures of the St. Lawrence Rift system.

Views on Physiographic Evolution

Some workers have postulated upland erosion surfaces in the Shield Region (Dawson 1897; Wilson 1903, p. 630; Cooke 1947, p. 11; Blanchard 1933, see Dresser and Dennis 1944, p. 200; Parry 1962). Several workers have correlated these surfaces with the presumed Cretaceous and/or Tertiary surfaces of the Appalachian Region (Crowl 1959, Parry 1962). Such correlations, however, seem untenable in the light of later studies by Ambrose (1964). He has advanced the hypothesis that the surface of the Shield Region is a pre-Ordovician paleoplain that has undergone little modification since its exhumation after being buried under a Paleozoic platform cover. The principal basis for this view is the presence of small outliers of Lower Paleozoic beds (Ordovician) in the Shield Region (Fig. 5). Few of these outliers seem to occupy hollows of the paleo-surface (e.g. see Osborne 1937, p. 9). Also in some places, Lower Paleozoic rocks extend as "tongues" along valleys cut in the old surface (e.g. see Osborne 1956, p. 159). These criteria do support Ambrose's view (1964). Yet, the majority of Lower Paleozoic outliers including all the ones that are far removed from present areas of extensive platform cover, appear to be preserved as downfaulted blocks in grabens. Thus, the preservation of these outliers do not necessarily prove Ambrose's (1964) contention that "no great lowering of the Precambrian surface has occurred since exhumation".

APPENDIX III

GRAVITY STUDIES OVER LOGAN'S LINE

INTRODUCTION

The western boundary of the Appalachian foldbelt lies along the eastern half of the St. Lawrence Valley and is known as Logan's Line or Logan's fault after Sir William Logan (1863) who first recognized it. In geological maps this feature is usually shown as a continuous, east-dipping thrust fault that extends from the north-east end of Lake Champlain to the vicinity of Quebec City and beyond (Fig. 9). However, outcrops indicating its position and nature are exposed only in Lake Champlain and Quebec City areas. In the former area, Logan's Line is, in fact, a zone of thrust faults (Clark and McGerrigle 1944, p. 396) dipping 20° towards the Appalachian fold-belt. In Quebec City area too, it is an east-dipping thrust fault, but the dip there ranges from 40° to 70° (Ells 1888; p. 15K; Raymond 1913; Graham and Jones 1931). In between the two above localities which are actually about 200 km apart, Logan's Line is placed to explain stratigraphic anomalies and widespread brecciated zones (e.g. see Clark 1964a, p. 75). The gravity traverses discussed below were carried out across the presumed trace of Logan's Line in this 200 km section (Fig. 60) where neither its exact position nor its fault geometry is known.

Kumarapeli and Saull (1966a, p. 665) discussed the nature of Logan's Line in the context of possible rifting along the St. Lawrence

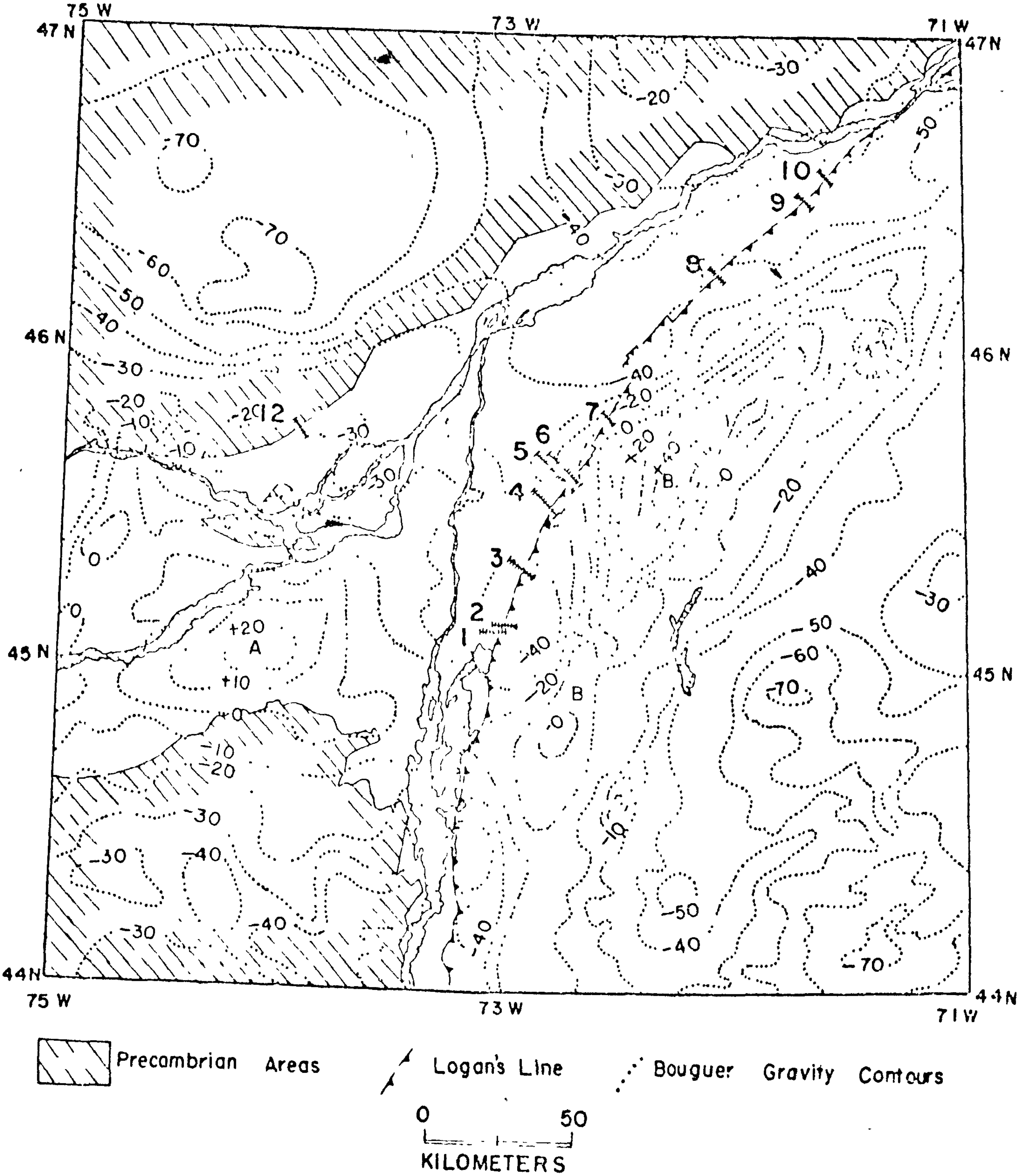

Fig.60. Gravity map of the St.Lawrence Valley and vicinity, after Innes and Argun-Weston (1967). The locations of gravity profiles (1 to 12) compiled by the writer are shown.

Valley and emphasized the problematic nature of structures included in Logan's Line. They suggested the possibility that the scale of thrusting postulated by some of the earlier workers may be exaggerated, implying that the thrust faults may steepen downwards and extend into the Precambrian basement. The gravity studies were undertaken (during the summers of 1966 and 1967) on the following assumptions (i) the subsurface structure of Logan's Line includes a basement step, the downdrop being towards the St. Lawrence Valley (ii) there is a significantly high density contrast between the basement and cover rocks. However, contrary to assumption (i) above, the concensus has grown in recent years, in support of gravity driven "thin-skinned" tectonics on Logan's Line (e.g. Williams et al. 1972, pp. 192-194; Zen 1972, p. 13).

GENERAL GEOLOGY

Rocks on the west side (foot-wall side) of Logan's Line are Middle and Late Ordovician in age. They belong to the parautochthonous St. Germain complex (Clark 1956, p. 4) and consist of an intensely deformed sequence of shales and calcareous shales with minor sandstone and limestone bands. An interruption in the St. Germain complex occurs in the St. Dominique area. There, an elongate "island like" block of relatively undeformed platformal rocks emerges through the crumpled St. Germain rocks (Fig. 61). The traditional interpretation has been that these platformal rocks have been pushed up as a thrust slice - the St. Dominique slice

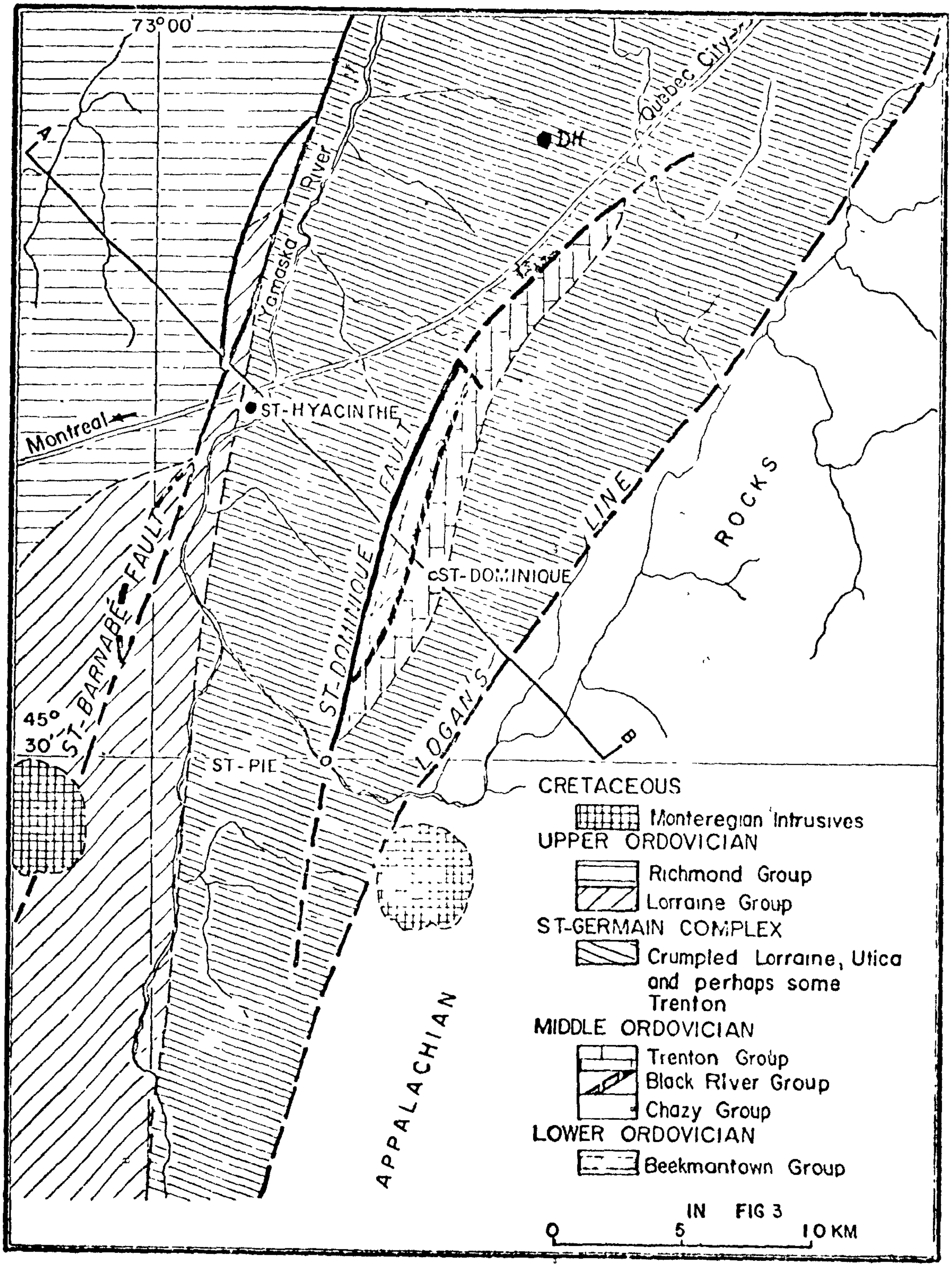

Fig.61. Geological map showing the relations of "St.Dominique slice" to Logan's Line and to St.Germain Complex.

(Clark 1947, p. 17) - during the general westward thrusting along Logan's Line. Recently, the writer suggested that the St. Dominique slice may have been pushed up by block uplift along high-angle faults that are related to the normal faults of the St. Lawrence Valley (Kumarapeli, 1969). Subsequent drilling at the west margin of the St. Dominique slice, however, has shown that certain formations are repeated several times by faulting (Quebec Dept. Natural Resources, open file) a situation supporting the thrust fault hypothesis.

The rocks on the east side (hanging-wall) of Logan's Line are Cambro-Ordovician. They consist mainly of sandstones and shales of the Sillery group except in the southern part where gravity traverses 1 and 2 are located. There, the hanging-wall rocks are a part of the Stanbridge complex (Clark and Eakins 1968), mainly composed of pelites with minor limestone lenses, and on the whole not too different from the ~~rocks on~~ the foot-wall side.

Along the gravity traverses bedrock is exposed only at a very few places. Elsewhere, the surface material consists of unconsolidated sands and clays whose mean density may be as low as 2 g/cm^3. The variations in thickness of this low density material is not known, but thickness up to about 30 m might be expected.

FIELD DATA COLLECTION

The gravity traverses were made along roads that cross Logan's Line at or nearly at right angles (Fig. 60). The lengths of the profiles varied from 5.6 to 18.9 km and the total length of the traverses is 106.8 km (Table VI). The station spacing was approximately 244 m (800 ft), except in profile 8 where a 122 m (400 ft) station spacing was adopted. The gravity measurements were carried out with a Worden gravimeter and the elevations of stations were determined by differential levelling using precision instruments. Drift of the gravimeter was checked every hour or so against a "floating" base station established on each traverse.

Over 125 samples of rock, most of them weighing more than 300 g, were collected for density determinations. The samples were taken from outcrops on or near the gravity traverses. A sampling bias can be expected because rocks that have a higher resistance to weathering and erosion tend to outcrop more frequently than those that are less resistant. Excepting for the Sillery group, in which sandstone members appear to outcrop far more frequently than the shale members, the sampling bias is probably not serious.

ROCK DENSITIES

Parasnis (1952), in a study of rock densities of the English Midlands, emphasized that the "field" density of a rock must lie between the dry and saturated densities. Because the amount of pore water in a rock _in situ_ is unknown, Parasnis suggested that the saturated density is probably the best approximation to "field" density. The rocks in the survey area are mostly sedimentary rocks. In sedimentary rocks the difference between the dry and the saturated density is often very significant. In the present study, densities

TABLE VI

SUMMARY DATA: GRAVITY PROFILES ACROSS LOGAN'S LINE

Profile No. in Fig.	No. of Stations	Length of Profile in km
1	41	10.46
2	39	9.49
3	41	9.49
4	51	12.43
5	80	18.88
6	69	16.41
7	24	5.63
8	68	8.05
9	33	7.56
10	35	8.37
	481	106.77

were determined after the samples had been immersed in water for about 1 week, using the following relation $\rho = w_1/(w_1-w_2)$ where ρ is the density, w_1 is the weight of the sample in air, w_2 is the weight of the sample in water. A total of 104 density determinations was made. Location of samples, their description and density values are given in Tables VII to X. The number of samples (N), the mean density (M) and the standard deviation are listed for each rock grouping in Table XI. The mean density obtained for the Sillery group is probably the least reliable. In addition to the sampling bias as already discussed, the samples taken from the shale members were found to be weathered, although every attempt was made to obtain samples as fresh as possible. On both counts above, the mean density of the Sillery group, as obtained from the density determinations, may be on the low side.

GRAVITY FIELD DATA REDUCTION

The land surface along the traverses and in the nearby areas is characterized by gentle slopes. The grades along the traverses are commonly less than 1%. The maximum grade encountered was about 4%. Accordingly, topographic corrections to the gravity readings were regarded as unnecessary.

The mean densities of rock groups (as determined from density determinations, see Table XI) on either side of Logan's Line vary from 2.62 g/cm^3 for the Sillery group to 2.67 g/cm^3 for the Stanbridge complex. A value of 2.65 g/cm^3 was taken as an approximation of the Bouguer density. Using this value for ρ_B, a combined Free Air and Bouguer correction of 0.1975 mgals/m (0.060 mgals/ft) was applied to every reading. On each profile, the "observed" Bouguer gravity values of stations were calculated relative to a "floating" base, to which a gravity value of zero was assigned. The Bouguer gravity profiles along the 10 traverses are shown in Figs. 62 to 71.

DISCUSSION OF GRAVITY PROFILES

Profiles 1, 2 and 3

These three profiles are discussed together because they are essentially similar (Figs. 62, 63, 64). Going from west to east, all three profiles show negative horizontal gravity gradients: 0.75 mgals/km, 0.5 mgals/km and 0.3 mgals/km respectively. Just about where the profiles cross the geologically inferred trace of Logan's Line, the gradients change from the above values to near zero in all three cases. An examination of the regional gravity map of the area (Fig. 60) shows that the three profiles lie across an approximately north-south trending gravity trough between two gravity highs (A and B in Fig. 60). The east-sloping linear segments of the gravity profiles probably lie on the east flank of the gravity high A. The eastward flattening of the slopes may be due to:

i) a gravity effect related to Logan's Line

ii) a change in the regional gravity trend reflecting proximity to gravity high B

iii) a cumulative effect of both (i) and (ii) above.

There is little basis for giving preference to any one of the three models given above. However on profile 3*, there is the suggestion that the slope after flattening resumes its earlier value further to the east. Such a feature, if real, indicates that the change in the gravity field on this profile may be explained using a "step model" (Grant and West 1965, p. 282) which in two dimensional gravity analysis is customarily used to represent a fault. Assuming that the above synthesis is correct (personal bias acknowledged!) for profile 3 and also for profiles 1 and 2, the model (i) was adopted for interpretation purposes. Of the three profiles, profile 1 shows the largest change in the horizontal gravity gradient and this profile was selected for detailed interpretation (see p. 337).

Profiles 4, 5 and 6

These three profiles lie not only across Logan's Line but also across the entire width of St. Dominique slice. Their locations with respect to the regional gravity picture (see Fig. 60) shows that the profiles are on the west flank of the Green-Sutton Mountain gravity high (B in Fig. 60) and as such they do have horizontal gravity gradients that are positive eastwards. On the average, the gradients are about 0.8 mgals/km, 1.1 mgals/km and 1.2 mgals/km, respectively. All three

*Profiles 1 and 2 had to be stopped 2 to 3 km east of the inferred trace of Logan's Line because the roads along which these traverses were carried out changed direction. But it was possible to extend profile 3 for a distance of over 5 km east of Logan's Line.

profiles show an eastward increase in horizontal gradient just about where they cross the inferred trace of Logan's Line, but soon settle again at about their average gradients. The increase (in horizontal gradient) is barely noticeable on profile 6 and is most conspicuous on profile 4. This latter profile was selected for detailed analysis (see p.334). Over the St. Dominique slice, the profiles show small but noticeable gravity variations which are compatible with the slightly higher density of rocks which make up the St. Dominique slice. Between stations 170 and 176 on profile 4, gravity values fluctuate rather sharply. In this section several relatively thin rock formations (of Beekmantown, Chazy and Black River groups; see Clark 1964b) succeed one another from west to east. The sharp fluctuations of the gravity field probably reflect the density differences of these rock formations. Similar fluctuations, although expected on profiles 5 and 6, were not recorded.

Profile 7

Profile 7, like profiles 4, 5 and 6, is located on the west flank of the Green-Sutton Mountain gravity high. The gradients on the profile are on the average about 2 1/2 mgals/km, positive eastwards. This profile, unlike the profiles discussed earlier, shows no noticeable change where it crosses the inferred trace of Logan's Line. Perhaps the very strong regional makes it difficult to discern a residual anomaly that might be present.

Profile 8

This profile was compiled with a station spacing of about 122 m (400 ft) i.e. about half that of the other profiles. The profile shows sharp fluctuations of gravity values, superimposed on a general gradient of about 0.25 mgals/km, positive eastwards. The sharp fluctuations of values may reflect variations of overburden thickness along the profile. The profile does cross several small river valleys which may be filled with wedge-shaped masses of alluvium. It is not at all clear whether this profile contains a gravity effect that can be correlated with the inferred trace of Logan's Line.

Profiles 9 and 10

Profiles 9 and 10 are similar to profiles 1, 2 and 3 in that their negative horizontal gradients (0.4 mgals/km and 1.3 mgals/km respectively) flatten eastwards to near zero just about where they cross Logan's Line. In both profiles there is a hint that the gradient may become positive further eastwards. If this be the case, the gravity pattern indicated by the profiles is consistent with the regional gravity pattern (see Fig. 60). In the case of these profiles too (as in profiles 1, 2 and 3), it is not clear whether the eastward flattening of slopes is a residual or a regional effect or a combination of both. The profile 10 was extended for a distance of over 5 km east of Logan's Line but there is no indication on it that the gradient resumes its earlier values.

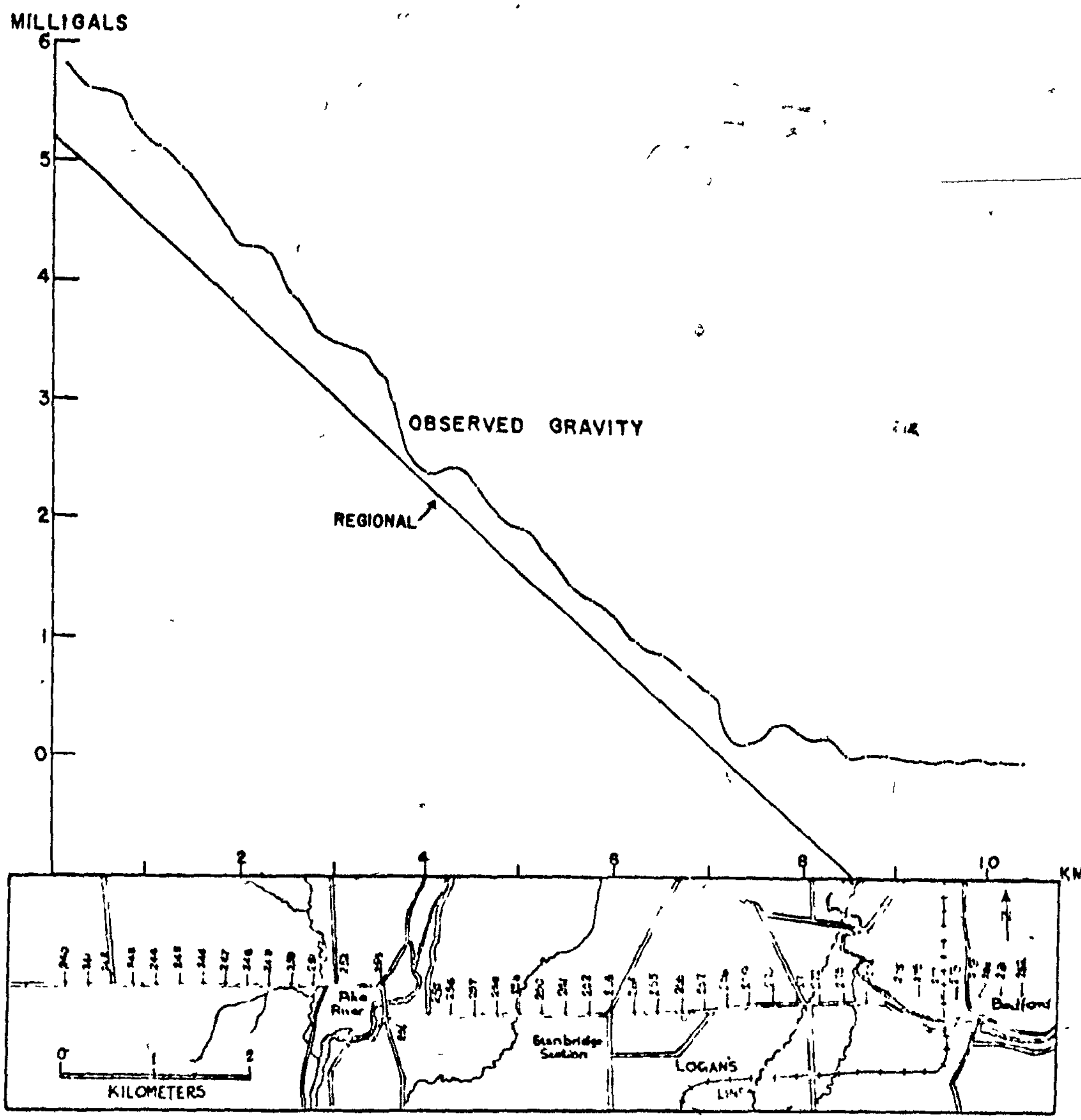

Fig.62.Bouguer gravity profile across Logan's Line. Traverse No.1.

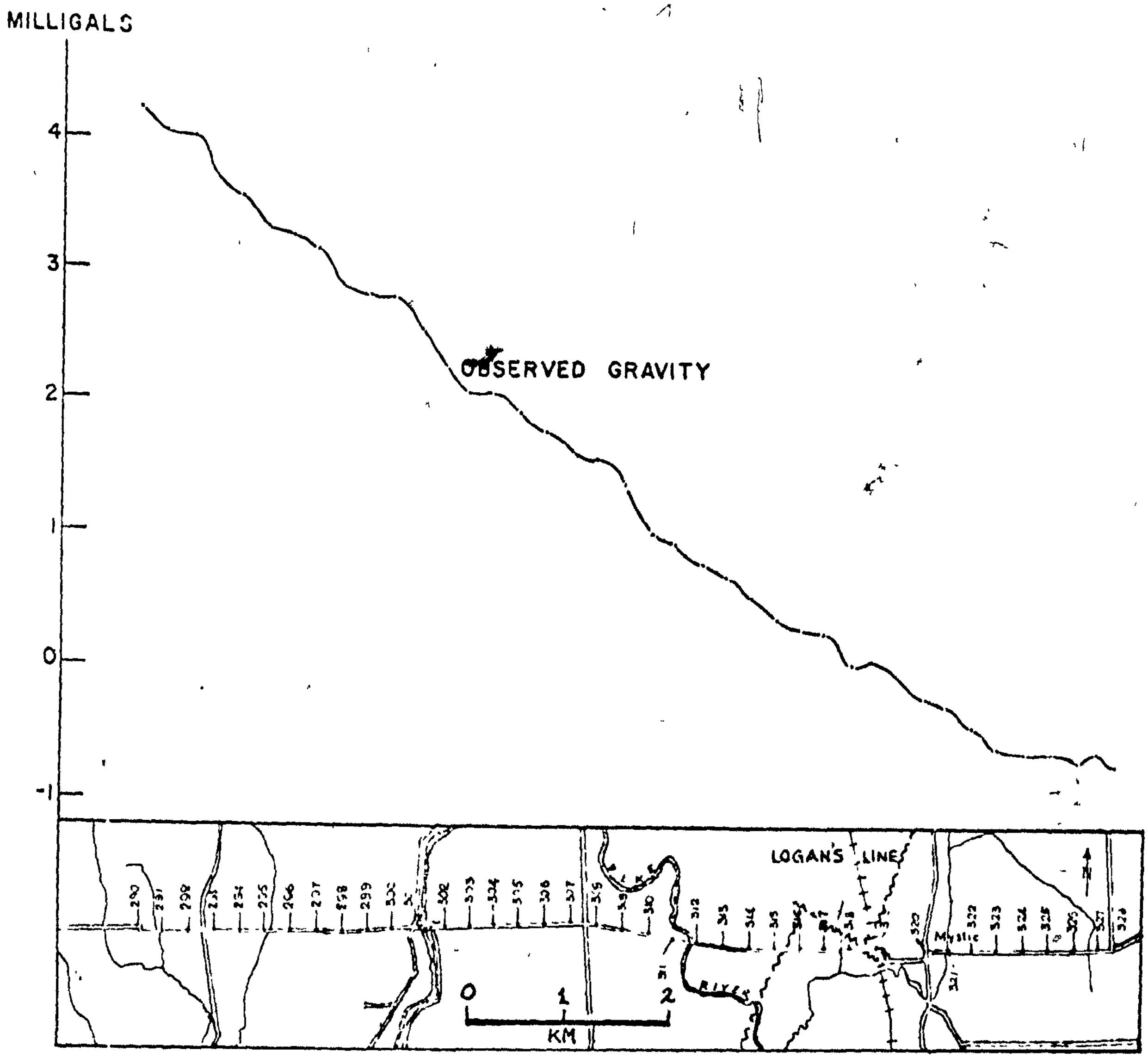

Fig.63. Bouguer gravity profile across Logan's Line. Traverse No. 2.

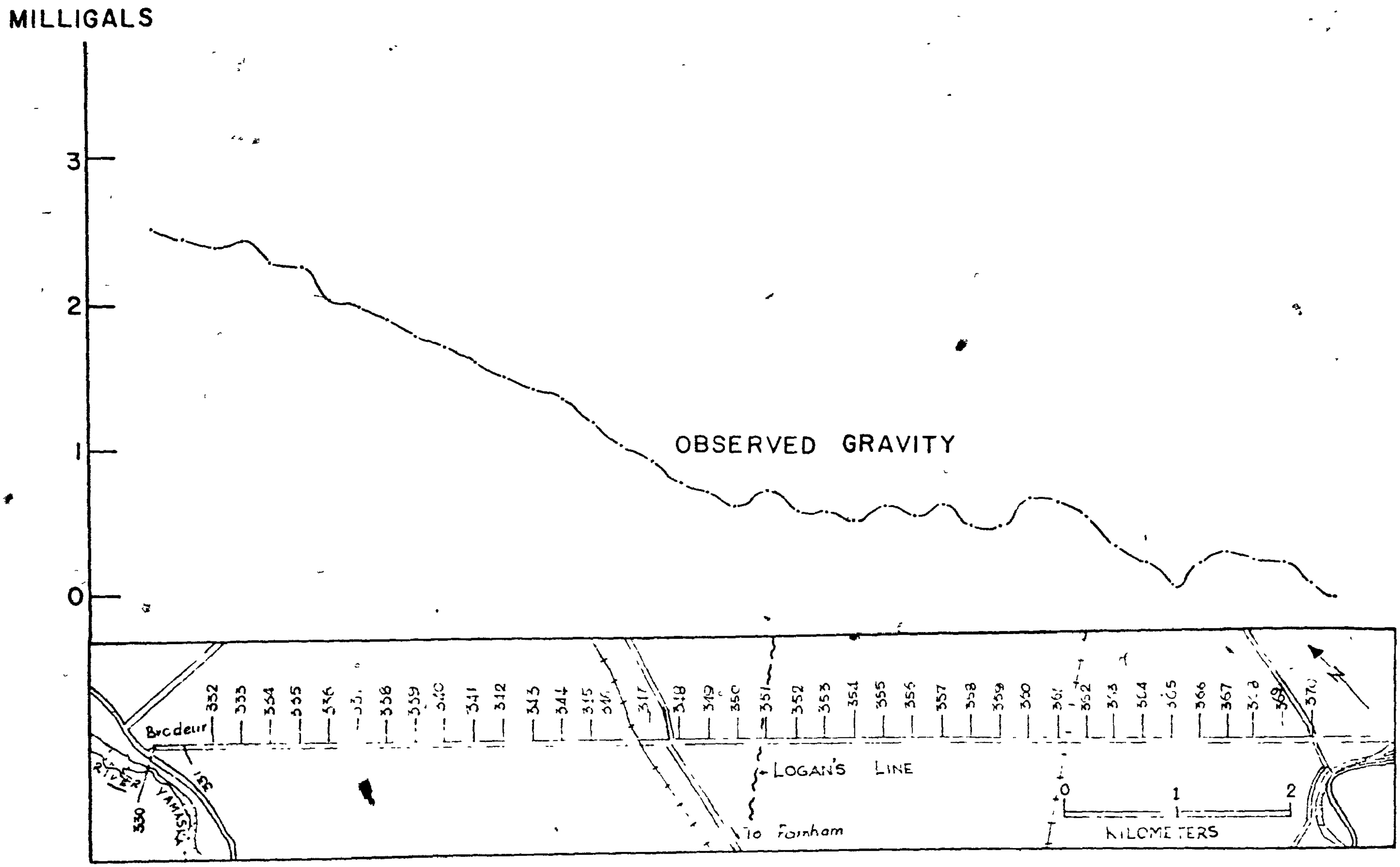

Fig.64. Bouguer gravity profile across Logan's Line. Traverse No.3.

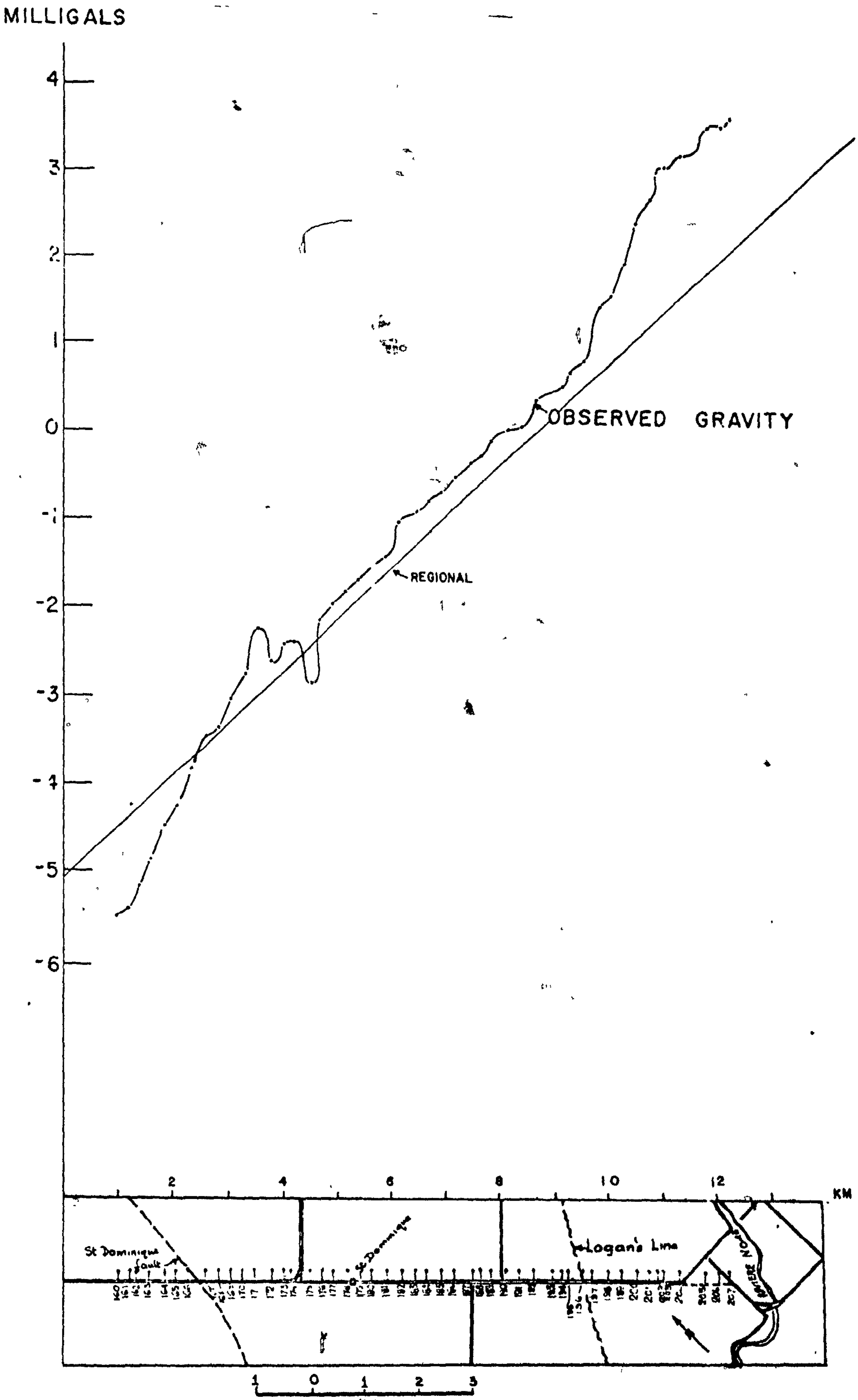

Fig.65.Bouguer gravity profile across Logan's Line. Traverse No.4.

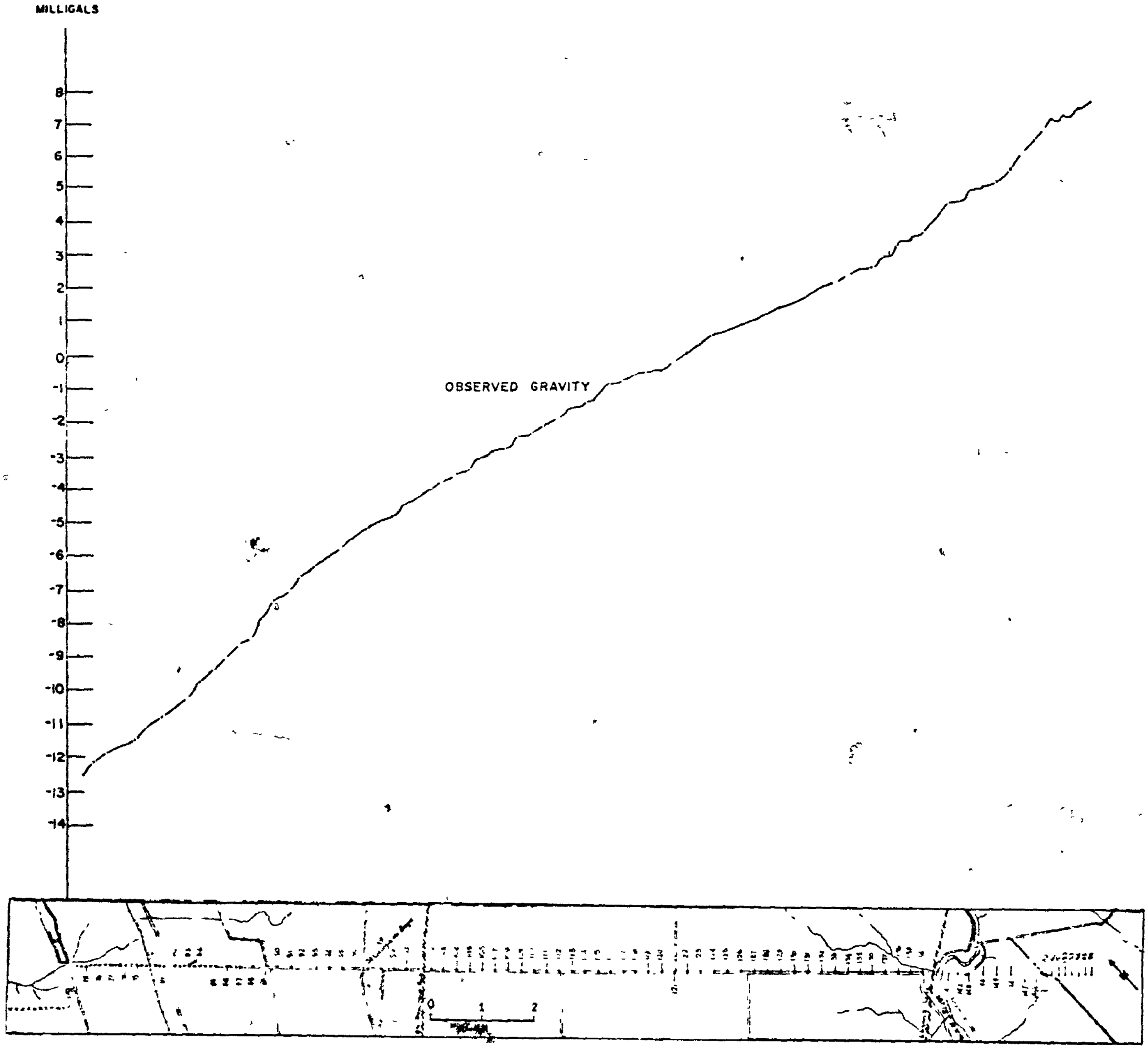

Fig.66.Bouguer gravity profile across Logan's Line. Traverse No.5.

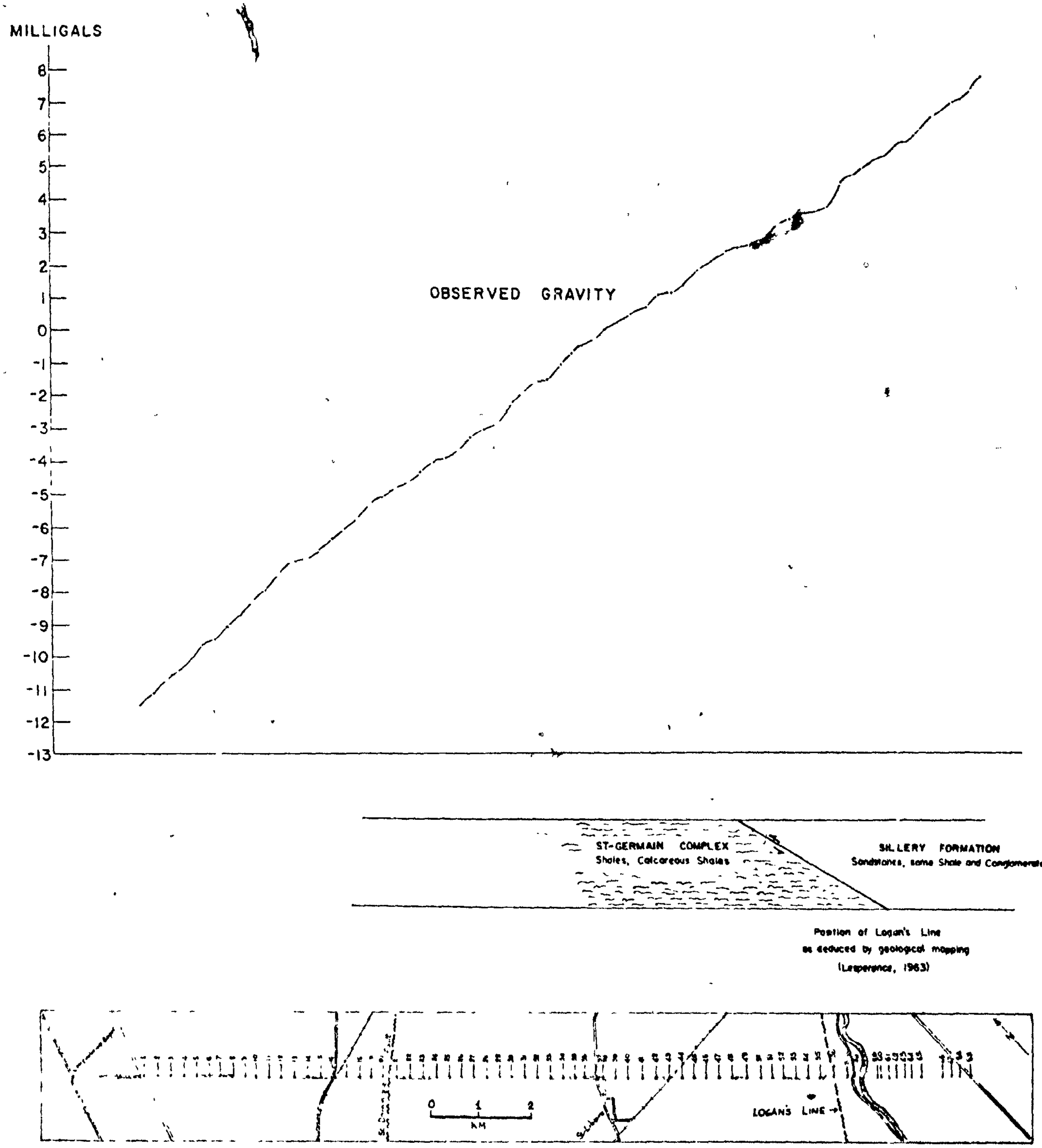

Fig.67. Bouguer gravity profile across Logan's Line. Traverse No.6.

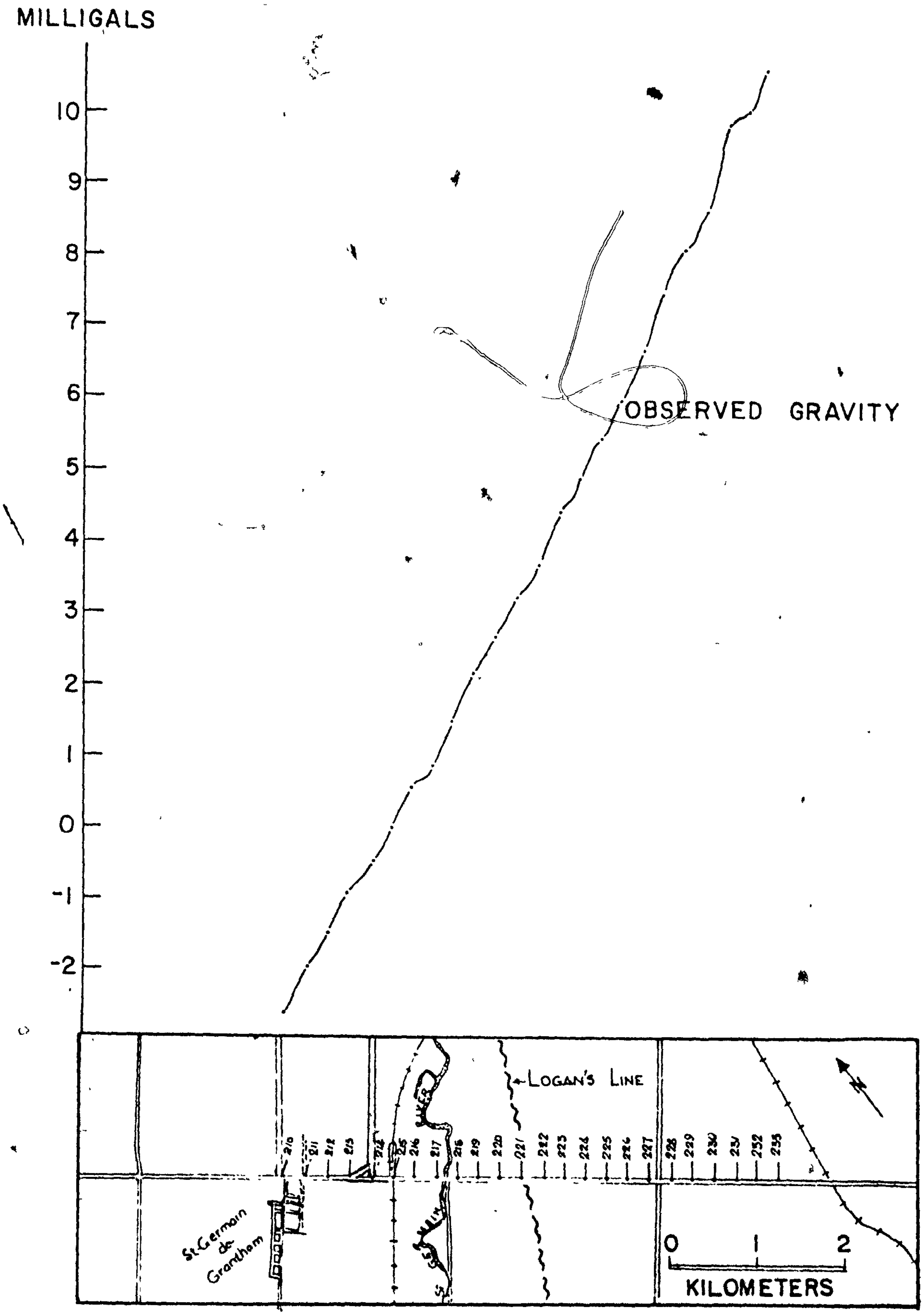

Fig.68. Bouguer gravity profile across Logan's Line. Traverse No.7.

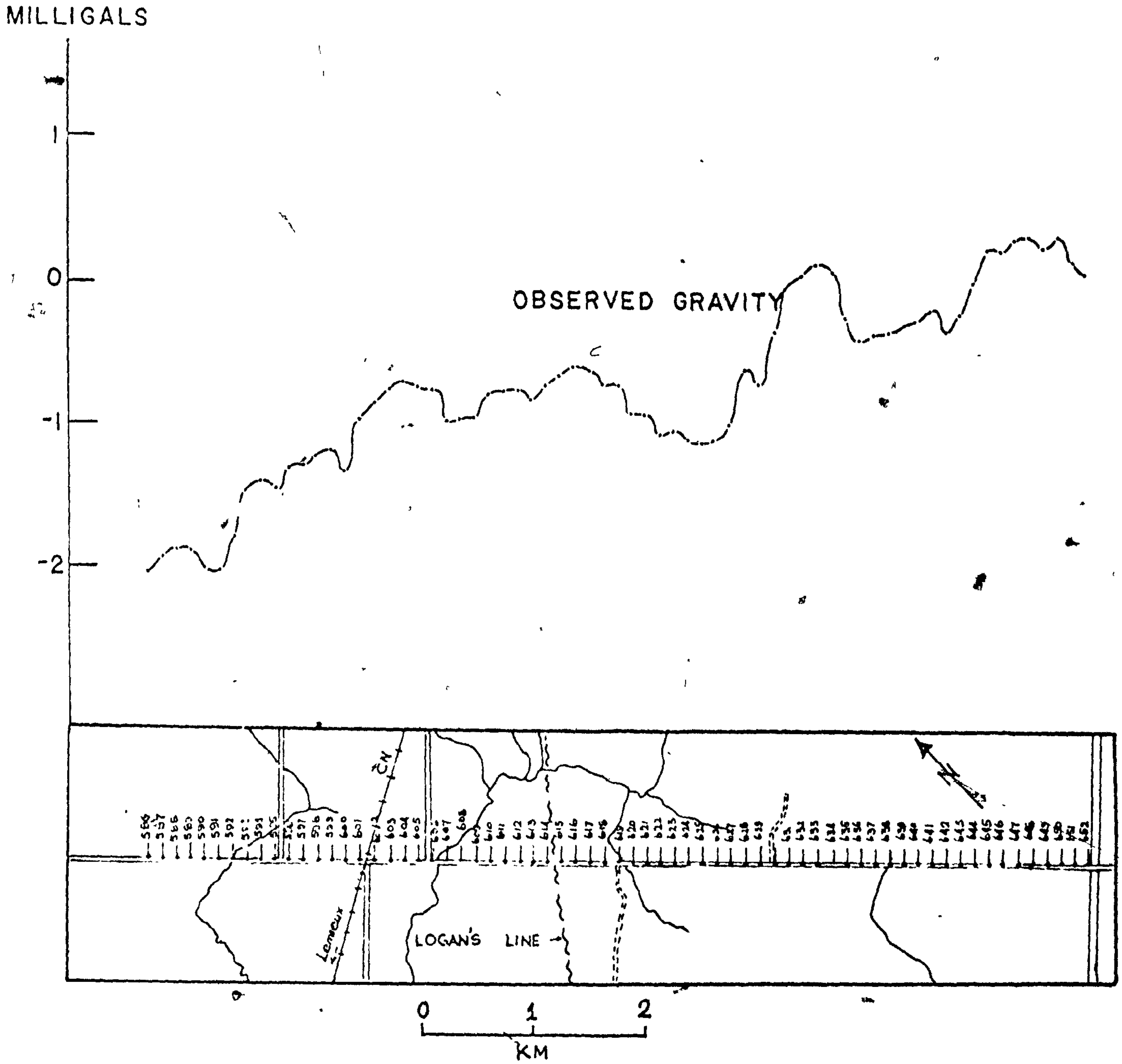

Fig.69. Bouguer gravity profile across Logan's Line. Traverse No.8.

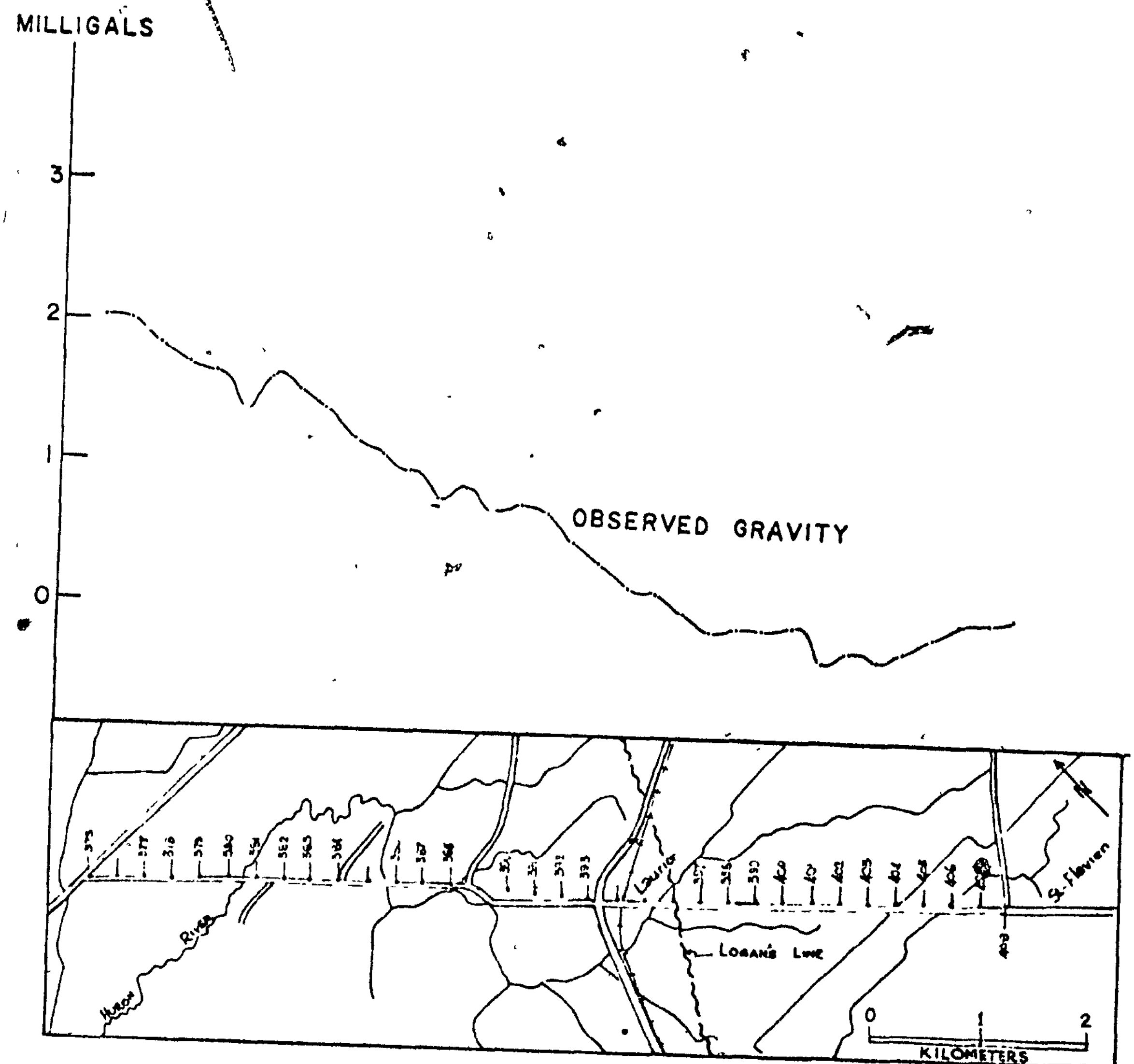

Fig.70.Bouguer gravity profile across Logan's Line. Traverse No.9.

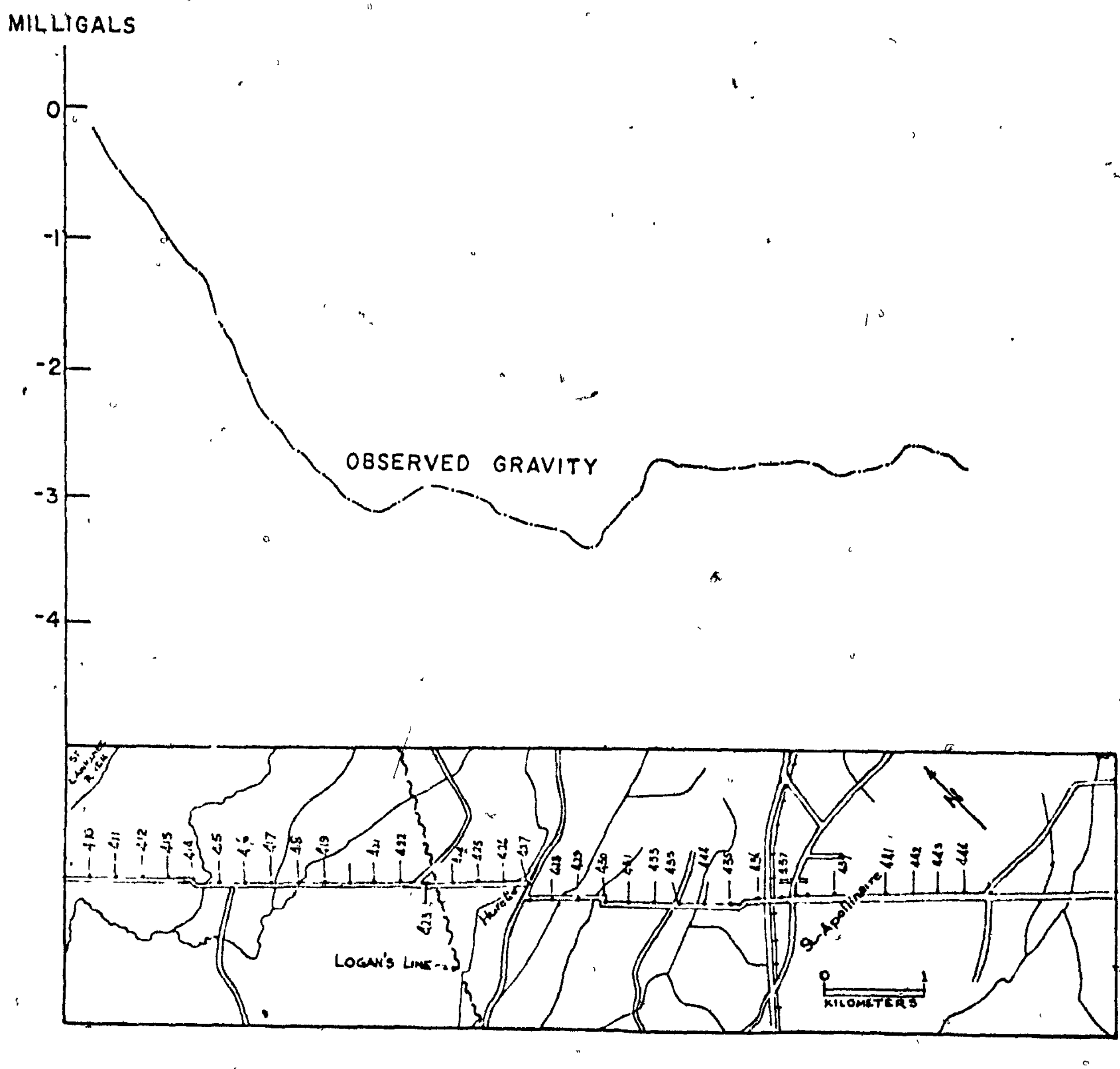

Fig.71. Bouguer gravity profile across Logan's Line. Traverse No.10.

INTERPRETATION

Profile 4

West of Logan's Line, profile 4 maintains a uniform horizontal gravity gradient of about 0.6 mgals/km for a distance of about 6 km. This linear segment of the profile was regarded as representing the regional trend. This trend was approximated by a straight line ($g = 0.566x - 5.071$, see Fig. 65) and removed from the observed gravity profile to give the residual anomaly shown in Fig. 72.

Interpretation of the residual anomaly was carried out by using a two-dimensional model of an east-dipping fault which can be approximated by a semi-infinite block of anomalous material of finite depth, truncated by a dipping plane. This semi-infinite block has the following parameters.

U_1, W_1 = horizontal and vertical positions of the point of intersection of the bottom of the block with the dipping plane.

U_2, W_2 = horizontal and vertical positions of the intersection of the top of the block with the dipping plane.

$\Delta\rho$ = the density contrast.

Fitting of theoretical curves was done by non-linear optimization (Al-Chalabi 1970; Hjett 1973) using a digital computer. The programming was done by Mr. A.K. Goodacre at the Earth Physics Branch, Ottawa and the computations were run on the computer at the Earth Physics Branch.

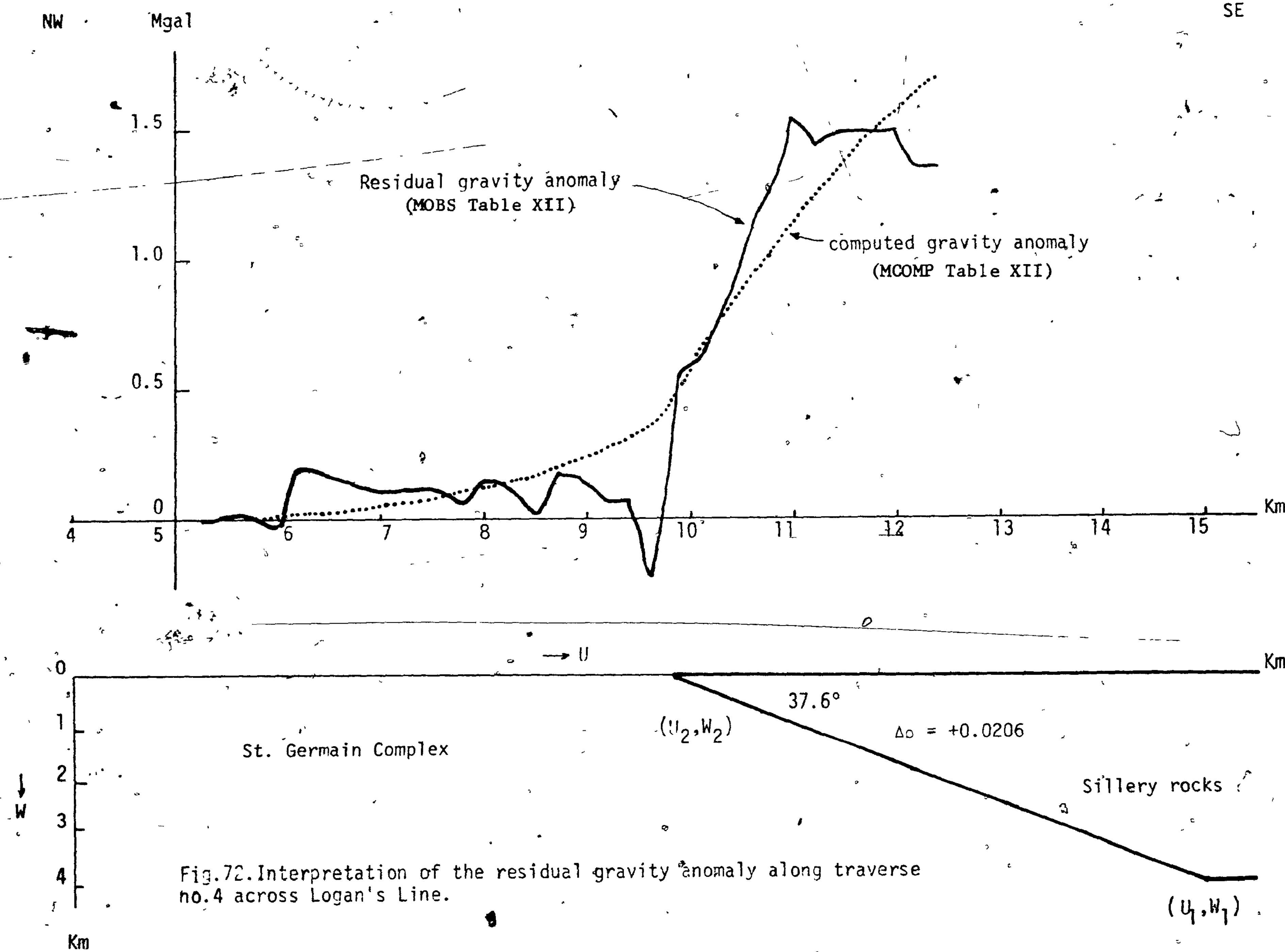

Fig.72. Interpretation of the residual gravity anomaly along traverse no.4 across Logan's Line.

In computing theoretical curves the positions of U_2 and W_2 were restricted to within 0.1 km of the location of Logan's Line, as given in geological maps of the area (see Clark 1964). The other parameters were unconstrained, although starting values were chosen by considering the amplitude of the anomaly and the density contrast as determined by density measurements. According to the values given in Table XI, the density contrast between the foot-wall and hanging-wall rocks is about 0.04 g/cm^3 and negative. A negative density contrast, however, cannot explain the residual anomaly. Therefore, $\Delta\rho$ was allowed to vary through small positive values. The root-mean-square of the deviations between the observed and theoretical anomaly curves (RMS in Table XII) was used as a measure of the best fit between the two curves. The minimum RMS value was 0.19480 and the corresponding parameters of the semi-infinite block are as follows (see also Table XII).

Thickness of the block = 4 km

Dip of the truncated edge (fault) = $\tan^{-1} \frac{W_1-W_2}{U_1-U_2}$ E = 37.6°E

$\Delta\rho$ = 0.0206 g/cm^3

The parameter which is difficult to reconcile with field measurements is the value of +0.0206 g/cm^3 for $\Delta\rho$, which as determined from density measurements on rock samples is −0.04 g/cm^3. This latter value as already discussed, may be on the low side, and as such it is possible that the density of the Sillery group may be as high as 2.68 g/cm^3.

It should be mentioned here that the gravity measurements on traverse 6 was used by Sharma (1968; also see Sharma and Geldart 1968) for analysis of the Bouguer gravity anomaly over Logan's Line. He used a ρ_B value of 2.0 g/cm^3 and found that the Bouguer anomaly consists of a trough and a peak simulating the gravity effect of a single faulted bed (Geldart et al. 1966, p. 378). However, when a ρ_B value of 2.65 g/cm^3 is used, the anomaly as computed by Sharma disappears. The latter value of ρ_B is more realistic (V.A. Saull, pers. comm.) in view of the density determinations as summarized in Table XI.

Profile 1

West of Logan's Line profile 1 maintains a uniform horizontal gravity gradient of about 0.75 mgals/km for a distance of about 7 km. This trend was regarded as representing the regional. It was approximated by a straight line ($g = -0.7350\ x\ +5.20$) and subtracted from the observed gravity to give the residual anomaly shown in Fig. 73.

Interpretation of the residual anomaly was carried out as with profile 4. For the best fit of the observed and theoretical profiles the parameters of the semi-infinite block is as follows (see also Table XIII).

Thickness of the block = 3 km

Dip of the truncated edge (fault) = 57.7°E

$\Delta\rho$ = +0.0217 g/cm^3

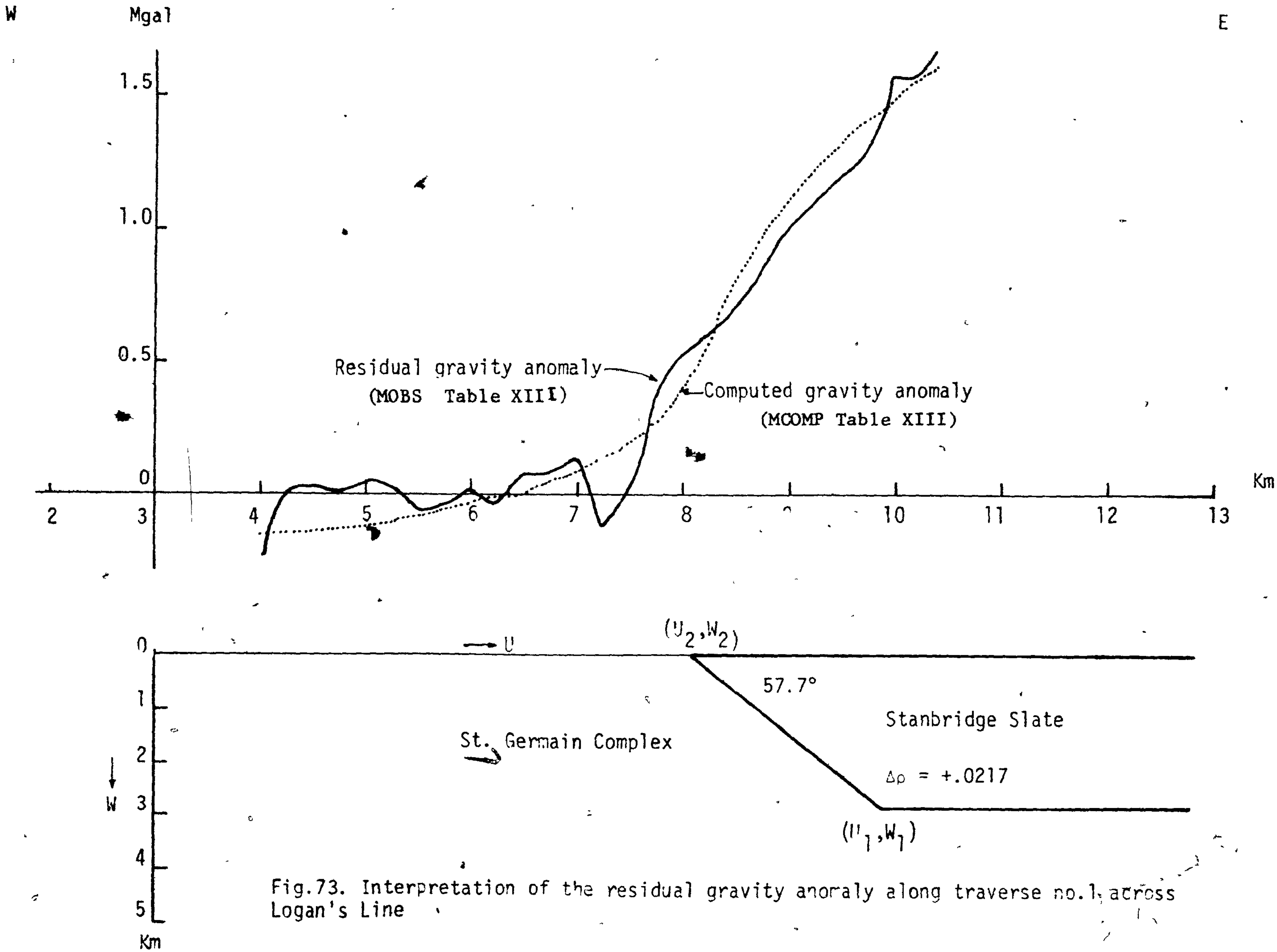

Fig.73. Interpretation of the residual gravity anomaly along traverse no.1 across Logan's Line

The dip angle is higher than the observed value of about 20° in the general area (Clark and McGerrigle 1944, p. 396) but falls within the range of dips (40° to 70°) of Logan's Line faults observed in Quebec City area.

CONCLUSIONS

The majority of the gravity profiles show rather sharp changes of the horizontal gravity gradient over Logan's Line. In most of the cases, however, it is far from certain as to what extent, if at all, the changes are caused by gravity effects related to Logan's Line. The changes may, to a smaller or larger extent, reflect variations of regional trends. However, with certain assumptions, interpretations consistent with the thrust fault hypothesis, can be given to observed gravity anomalies. If the assumptions are correct, the bulk density of the hanging wall rocks (Sillery, Stanbridge) should be about 0.02 g/cm^3 higher than the bulk density of St. Germain Complex.

TABLE VII

Densities of rock samples from the St. Germain Complex

Sample No.	Location	Description	SG in gm/cm^3
1	Long. 72°51', Lat. 45°39' approx. Orme brook, St. Hyacinthe area (Clark 1964b)	thinly bedded dark grey shale	2.67
2	" " "	thinly bedded dark grey shale with some grey sandstone	2.65
3	" " "	light grey sandstone	2.58
4	" " "	thinly bedded dark grey shale	2.69
5	" " "	" " " "	2.67
6	" " "	" " " "	2.66
7	Long. 72°52', Lat. 45°39' approx. Orme brook, St. Hyacinthe area (Clark 1964b)	" " " "	2.67
8	" " "	thinly bedded black shale	2.67
9	" " "	thinly bedded dark grey shale with some grey sandstone	2.59
10	" " "	" " " "	2.66
11	Long. 72°51½', Lat. 45°39' approx. Diamond Drill core from the Ste.-Rosalie hole no. 2 (Clark 1964b, p. 129, samples supplied by T.H. Clark)	thinly bedded grey shale with bands of light grey sandstone	2.57
12	" " "	thinly bedded dark grey shale	2.67

TABLE VII continued

Sample No.	Location	Description	SG in gm/cm^3
13	Long. 72°51½', Lat. 45°39' approx. Diamond Drill core from the Ste.-Rosalie hole no. 2 (Clark 1964b, p. 129, samples supplied by T.H. Clark)	thinly bedded dark grey shale	2.67
14	" " "	" " " "	2.66
15	" " "	" " " "	2.68
16	" " "	grey sandstone	2.56
17	Long. 72°36', Lat. 45°52' approx. Upton area (Clark 1964c)	black calcareous shale	2.68
18	" " "	" " "	2.69
19	" " "	" " "	2.67
20	" " "	black shaly limestone	2.72
21	" " "	black calcareous shale	2.67
22	" " "	" " "	2.68
23	" " "	" " "	2.68
24	" " "	" " "	2.68
25	" " "	" " "	2.67
26	" " "	" " "	2.68

TABLE VIII

Densities of rock samples from the St. Dominique Slice

Sample No.	Formation	Description	SG in gm/cm^3
31	Beldens Formation	light grey limestone	2.69
32	" "	medium grey, fine-grained dolomite	2.75
33	" "	light grey limestone	2.72
34	St. Dominique Formation	massive grey limestone	2.72
35	" " "	massive grey limestone	2.74
36	" " "	grey crystalline limestone	2.73
37	" " "	fine-grained calcareous sandstone	2.56
38	" " "	fine-grained sandstone	2.53
39	Formations of the Black River Group	dark, fine-grained fragmental limestone	2.71
40	" " "	dark, fine-grained limestone	2.72
41	" " "	dark grey, shaly limestone	2.67
42	Formations of the Trenton Group	dark grey limestone	2.74
43	" " "	black shaly limestone	2.71
44	" " "	black limestone	2.72
45	" " "	thinly bedded shaly limestone	2.69
46	St. Germain Complex	slaty rock	2.66

TABLE IX

Densities of rock samples from the Sillery Group

Sample No.	Location	Description	SG in gm/cm^3
51	Long. 72°46', Lat. 45°30½' approx. St. Hyacinthe area (Clark 1964b)	fine-grained greenish-grey siltstone	2.66
52	" " "	coarse-grained grey sandstone	2.56
53	" " "	medium-grained greenish-grey sub-greywacke	2.61
54	" " "	" " "	2.60
55	" " "	" " "	2.62
56	" " "	red slate	2.67
57	" " "	" "	2.67
58	" " "	red and green slate	2.68
59	" " "	red slate with minor bands of green	2.69
60	Long. 72°46', Lat. 45°32' approx. St. Hyacinthe area (Clark 1964b)	medium-grained greenish-grey sandstone	2.57
61	" " "	" " "	2.58
62	" " "	medium-grained greenish-grey sub-greywacke	2.61
63	" " "	" " "	2.60
64	" " "	red slate	2.66

TABLE IX continued

Sample No.	Location	Description	SG in gm/cm^3
65	Long. $72^{\circ}20\frac{1}{2}'$ to $72^{\circ}21\frac{1}{2}'$, Lat. $46^{\circ}00\frac{1}{2}'$ approx. Aston area (Clark 1964a)	medium-grained grey sandstone	2.59
66	" " "	" " "	2.57
67	" " "	medium-grained reddish-grey sub-greywacke	2.61
68	" " "	medium-grained grey sandstone	2.58
69	Long. $72^{\circ}20\frac{1}{2}'$ to $72^{\circ}21\frac{1}{2}'$, Lat. $46^{\circ}00\frac{1}{2}'$ approx.	medium-grained reddish-grey sub-greywacke	2.61
70	" " "	grey sandstone	2.57
71	" " "	grey quartzite	2.56
72	Along Nicolet River between Route 9 and St. Lawrence (Yamaska-Aston Map area (Clark 1964a)	red slate	2.66
73	" " "	" "	2.66
74	" " "	red and grey slate	2.68
75	" " "	grey sandstone	2.57
76	" " "	red crumbly shale	2.64
77	" " "	black slate	2.64
78	" " "	red slate	2.66
79	" " "	reddish grey sandstone	2.57

TABLE IX continued

Sample No.	Location	Description	SG in gm/cm^3
80	Along Nicolet River between Route 9 and St. Lawrence (Yamaska-Aston Map area (Clark 1964a)	red and grey slate	2.66
81	" " "	black slate	2.66
82	" " "	red and grey slate	2.67
83	" " "	medium-grained grey sandstone	2.58
84	" " "	medium-grained greenish-grey sandstone	2.57
85	" " "	coarse-grained grey sandstone	2.56
86	" " "	red and green slate	2.68
87	" " "	red slate	2.67
88	" " "	dark, almost black, slate	2.64
89	" " "	medium-grained grey sandstone	2.57
90	Long. 71°31', Lat. 46°37'. From road-crop, traverse no. 10	medium-grained grey sandstone	2.56
91	" " "	reddish-grey siltstone	2.63
92	" " "	medium-grained red brown sandstone	2.60
93	" " "	red slate	2.68
94	" " "	" "	2.63
95	" " "	green slate	2.65
96	" " "	red and green slate	2.65

TABLE X

Densities of rock samples from the Stanbridge Complex

Sample No.	Location	Description	SG in gm/cm^3
101	Long. 72°57½', Lat. 45°10' approx.	dark grey, cleaved shale (slate)	2.64
102	" " "	" " " "	2.63
103	" " "	" " " "	2.64
104	" " "	" " " "	2.62
105	" " "	" " " "	2.66
106	" " "	black shaly limestone	2.70
107	" " "	" " "	2.71
108	" " "	dark grey slate	2.67
109	" " "	black slate	2.66
110	Long. 72°58', Lat. 45°11' approx.	black slate	2.68
111	" " "	black limestone	2.71
112	" " "	black shaly limestone	2.70
113	" " "	black slate	2.64
114	" " "	dark grey slate	2.63
115	" " "	black shaly limestone	2.70
116	" " "	black slate	2.67

TABLE XI

Rock Densities: Summary

Classifications	Rock Type (s)	No. of Samples (N)	Mean Density (M)	Sample Standard Deviation (σ)	Standard error of mean ($\sigma/\sqrt{N}$)
St. Germain Complex	shales, calcareous shale, minor sandstone	26	2.66	.039	.008
St. Dominique Slice	limestone, sandstone, shale	16	2.69	.062	.015
Sillery Group	shale, sandstone	46	2.62	.044	.006
Stanbridge Complex	slate, minor limestone	16	2.67	.031	.008

TABLE XII. Computed data for semi-infinite block model, gravity profile 4 across Logan's Line

MOBS	MCOMP	RESID	XCOORD
-.00732	-.01760	.01028	5.20000
.00403	-.00994	.01397	5.42000
.00071	.00085	.00014	5.71000
-.03827	.01170	-.04997	5.98000
.20908	.02117	.18791	6.20000
.17610	.03513	.14097	6.50000
.14745	.04779	.09966	6.75000
.10480	.06153	.04327	7.00000
.11182	.07965	.03217	7.30000
.10517	.09636	.00881	7.55000
.06719	.11400	-.04682	7.79000
.15353	.12678	.02675	7.95000
.13488	.14864	-.01376	8.20000
.00457	.17752	-.17295	8.49000
.18691	.20260	-.01569	8.71000
.06195	.28164	-.21969	9.25000
.06796	.31052	-.24256	9.40000
-.22069	.36332	-.58401	9.62000
.57566	.51283	.06283	9.90000
.61534	.65112	-.03578	10.10000
.83369	.80195	.03174	10.35000
1.14904	.93888	.21016	10.60000
1.31739	1.06588	.25151	10.85000
1.53307	1.13343	.39964	10.99000
1.44008	1.23054	.20954	11.20000
1.48710	1.33594	.15116	11.44000
1.49113	1.55389	-.06276	11.98000
1.34848	1.63577	-.28729	12.20000
1.35483	1.70348	-.34865	12.39000

RMS
.19480

U1 U2 W1 W2 RHO REG

15.01823
9.81087
4.01810
.00356
.20620
-.29066

TABLE XIII. Computed data for semi-infinite block model, gravity profile 1 across Logan's Line

MOBS	MCOMP	RESID	XCOORD
-.22730	-.15546	-.07184	4.02000
.00215	-.14428	.14643	4.29000
.02420	-.13365	.15805	4.52000
.01365	-.12035	.13401	4.79000
.04875	-.10591	.15466	5.05000
.01750	-.09043	.10793	5.30000
-.06715	-.07601	.00886	5.51000
-.05715	-.06089	.00374	5.71000
.01565	-.03704	.05269	5.99000
-.05230	-.01461	-.03769	6.22000
.07050	.01699	.05351	6.50000
.06720	.04597	.02123	6.72000
.13165	.08803	.04362	6.99000
-.12165	.12928	-.25093	7.21000
.00710	.18688	-.17978	7.46000
.37185	.26192	.10993	7.71000
.52565	.38826	.13739	7.99000
.58600	.59176	-.00576	8.20000
.70380	.81805	-.11425	8.48000
.87385	.96519	-.09134	8.71000
.97620	1.08084	-.10464	8.92000
1.54790	1.48179	.06611	9.94000
1.21450	1.33554	-.12104	9.50000
1.25605	1.38236	-.12631	9.63000
1.42230	1.43251	-.01021	9.78000
1.57120	1.47586	.09534	9.92000
1.56260	1.54311	.01949	10.16000
1.65765	1.60007	.05758	10.39000

RMS
.10709

U1 U2 W1 W2 RHO REG

9.90982
8.09155
2.89256
.01215
.21680
-.37668

TABLE XIV

GRAVITY DATA LINE I

Station No.	Elevation (ft)	Gravity Value(mgals)
240	128.724	5.809
241	131.434	5.604
242	129.739	5.580
243	127.984	5.321
244	135.974	5.130
245	134.334	4.948
246	131.599	4.692
247	127.849	4.432
248	126.214	4.301
249	123.479	4.217
250	120.679	3.876
251	118.484	3.580
252	120.539	3.481
253	124.199	3.204
254	115.774	2.980
255	119.089	2.398
256	120.424	2.429
257	126.504	2.282
258	127.819	2.073
259	126.159	1.917
260	126.489	1.702
261	127.759	1.463
262	131.489	1.326
263	148.799	1.193
264	160.979	0.956

GRAVITY DATA LINE I

Station No.	Elevation(ft)	Gravity Value(mgals)
265	167.349	0.873
266	179.499	0.708
267	188.029	0.574
268	193.734	0.159
269	211.589	0.104
270	214.549	0.285
271	184.354	0.233
272	167.959	0.139
273	163.039	0.051
274	156.654	0.052
275	161.184	0.000
276	179.994	-0.178
277	174.384	-0.188
278	183.294	-0.242
279	173.529	-0.186
280	179.994	-0.140
281	190.119	-0.325
282	194.969	-0.399

GRAVITY DATA LINE 2

Station No.	Elevation(ft)	Gravity Value(mgals)
290	133.469	4.200
291	140.249	4.076
292	141.254	3.907
293	133.529	3.696
294	132.174	3.519
295	132.489	3.295
296	141.089	3.234
297	137.719	3.107
298	135.129	2.857
299	136.169	2.768
300	125.029	2.759
301	136.169	2.509
302	145.209	2.236
303	150.099	2.099
304	144.599	2.038
305	143.369	1.870
306	144.519	1.724
307	146.219	1.559
308	149.054	1.524
309	151.954	1.321
310	149.289	1.077
311	149.059	0.872
312	151.319	0.718
313	157.619	0.641
314	154.129	0.488

GRAVITY DATA LINE 2

Station No.	Elevation (ft)	Gravity Value (mgals)
315	169.954	0.306
316	162.494	0.269
317	160.679	0.237
318	168.519	-0.044
319	179.979	0.000
320	209.134	-0.294
321	199.854	-0.328
322	212.854	-0.484
323	229.214	-0.635
324	226.434	-0.675
325	225.814	-0.670
326	233.364	-0.733
327	224.314	-0.686
328	223.069	-0.766

GRAVITY DATA LINE 3

Station No.	Elevation (ft)	Gravity Value (mgals)
330	149.272	2.514
331	150.042	2.433
332	151.492	2.388
333	155.202	2.403
334	157.562	2.278
335	159.292	2.247
336	169.557	2.050
337	180.032	1.912
338	179.592	1.872
339	183.032	1.751
340	187.182	1.668
341	190.432	1.581
342	202.952	1.477
343	206.262	1.383
344	206.882	1.303
345	221.432	1.148
346	233.072	0.979
347	247.847	0.885
348	256.857	0.714
349	252.287	0.652
350	246.337	0.565
351	246.337	0.679
352	241.292	0.510
353	244.892	0.501
354	241.632	0.464

GRAVITY DATA LINE 3

Station No.	Elevation(ft)	Gravity Value(mgals)
355	240.747	0.544
356	245.007	0.491
357	244.727	0.561
358	248.737	0.412
359	247.932	0.410
360	244.232	0.592
361	234.652	0.585
362	226.107	0.479
363	222.567	0.289
364	220.677	0.165
365	220.627	0.068
366	220.077	0.157
367	221.027	0.221
368	226.007	0.189
369	227.857	0.173
370	223.462	0.000

TABLE XIV continued

GRAVITY DATA LINE 4

Station No.	Elevation(ft)	Gravity Value(mgal)
160	116.401	-5.503
161	116.231	-5.445
162	116.671	-5.172
163	118.301	-4.844
164	120.231	-4.475
165	121.77	-4.261
166	126.231	-3.847
167	137.626	-3.477
168	154.281	-3.357
169	164.361	-3.031
170	190.506	-2.755
171	229.901	-2.253
172	222.561	-2.617
173	228.161	-2.431
174	227.531	-2.400
175	233.681	-2.860
176	236.921	-2.146
177	228.756	-1.985
178	243.511	-1.842
179	251.381	-1.706
180	252.571	-1.545
181	244.776	-1.431
182	235.221	-1.059
183	232.426	-0.922
184	239.311	-0.809

TABLE XIV CONTINUED ON PAGE 358

TABLE XIV continued

GRAVITY DATA LINE 4

Station No.	Elevation(ft)	Gravity Value(mgals)
185	248.506	-0.710
186	260.096	-0.533
187	267.386	-0.398
188	269.476	-0.300
189	265.256	-0.123
190	264.906	0.000
191	264.496	0.034
192	248.426	0.341
193	243.156	0.522
194	236.781	0.613
195	214.441	0.449
196	196.856	1.210
197	190.236	1.404
198	184.586	1.557
199	178.171	1.917
200	172.711	2.374
201	167.051	2.684
202	160.651	2.979
203	164.191	3.005
204	162.351	3.188
205	156.801	3.498
206	159.491	3.480
207	157.501	4.594

TABLE XIV continued

GRAVITY DATA LINE 5

Station No.	Elevation(ft)	Gravity Value(mgals)
75	111.102	-12.480
76	111.592	-12.160
77	112.322	-11.847
78	113.232	-11.658
79	113.822	-11.504
80	114.512	-11.082
81	115.182	-10.802
82	115.282	-10.259
83	115.692	- 9.798
84	116.632	- 9.422
85	117.872	- 9.048
86	118.772	- 8.653
87	119.972	- 8.415
88	121.152	- 7.812
89	124.602	- 7.362
90	144.172	- 7.037
91	152.602	- 6.605
92	158.062	- 6.350
93	166.387	- 6.077
94	171.647	- 5.819
95	178.577	- 5.478
96	188.817	- 5.254
97	198.872	- 5.014
98	204.352	- 4.876
99	199.067	- 4.542

TABLE XIV continued

GRAVITY DATA LINE 5

Station No.	Elevation(ft)	Gravity Value(mgals)
100	197.332	-4.343
101	198.512	-4.072
102	201.212	-3.832
103	204.622	-3.645
104	205.772	-3.450
105	210.652	-3.139
106	215.912	-2.934
107	213.532	-2.839
108	215.732	-2.504
109	218.932	-2[illegible]
110	220.912	-2.167
111	228.702	-1.982
112	230.522	-1.685
113	240.982	-1.570
114	240.137	-1.328
115	240.102	-1.121
116	240.072	-0.986
117	238.652	-0.803
118	238.142	-0.662
119	241.687	-0.495
120	240.517	-0.268
121	238.267	0.000
122	242.567	0.230
123	250.992	0.495

TABLE XIV continued

GRAVITY DATA LINE 5

Station No.	Elevation(ft)	Gravity Value(mgals)
124	266.782	0.633
125	273.967	0.800
126	273.297	0.948
127	276.142	1.141
128	282.822	1.312
129	281.212	1.413
130	276.412	1.614
131	266.667	1.891
132	257.972	2.026
133	253.627	2.262
134	228.167	2.452
135	215.762	2.538
136	208.792	2.835
137	204.107	3.233
138	194.987	3.417
139	188.817	3.601
140	185.737	3.798
141	182.066	4.489
142	190.496	4.553
143	189.656	4.814
144	188.656	4.991
145	191.656	5.151
146	191.156	5.480
147	193.576	6.004
148	197.006	6.541

GRAVITY DATA LINE 5

Station No.	Elevation(ft)	Gravity Value(mgals)
149	202.466	6.704
150	207.236	7.021
151	205.816	6.980
152	205.156	7.180
153	205.471	7.138
154	206.701	7.334
155	208.351	7.391
156	211.086	7.584

TABLE XIV continued

GRAVITY DATA LINE 6

Station No.	Elevation(ft)	Gravity Value(mgals)
1	126.975	-11.503
2	126.475	-11.084
3	129.015	-10.718
4	139.480	-10.408
5	151.335	-10.015
6	166.400	- 9.764
7	175.215	- 9.463
8	174.545	- 9.022
9	174.665	- 8.708
10	181.090	- 8.328
11	189.900	- 7.949
12	193.545	- 7.465
13	197.240	- 7.131
14	210.830	- 6.981
15	214.350	- 6.647
16	220.730	- 6.349
17	229.970	- 6.059
18	228.010	- 5.806
19	231.030	- 5.358
20	232.740	- 5.107
21	231.690	- 4.830
22	232.360	- 4.662
23	234.100	- 4.380
24	238.910	- 4.031
25	245.120	- 3.965

GRAVITY DATA LINE 6

Station No.	Elevation(ft)	Gravity Value(mgals)
26	255.720	-3.679
27	257.740	-3.271
28	258.050	-3.031
29	260.760	-2.826
30	253.750	-2.380
31	252.940	-2.059
32	251.910	-1.735
33	252.980	-1.473
34	255.910	-1.078
35	263.970	-0.722
36	261.490	-0.474
37	284.650	-0.319
38	280.770	0.000
39	283.010	0.202
40	287.630	0.553
41	280.036	0.711
42	270.620	1.133
43	284.230	1.087
44	273.420	1.463
45	251.920	1.884
46	247.330	2.127
47	243.130	2.259
48	230.420	2.528
49	229.160	2.596
50	228.956	2.842

TABLE XIV continued

GRAVITY DATA LINE 6

Station No.	Elevation(ft)	Gravity Value(mgals)
51	206.546	3.189
52	208.156	3.281
53	213.336	3.485
54	223.696	3.503
55	211.190	3.730
56	209.956	4.281
57	199.281	4.604
58	186.241	4.811
59	187.242	5.295
60	186.822	5.458
61	189.302	5.675
62	193.342	5.770
63	198.232	5.891
64	200.272	6.209
65	205.292	6.506
66	219.602	7.078
67	226.067	7.157
68	223.987	7.509
69	226.212	7.799

TABLE XIV continued

GRAVITY DATA LINE 7

Station No.	Elevation(ft)	Gravity Value(mgals)
210	290.845	-2.614
211	280.325	-1.966
212	282.925	-1.510
213	279.005	-0.958
214	283.195	-0.536
215	260.00	0.000
216	253.395	0.584
217	255.435	0.885
218	254.480	1.442
219	252.485	2.179
220	251.685	2.646
221	252.485	3.203
222	253.005	3.721
223	253.085	4.398
224	254.615	4.887
225	256.370	5.461
226	258.100	6.011
227	260.060	6.673
228	260.040	7.472
229	262.655	8.161
230	264.635	8.630
231	266.655	9.669
232	272.205	9.979
233	273.365	10.526

GRAVITY DATA LINE 8

Station No.	Elevation(ft)	Gravity Value(mgals)
586	309.010	-2.037
587	313.00	-1.925
588	316.520	-1.890
589	313.160	-1.882
590	307.170	-2.009
591	303.440	-2.034
592	303.915	-1.729
593	303.575	-1.475
594	307.050	-1.415
595	309.170	-1.468
596	309.110	-1.326
597	306.580	-1.306
598	307.110	-1.242
599	309.330	-1.204
600	314.215	-1.365
601	316.515	-0.997
602	316.065	-0.888
603	310.280	-0.792
604	306.345	-0.743
605	301.525	-0.762
606	299.265	-0.793
607	301.325	-1.055
608	298.900	-1.000
609	299.595	-0.979
610	299.945	-0.801

GRAVITY DATA LINE 8

Station No.	Elevation(ft)	Gravity Value(mgals)
611	300.675	-0.802
612	301.570	-0.802
613	302.285	-0.868
614	302.670	-0.782
615	303.575	-0.706
616	304.200	-0.636
617	305.140	-0.669
618	307.785	-0.769
619	309.945	-0.783
620	307.235	-0.967
621	306.180	-0.963
622	306.340	-1.111
623	306.830	-1.101
624	307.270	-1.159
625	307.285	-1.176
626	306.945	-1.165
627	306.945	-1.048
628	307.330	-0.671
629	307.485	-0.763
630	308.650	-0.412
631	310.415	-0.119
632	317.395	-0.012
633	317.165	0.073
634	313.195	0.000
635	310.895	-0.319

GRAVITY DATA LINE 8

Station No.	Elevation(ft)	Gravity Value(mgals)
636	310.485	-0.483
637	310.925	-0.432
638	311.190	-0.417
639	312.485	-0.382
640	311.645	-0.332
641	315.105	-0.270
642	315.555	-0.408
643	316.780	-0.294
644	317.300	-0.070
645	325.170	0.177
646	336.610	0.159
647	336.640	0.236
648	328.385	0.253
649	319.665	0.176
650	317.740	0.256
651	313.905	0.084
652	313.840	-0.054

TABLE XIV continued

GRAVITY DATA LINE 9

Station No.	Elevation(ft)	Gravity Value(mgals)
375	292.410	2.096
376	295.335	1.982
377	298.265	1.803
378	302.535	1.675
379	308.715	1.630
380	307.640	1.360
381	316.275	1.593
382	319.270	1.502
383	332.775	1.370
384	339.880	1.149
385	337.550	1.051
386	338.610	0.944
387	343.680	0.733
388	347.375	0.840
389	354.795	0.695
390	354.080	0.711
391	357.130	0.651
392	364.290	0.481
393	361.760	0.314
394	371.72	0.137
395	370.060	0.113
396	371.590	-0.009
397	377.715	-0.150
398	379.830	-0.131
399	381.310	-0.126

TABLE XIV continued

GRAVITY DATA LINE 9

Station No.	Elevation(ft)	Gravity Value(mgals)
400	390.385	-0.109
401	404.300	-0.354
402	400.240	-0.283
403	401.530	-0.323
404	402.340	-0.266
405	403.310	-0.162
406	410.545	-0.083
407	416.485	-0.043
408	424.690	0.000

TABLE XIV continued

GRAVITY DATA LINE 10

Station No.	Elevation(ft)	Gravity Value(mgals)
410	192.400	-0.193
411	205.115	-0.504
412	217.640	-0.728
413	223.010	-1.047
414	223.970	-1.264
415	230.970	-1.703
416	238.605	-2.070
417	248.805	-2.420
418	260.700	-2.648
419	266.545	-2.839
420	273.235	-3.028
421	272.685	-3.127
422	275.270	-3.046
423	278.790	-2.920
424	282.875	-2.964
425	285.675	-3.013
426	284.445	-3.194
427	280.490	-3.221
428	275.650	-3.294
429	270.980	-3.313
430	292.190	-3.253
431	294.675	-3.030
432	314.380	-2.734
433	312.905	-2.762
434	310.990	-2.861

TABLE XIV continued

GRAVITY DATA LINE 10

Station No.	Elevation(ft)	Gravity Value(mgals)
435	310.760	-2.899
436	312.105	-2.774
437	314.745	-2.759
438	350.490	-2.411
439	364.545	-2.849
440	370.490	-2.811
441	380.000	-2.760
442	392.835	-2.617
443	397.715	-2.670
444	403.720	-2.806

APPENDIX IV

A GRAVITY PROFILE ACROSS THE SHIELD MARGIN IN THE VICINITY OF ST. JEROME, QUEBEC

(Also see Kumarapeli and Sharma 1969)

INTRODUCTION

In southern Quebec, the margin of the Canadian Shield trends approximately northeast, and is largely outlined by a system of en echelon normal faults (Fig. 74). The downthrown side of these faults is to the southeast and in that direction the shield surface dips under the sedimentary cover of the St. Lawrence Platform. This cover consists of nearly flat-lying Lower Paleozoic rocks, some 1500 to 3000 m thick, and there is reason to believe that it may once have covered much of the shield surface. The exhumation of the shield surface, in the immediate vicinity of the present Precambrian-Paleozoic boundary, must have taken place in recent times.

The basal member of the platform succession is the Upper Cambrian Potsdam formation, which for the most part is a crossbedded sandstone. It outcrops next to the shield margin a few kilometers southwest of St. Jérome (Fig. 74). Near St. Jérome, however, the platformal rocks that abut against the shield margin are not Potsdam, but belong to the Lower Ordovician Beekmantown group, whose normal stratigraphic position is immediately above the Potsdam. It is very likely that the Potsdam formation in the St. Jérome area is faulted out,

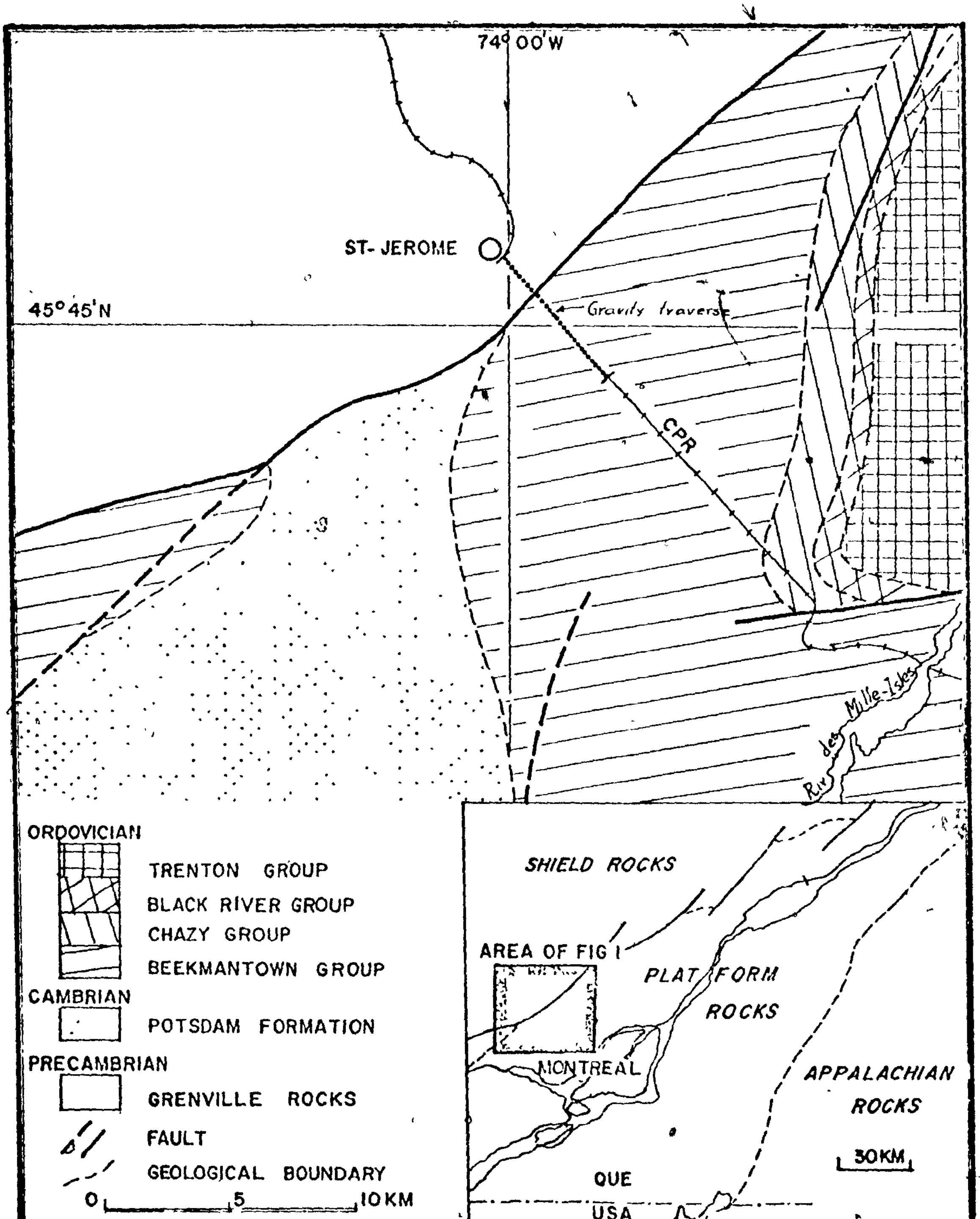

Fig.74. Geologic map of St.Jérome area. After Houde and Clark 1961.

but the possibility that it was never deposited there has not been ruled out (Osborne and Clark 1960, p. 20). The gravity profile discussed below was compiled in order to test the above possibility and to estimate the thickness of the Potsdam formation at the shield margin if the Potsdam is indeed faulted out.

GENERAL GEOLOGY

The geology of the area of the gravity traverse has been described by Osborne and Clark (1960). Precambrian rocks consist mainly of a diverse group of gneisses that vary from granitic to gabbroic in composition. Their foliation direction is commonly north-south. The Beekmantown group rocks are composed mainly of dolomite. Their dips are to the southeast and average between 2^{o} and 3^{o}. The thickness of the Beekmantown group in the area is believed to be around 900 ft (275 m) (Osborne and Clark 1960, p. 20).

To a large extent, bedrock along the gravity traverse is mantled by a veneer of glacial deposits composed mainly of clay, sand, and gravel. Precambrian rocks outcrop only at the north end of the traverse. There they consist of mafic schistose rocks that are generally referred to as mangerite.

FIELD DATA COLLECTION

The gravity traverse was made along the Canadian Pacific railway line and was 6.76 km long (Figs. 74, 75). The distance between stations was approximately 250 m. The gravity measurements were done with a Worden gravimeter and the relative elevations of stations were determined by precise levelling. The total number of stations was 28. The gravity readings were tied to a floating base established on the traverse.

FIELD DATA REDUCTION

Along the traverse variations of the thickness of overburden are not known with any certainty. The maximum thickness is probably no more than 10 m. In reducing gravity data, no attempt was made to correct for the possible variation of overburden.

The grades along the profile are low and there are no sharp upstanding masses in its vicinity. Therefore, no topographic corrections were deemed necessary.

The material constituting the overburden in the area has an average density of about 2 g/cm^3. This value was originally used as ρ_B in reducing gravity field data (1972 submission of this thesis; Kumarapeli and Sharma 1969). However, the densities of the various rock types that underlie the area, are appreciably higher (V.A. Saull, pers. comm.). The ρ_B value used to calculate the Bouguer gravity values given in Table XV is 2.67 which is the average density of granitic rocks. It is thought that the latter value of ρ_B is a good approximation of the mean density of the rocks in the area.

DENSITIES USED

Ten Precambrian rock samples collected from the outcrop at the north end of the profile gave an average density of 2.82 g/cm^3. This value was used as the density of Precambrian rocks along the profile. The average density values for Potsdam sandstone and Beekmantown dolomite were adopted from measurements made by earlier workers: 2.52 g/cm^3 for Potsdam (Saxov 1956) and 2.71 g/cm^3 for Beekmantown (McDonald 1965).

THE GRAVITY ANOMALY

Figure 75 shows the Bouguer anomaly profile along the traverse after making free-air, Bouguer, and latitude corrections. To separate this anomaly profile into its probable regional and residual components, the traditional method of visual smoothing was used. The gravity profile is steepest over the area of the Precambrian-Paleozoic boundary, and away from this contact zone the profile flattens and seems to settle at a gradient of about 0.30 mgal/km (Fig. 75). This gradient was assumed as the slope of the regional anomaly profile along the traverse (Fig. 75). It is possible that the actual gradient is flatter than the one assumed above, but a steeper gradient seems to be unlikely. The residual anomaly profile obtained by removing the regional as estimated above is shown in Fig. 75. It is a typical gravity profile across a faulted bed or a flat step of anomalous material.

DISCUSSION

For the purpose of mathematical analysis, a faulted bed or a flat step of anomalous material can conveniently be approximated by

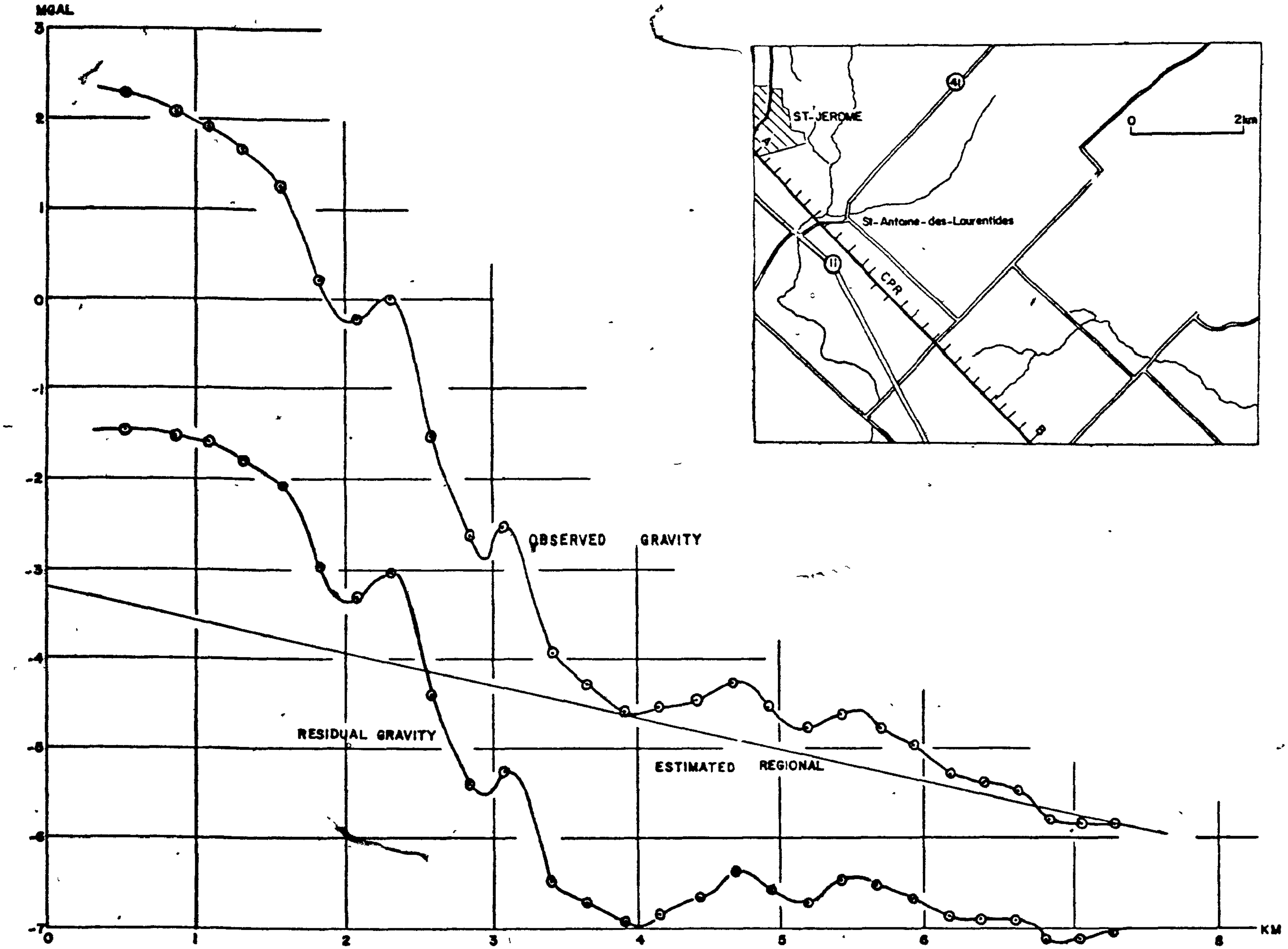

Fig.75. Bouguer gravity profile across the Shield margin in St.Jérome area. Traverse No. 12.

a semi-infinite block. Geldart et al. (1966) have shown that the total change in gravity due to a semi-infinite block truncated by a dipping plane (representing a fault) at an arbitrary angle of inclination is $2\pi G\rho t$, where G is the universal gravitational constant, ρ the density contrast of the block from the surrounding medium, and t the thickness of the semi-infinite block. From Fig. 75, the observed change in gravity is about 5.5 mgal. (This value is the same as the amplitude of the residual anomaly obtained using a ρ_B of 2 g/cm^3). If this anomaly is entirely due to the density difference of about 0.11 g/cm^3 between Precambrian and Beekmantown rocks, then the total thickness of the Beekmantown group has to be about 4000 ft (1220 m). But the total thickness of the Beekmantown group is known to be much smaller, being of the order of 900 ft (275 m) or so (Osborne and Clark 1960, p. 20). Hence, the large change in gravity cannot be explained by the density difference between the Precambrian and the Beekmantown alone.

The presence of a layer of low density Potsdam rock (average density 2.52 g/cm^3) below the Beekmantown can, however, account for the large change in the observed gravity. As shown in Fig. 76, if a layer of Potsdam is present below the Beekmantown, then assuming each layer as a semi-infinite block, it can be shown that the total change in gravity due to the combined effect of the two blocks is $2\pi G(\rho_1 t_1 - \rho_2 t_2)$, where ρ_1 refers to the density contrast between Precambrian and Beekmantown, ρ_2 the density contrast between Precambrian and Potsdam, t_1 and t_2 thicknesses of Beekmantown and Potsdam respectively. The measured value of ρ_2 is 0.32 g/cm^3, and substituting this value of ρ_2 in the expression for the total change in gravity due to two semi-infinite blocks and taking ρ_1 = 0.11 and t_1 = 900 ft (275 m), the value for the thickness of the Potsdam Formation comes to be about 1050 ft (320 m). This value probably represents the minimum thickness of Potsdam present, for if the actual regional gravity gradient is less steep than the one assumed in the regional-residual analysis, then the residual anomaly will have an amplitude greater than 5.5 mgal and the thickness of the Potsdam Formation will come out correspondingly greater. In this connection it is interesting to note that a drill hole located about 15 km southeast of St. Jérome passed through 1696 ft (517 m) of Potsdam without reaching the Grenville basement (Clark 1952, p. 17).

The equation given by Geldart et al. (1966) was used to compute the combined gravity anomaly of two semi-infinite blocks representing the Beekmantown Group and the Potsdam Formation. For the calculations, the dip of the fault plane is taken as 60^0, t_1 = 900 ft (275 m), t_2 = 1050 ft (320 m), ρ_1 = 0.11 g/cm^3, ρ_2 = 0.32 g/cm^3, and the overburden is taken to be 20 ft (6 m) thick, having a density of 2.00 g/cm^3. The agreement between the theoretical and the residual curves (Fig. 76) is good except for small irregularities in the residual curve. Calculations were also made with different values of t_1 and t_2, dip of the fault plane, and with slightly different values of ρ_1 and ρ_2, but the computed curve of Fig. 76 seems to give the best fit with the observed data.

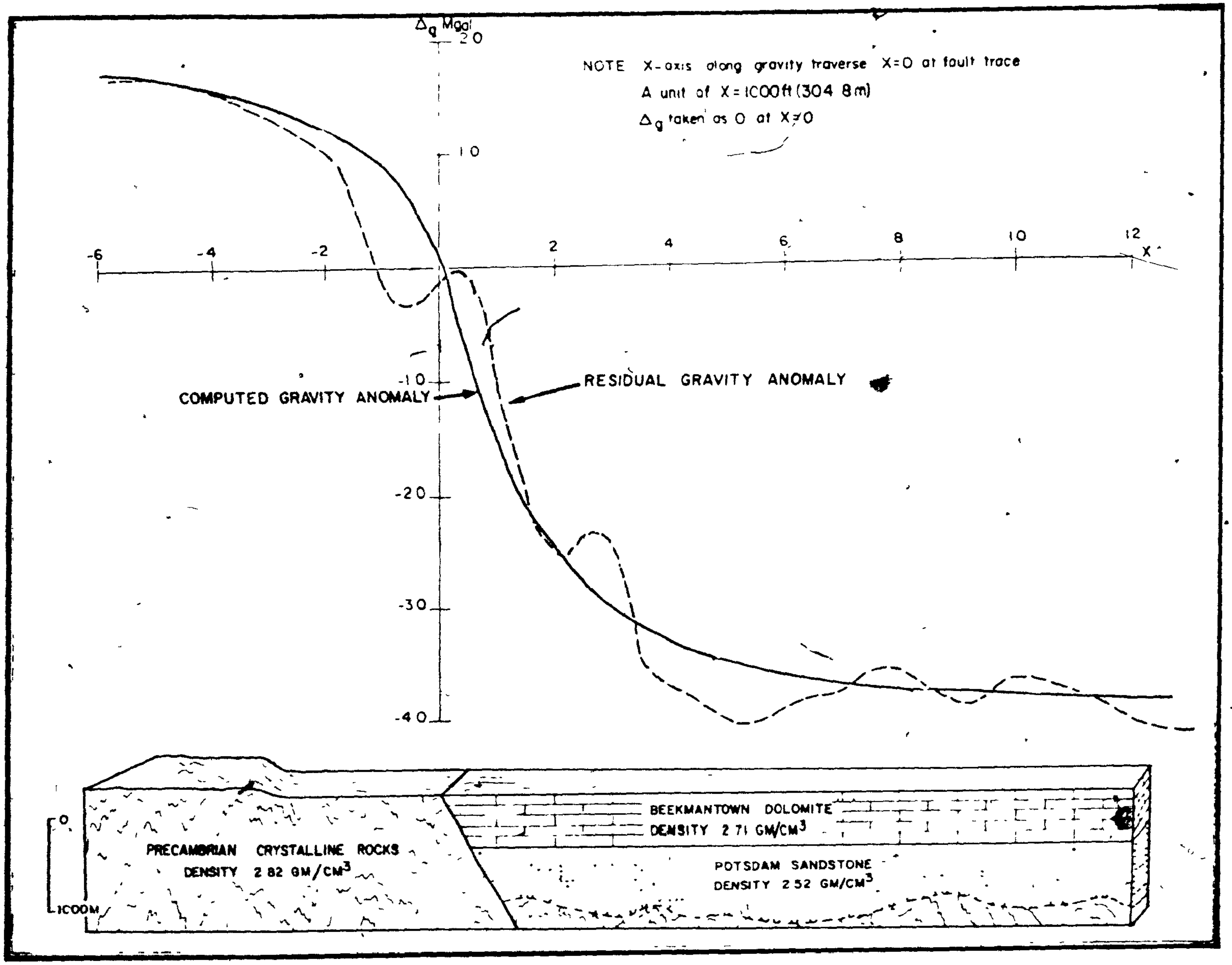

Fig.76. Interpretation of the gravity profile across the Shield margin in St.Jérome area.

For the application of the semi-infinite block formulae it is necessary to assume that the top and bottom of each faulted block are horizontal planar surfaces and that the faulted end itself is a single planar surface. It is very likely that the Beekmantown-Potsdam interface is practically planar. The sub-Potsdam surface on the other hand may be irregular and may have a relief of 60 to 90 m (see Ambrose 1964, p. 850). The upper surface of the Beekmantown on which the glacial deposits rest may also be somewhat irregular and these irregularities may give rise to significant variations of the overburden thickness. Also, the 'fault' itself, instead of being a single break, may consist of a major fault and several closely associated antithetic and synthetic faults. The fault surfaces themselves may not be planar. Some of them may be curved. The above possible departures in the actual and assumed conditions may be contributory to differences in the observed and computed anomalies. The 'humps' in the residual curve are quite sharp, indicating that they are caused by near-surface sources such as variation of overburden thickness.

The presence of a minimum of 1050 ft (320 m) of Potsdam beds below 900 ft (275 m) of Beekmantown rocks indicates that the shield surface descends below the Lower Paleozoic platform cover in a step, representing a drop of about 1950 ft (595 m) or more. This is evidently a feature related to faulting, not only because of its abruptness and large amplitude, but also because it transects the bedding and foliation directions of Precambrian rocks at a large angle (about 45°).

CONCLUSION

Near St. Jérome, Beekmantown rocks that abut against the margin of the Precambrian Shield seems to be underlain by a minimum of 1050 ft (320 m) of Potsdam sandstone. The absence of outcropping Potsdam beds along the shield margin can be explained by faulting.

TABLE XV

GRAVITY DATA LINE 12

Station No.	Elevation (ft)	Gravity Value (mgal)
560	211.805	-5.865
561	212.275	-5.818
562	212.570	-5.789
563	216.140	-5.495
564	218.160	-5.370
565	220.050	-5.281
566	223.420	-4.951
567	225.040	-4.785
568	225.140	-4.613
569	225.120	-4.761
570	225.220	-4.530
571	226.900	-4.288
572	231.190	-4.462
573	235.215	-4.524
574	239.085	-4.558
575	243.035	-4.293
576	245.295	-3.928
577	249.115	-2.538
578	249.335	-2.660
579	250.000	-1.554
580	255.730	0.000
581	261.000	-0.212
582	267.810	+0.223
583	276.640	+1.238
584	287.970	+1.664
585	306.790	+1.886
586	309.112	+2.183
587	312.560	+2.288

APPENDIX V

HYPOTHESES ON THE ORIGIN OF THE LAURENTIAN CHANNEL

Hypothesis of Spencer

Spencer (1890, 1903) was the first to propose a hypothesis to explain the origin of the Laurentian Channel. He believed it to have formed along the continuation of Logan's Line by river erosion followed by drowning. This view was later advocated by Upham (1894), Goldthwait (1924) and Johnson (1925). Objections to this hypothesis are:

1. Logan's Line does not continue along the outer part of the channel and hence cannot provide the necessary structural control for this part of the channel.
2. In the upriver continuation of the topographic low of the Laurentian Channel i.e. in the St. Lawrence Valley, the normal faulted structurally downbowed area clearly provides the necessary structural control for the topographic low (Chapter IV) and the association of Logan's Line seem incidental. It is likely that the morphotectonic relations are no different at least along the inner part of the Channel.
3. The channel does not show features typical of drowned river valleys (sinuous trends, seaward sloping valley floor, etc.) but has many features indicative of glacial erosion.

In view of the above objections Spencer's hypothesis is now abandoned, although it can hardly be denied that river erosion has played an important role in the development of the channel.

Hypothesis of Gregory

The Grand Banks earthquake of 1929 (Hodgson 1930) which occurred on the continental slope near the mouth of the channel, breaking the Trans-Atlantic cables, stimulated new thinking on the origin of the channel. Gregory (1929) interpreted the trough-like form of the channel as having formed by subsidence of a strip of land along parallel faults, i.e. a rift valley in the strictest sense. According to him, the earthquake shock and cable breaks were caused by fault movements related to further subsidence of the channel floor. Hodgson (1930), Keith (1930) and Doxee (1948) advocated a similar origin.

In interpreting the Laurentian Channel as a rift valley, Gregory (1929) took into account the possible downriver continuation of the St. Lawrence Valley structure, which Laflamme in 1908 had interpreted as a strip sunk between parallel fualts. This particular line of Gregory's argument is perhaps his greatest contribution to the understanding of the channel's origin. However, other lines of Gregory's argument are untenable in the light of later work. For instance, Heezen and Drake (1964) have shown that the Trans-Atlantic cables were cut by turbidity currents triggered by the earthquake, and not by fault-movements as Gregory suggested. Also, the view that the Laurentian Channel is a land form produced purely as a result of down-faulting is untenable in view of later studies Shepard (1931) and King and MacLean (1970a) who have made a strong case for glacial and fluvial erosion along the channel.

Hypothesis of Shepard

Shepard in 1931 found clear evidence of glacial erosion (hanging valleys, undulating channel floor with unconnected depressions etc.) along the channel. He suggested that prior to Pleistocene glaciation, the ancestral channel was probably formed by river erosion, possibly in part along fault lines. Later during the glacial period, tongues of ice moved down the ancestral channel, caused great deepening and widening to produce the present channel. Thus, despite his admission of some possible fault control, Shepard's hypothesis advocates a largely erosional origin for the channel.

Shepard's recognition that glacial erosion has played a major role in carving out the Laurentian Channel is a very important step towards understanding its origin. Shepard emphasized that his hypothesis explains the origin of the channel's present morphology. In this restricted scope, the hypothesis has been wholly substantiated by later work (King and MacLean 1970a). However, Shepard's hypothesis is perhaps inadequate to explain the channel's origin in a broader sense (see pp.179 to 189). Shepard's arguments against a large scale structural control for the channel are no longer relevant because they were directed specifically against Gregory's graben hypothesis which as discussed earlier, is outmoded. It should also be pointed out here that both

Shepard's and Gregory's hypotheses on the origin of the channel were formulated in the context of a broader problem: the origin of submarine canyons. As a result both hypotheses are based to a certain degree on generalities which tend to ignore some of the unique features of the channel.

Gregory's structural hypothesis and Shepard's erosional hypothesis represent a sharp polarization of views on the origin of the channel. This polarization has had a profound influence on later thinking of the channel's origin. This is unfortunate, because as will be evident later it now seems that its origin is related to an interplay of long continued tectonic activity and erosion.

<u>Hypothesis of McNeil</u>

McNeil (1956) suggested that the Channel was a fault controlled river valley and that the valley was subsequently drowned and modified by turbidity currents to produce the present channel. Later work has failed to reveal turbidities within the channel (Connolly <u>et al</u>. 1967) suggesting the absence of turbidity currents, as suggested by McNeil. The current that broke the Trans-Atlantic cables in 1929 may have originated on the continental slope.

APPENDIX VI

ORIGIN OF CONTINENTAL RIFTS

There have been two schools of thought concerning the origin of continental grabens. The first postulates that the floor of the graben has been lowered in relation to its sides between normal faults due to tensional forces (for example see Suess 1904; Gregory 1921). The second school considers that the floor of the graben is held down between reverse faults that formed under compressional forces (e.g. Wayland 1921; Willis 1936). The second hypothesis is now largely discredited in view of the almost ubiquitous evidence for tensional normal faulting associated with rift zones.

It is possible to explain the tensional environment and normal faulting along a rift zone as a result of crustal upwarping (Cloos 1939), and the subsidence of the graben block and also the rising of graben edges as a response to isostatic adjustment of crustal blocks (Taber 1927) but both the observed crustal stretching and the amount of block subsidence are too large to be explained this way (Vening Meinesz 1950 in Freund 1966; Zorin 1966b; Freund 1966; Illies 1970). But there is little doubt that the growth of crustal swells and formation of grabens are integral parts of a single phenomenon. Another aspect of this same phenomenon seems to be volcanism. It appears that growth of swells was accompanied by cracking of the crust and reopening of ancient joints and lineaments and that volcanism burst along some of these cracks. However, the interrelationships between volcanism, crustal updoming and graben formation are not clear. Some insight to this problem can, however, be gained from results of deep crustal seismological studies of graben zones, especially those conducted in the Rhine Rift zone (Mueller et al. 1969; Meissner et al. 1970, Ansorge et al. 1970). Models of crustal structure in the Rhine Rift zone, as interpreted from seismic data, have one common feature: a cushion-shaped intermediate layer with compressional wave velocities between 7.5 and 7.7 km/s is present between the crust and the mantle (Fig. 77); (Ansorge et al. 1970). Directly beneath the graben zone the "rift cushion" occurs between depths of 25 and 40 km approximately. Laterally it is believed to extend as far as the margin of the Rhenish upwarp (Illies 1970). Intermediate layers with similar P-wave velocities have also been reported from beneath the Baikal Rift and also from the Basin and Range Rift system (Meissner et al. 1970). The Red Sea too has an intermediate layer (Drake and Girdler 1964), but it is not yet known whether a similar feature exists along the rest of the East African Rift system. A cushion-shaped intermediate layer (P-wave velocity 7.3 km/s) is also present along the graben crested mid-Oceanic ridges (e.g. the mid-Atlantic Ridge), which are believed to be the oceanic counterparts of rifted continental crustal swells. It is generally believed that "rift cushions" beneath graben zones represent linear injections into the lower crust of hot, buoyant, mantle-derived

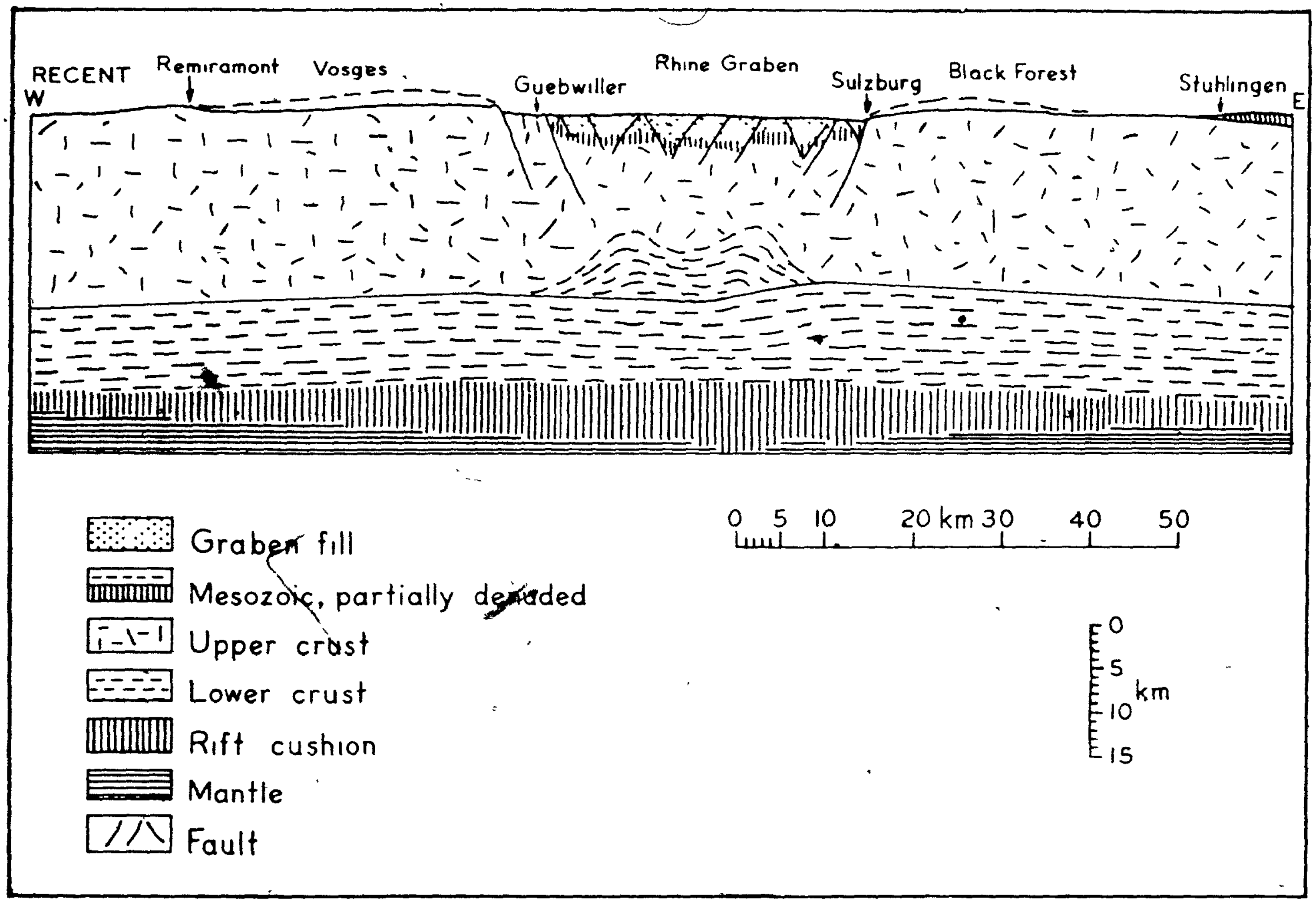

Fig.77. Model of the crust and upper mantle beneath the Rhenish upwarp as inferred from deep crustal seismology. After Illies(1970).

material (basalt) that flowed laterally into the shape of laccolithic masses under the influence of gravity (Ramberg 1967, Illies 1970). In the case of the Rhine Rift, Illies (1970) has remarked that there are many indications to suggest that the emplacement of the "rift cushion" may have led to the development of the crustal swell and the graben. These remarks should also apply equally well to rifted crustal swells in other areas. Moreover, as mentioned earlier, simple bending of the rigid crust by upwarping is insufficient to generate the observed crustal stretching along rift zones, whereas the lateral spreading of the injected masses provides an adequate mechanism. There is one basic question that still needs explanation. What factors determine the space distribution of these buoyancy injected masses? In the writer's opinion, the hypothesis that they represent upcurrents of mantle convection is appealing.

The concepts summarized above have been developed by research work carried out mainly in the classical rift zones and the mid-oceanic ridges. They should, however, with pertinent modifications, be applicable to the problem of origin of the St. Lawrence Rift system. At the present time it is difficult to rationalize further along these lines, because research studies on the St. Lawrence Rift system are not yet sufficiently advanced. For instance, it is not known whether the St. Lawrence Rift zone is underlain by a "rift cushion" or not. The available information, although admittedly scanty, indicates it probably does not have one (Rankin et al. 1969, Overton 1972). Such a situation, however, is not unexpected because of the very long period of relative inactivity of the rift system.